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REPORT OF ACTIVITIES PART A

Energy, Mines and Resources Canada
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REPRINTS

A limited number of reprints of the papers that appear in this volume are available by direct request to the individual authors.

INTRODUCTION

A geological survey provides all kinds of information on the geology of a country, interpreted in terms of those national activities that make use of, or are affected by, geology. Such information is essential for developing national policies for the rational use of energy and mineral resources and of the landmass.

The formal objectives of the Geological Survey of Canada are to provide a comprehensive inventory and understanding of the geological framework and processes in Canada as a basis for national policy and planning in all matters affected by geology with special emphasis on:

- *ascertaining Canada's energy and mineral resources;*
- *facilitating their exploration and development;*
- *promoting regional development in Canada;*
- *identifying and describing geological features and processes that affect environmental and ecological equilibrium, with special emphasis on the effects of the development of energy and mineral resources;*
- *identifying and studying features of the recent geological past and ongoing processes that affect the use of the landmass, engineering design, urban development, and renewable resource industries, such as forestry, agriculture and fisheries;*
- *identifying and assessing natural hazards, and*
- *making available information on the landmass of Canada and its surrounding continental shelves for use by other government agencies, industry and the general public.*

Most of the information obtained by the Geological Survey is designed for use by both government and the general public, including industry and, therefore, must be widely disseminated. This is accomplished by means of an extensive publication program which in 1974-75 involved the publication of 75 reports, as well as numerous geological and aeromagnetic maps and interim Open File releases. The "Report of Activities" series has developed rapidly in the past few years and in 1975 nearly 300 short reports comprising 1300 pages of text and covering nearly all the fields of activity in which the Geological Survey is carrying out research, were published in this series.

In this report, the papers are grouped by subject in the Contents Section but in the main part of the publication the order followed is that in which the manuscripts were received. Although it might be preferable to group the papers together, such a procedure considerably complicates the preparation of the report and increases the production time. As many of the reports included in the "Report of Activities" represent the first release of new information, which in some cases may be of considerable value to industry in planning future programs, it is felt that the loss resulting from the random sequence is more than offset by the earlier release of data. Material for this report was received and edited between November 3rd and 14th; production editing, typing, proofreading and preparation of camera-ready copy were carried out between November 3rd and December 8th. Printing was scheduled to permit a mid-January release date.

All of the reports that make up this publication result from projects that form part of one or several of the Survey's programs. It may be of interest to users of this report to learn of the general objectives of some of the activities reported herein.

Studies of the composition, stratigraphy and structure of the rocks of the Appalachian geosyncline and eastern Canadian platform regions are important in evaluating the mineral and hydrocarbon resources of the area. After some years during which activity by the Geological Survey in this region was limited due to the need to undertake programs in the frontier areas, renewed emphasis is being given to studies in bedrock and economic geology and reports 35-38 present interim results of some of these.

The Cordilleran Region has, over the years, been a major source of Canadian mineral wealth and will undoubtedly continue to be so. Determining in advance what the resources are depends to a large degree on the existence of a comprehensive data base which will lead to a better understanding of the geology of the Cordilleran Orogen. The Geological Survey's current activities in this region are designed to complete the reconnaissance phase of regional investigations to provide a broad geological and tectonic framework for the region and to carry out detailed studies of specific problems to further the understanding of the nature and sequence of geological processes, particularly the formation and localization of mineral deposits. Reports 10 to 16 present the results of the "Takla Project", a study designed to focus on the plutonic, volcanic, sedimentary and structural history of the late Triassic and early Jurassic rocks of north-central British Columbia, and in part undertaken in order to better understand the nature and origin of the numerous copper showings common in rocks of these ages throughout the Cordilleran. Reports 17-19 present the results of the 1975 field component of "Operation Finlay", a project designed to complete the geological reconnaissance of parts of north-central British Columbia. These reports add data concerning the presence of copper showings and the authors point out that, on the basis of geological characteristics, other base metals should also be expected.

During the latter part of the field season, a party working in the Proterozoic rocks of the Mackenzie Mountains encountered a significant copper showing. This discovery was made known to the public through an Open File release and the news led to considerable activity. Report 24 presents the general geological setting of the Redstone River area, information that was not available for publication at the time the Open File was released. This information, coupled with the knowledge that significant mineralization does occur, should prove useful in predicting favourable areas for further prospecting for base metals.

The Geological Survey carries out many studies in the field of Economic Geology. Some of these are designed to identify and interpret the geological characteristics of mineral deposits and to investigate their relationships to their geological environments. The results of such studies permit us to make more accurate evaluations of the distribution, character and amount of our mineral resources as well as providing guides to the discovery of a large variety of mineral deposits.

Reports 64 and 67 present general observations in the nature of Canadian lead and zinc deposits and are part of continuing studies in commodity metallogeny, the comprehensive study of all aspects of the geology of specific mineral commodities to determine the ways in which they are concentrated in the crust and how these affect quality, distribution and identification of such concentrations. Report 70, also concerned with commodity studies, is a progress report on a general study of the barite, fluorite and celestite occurrences of Canada.

Nuclear energy offers a possible solution to the world's long-term energy needs and as part of its role of providing the fundamental data needed for the exploration and assessment of mineral deposits, the Geological Survey is engaged in a Uranium Program that involves field and laboratory studies. Reports 71 and 73 are general assessments of the mode of occurrence of uranium whereas reports 72 and 74 present more specific data. Closely related to the objectives of the Uranium Program is the Federal-Provincial Uranium Reconnaissance Program, designed to provide industry with high quality reconnaissance exploration data. Results of the 1975 field component of this program were released in early November in Paper 75-1C.

More than one quarter of the reports that make up this volume present the results of projects that are concerned with Quaternary and Holocene geology. Such projects are designed to meet a wide range of needs: to provide systematic coverage of the surficial geology of the Canadian landmass necessary for effective use of the terrain; to identify and assess the occurrence and magnitude of natural terrain hazards; and to provide the geoscience data needed to assist in the maintenance and restoration of the physical environment. Environmental and Engineering Geology studies are fundamental in evaluating terrain hazards and in the past few years have been concerned mainly with the proposed Mackenzie Valley highway and pipeline. The possibility that exploitable deposits of natural gas exist in the Arctic Islands has accentuated the need for geoscience information on potential pipeline routes to southern Canada. Reports 58 and 59 present the results of terrain performance studies along the Somerset-Boothia corridor. Geomorphological studies such as those presented in reports 40, 56 and 62 also contribute significantly to the evaluation of transportation routes. The eight reports grouped together as Inventory Mapping and Stratigraphic Studies are, like the comparable bedrock studies, essential to building a data base from which many other studies draw.

The Canadian Shield underlies about half of Canada's landmass. The Geological Survey's activity in this vast region is currently concentrated in the Northwest Territories and is designed to provide a geological data base consistent with modern standards and with that in the better-known, more southern parts of the Shield. The more detailed studies now being undertaken became possible with the completion of the field phase of the reconnaissance mapping of the Shield in 1973 when staff became available. Reports 75 to 82 are concerned primarily with the systematic mapping of northern Precambrian areas. Although no dramatic discoveries of economic importance are reported, the new information enhances the potential for discovery of mineral deposits, especially within the Foxe Fold belt where ultramafic flows and acid volcanic centres have been identified.

Stratigraphic studies are a key requirement in meeting one of the Survey's objectives--a comprehensive inventory and understanding of the geology of the sedimentary basins of western and Arctic Canada. Such studies facilitate exploration and development by providing geological information related to the occurrence of hydrocarbons, coal and other minerals. Reports 91 to 96 present progress reports of a study initiated in 1973 which is designed to give systematic coverage at 1: 250 000 scale of northern Ellesmere Island by means of establishing a coherent stratigraphic framework. Reports 104 to 107 present the results of similar studies on Somerset Island where the data on bedrock geology are essential in evaluating the feasibility of constructing a proposed natural gas pipeline. As part of this study, the Somerset Island kimberlite occurrences were examined and brief descriptions for all known exposures are given in report 107.

As the "Report of Activities" is probably one of the most widely circulated and generally available publications of the Geological Survey, a list of its senior management and their areas of responsibility has been included with this issue in order to acquaint the general public with the names of those responsible for the overall direction of the Survey's research program.

R.G. Blackadar
Chief Scientific Editor.

1. PALEOZOIC GEOLOGY OF NORTHEAST GULF OF ST. LAWRENCE

Projects: 750061, 750001

R.T. Haworth[1] and B.V. Sanford[2]

Introduction

A preliminary attempt to define the distribution of Paleozoic formations in the northern Gulf of St. Lawrence was made by Shearer (1973). His seismic reflection (airgun) profiles showed little penetration due to the seismic velocity characteristics of the formations. Even the more sophisticated seismic reflection systems used by petroleum exploration companies give only slightly better penetration but lose considerable definition of the Quaternary units. Shearer (1973) confirmed that the gentle regional dip of the Paleozoic formations exposed on Anticosti Island (Bolton, 1972) extends almost entirely across the Gulf of St. Lawrence between Quebec and Newfoundland. A pronounced dip reversal about 8 km offshore from western Newfoundland, between Port au Port Peninsula and Bonne Bay, was interpreted as indicating where the autochthonous rocks onlapped the offshore extension of the allochthon (Cumming, 1967). In the extreme northern part of the Gulf the observed onshore contacts between formations were extrapolated offshore primarily on the basis of morphology, since little or no seismic penetration was obtained. Apart from the samples collected offshore from central western Newfoundland (Lilly, 1966), no bedrock samples were available as control for the interpretation of geophysical surveys. On the basis of widespread sampling of surficial sediments (Loring and Nota, 1973; Canadian Hydrographic Service, 1972) it was known that outcrop occurred on the ridge separating Mecatina and Esquiman troughs. It was proposed, therefore, that further seismic reflection work would be carried out to delineate these and other areas of outcrop for sampling with a rock-core drill.

The Bedford Institute electric rock-core drill (Fowler and Kingston, in press) is capable of taking a solid rock core that is 5 m long and 25 mm in diameter. In order for a bedrock core to be obtained with this drill, the surficial cover must therefore by very thin. Low energy seismic sources conventionally used for surficial geological studies, e.g. $16.10^{-6}m^3$ (1 in^3) airguns, do not provide adequate resolution of the surficial layer to determine whether recovery of a bedrock core is probable. As a result, the pre-1975 efficiency of drilling has been relatively low (Harris, 1973). The development of a high resolution deep towed seismic system by Huntec '70 Ltd. (McKeown, 1975) provided the capability to resolve the thickness of the overburden to within approximately 1 m and hence provided potential for increasing the efficiency of the rock-core drill. In order that the drilling could be carried out within a known seismo-stratigraphic framework, a small airgun (156 or $625.10^{-6}m^3$, 10 or 40 in^3) was used to provide as much penetration as possible without totally relinquishing resolution of the surficial material.

Bathymetry, gravity and magnetic field data were collected on all survey lines to assist in the location of drill sites and interpretation of the regional geology, although for these parameters the whole eastern half of the Gulf of St. Lawrence and already been surveyed in much greater detail (Haworth and MacIntyre, 1975).

Data Collection

The philosophy behind survey procedures at the beginning of the cruise was quickly modified through experience. After completion of two long survey lines, part of an overall pattern planned to supplement the earlier Shearer (1973) data, it was apparent that where seismic penetration occurred, even though the new records were of slightly better quality than Shearer's, the continued regularity in the dip of the presumed Paleozoic sedimentary horizons meant that additional data contributed relatively little to an increase in knowledge. Because of time lost through unforeseen incidents early in the cruise, it was therefore decided to proceed by basing the approximate location of drill sites primarily on Shearer's data and to concentrate our efforts on precise location of the sites through use of the high-resolution system. Thus, the majority of the regional seismic reflection coverage (Fig. 1.1) is still provided by Shearer's data.

Results

In the southeastern part of the survey area there were few areas where the overburden was sufficiently thin to permit drilling, and most of these areas were less than 1 km in extent. Unfortunately, it is not possible to deploy the Huntec gear when the ship is stopped on station, so that the best drilling site determined on the basis of several Huntec traverses had to be returned to either by the use of LORAN-C, dead reckoning, or by reference to variations of bathymetry observed during the site survey. When LORAN-C was working this was a relatively easy task, but unfortunately there were continuous problems with that equipment. The addition of a "mini" high-resolution seismic reflection system on the drill frame to indicate depth of overburden immediately beneath the drill would go a long way towards removing the final uncertainty about drill positioning. Given these constraints our general objective was to concentrate on getting samples, preferably evenly distributed through the stratigraphic sequences south of the Esquiman Trough. Within the time available drilling was attempted in three areas at sites 1 through 4 (Fig. 1.3) and core was recovered in each case.

[1]Atlantic Geoscience Centre, Dartmouth

[2]Regional Economic and Geology Division

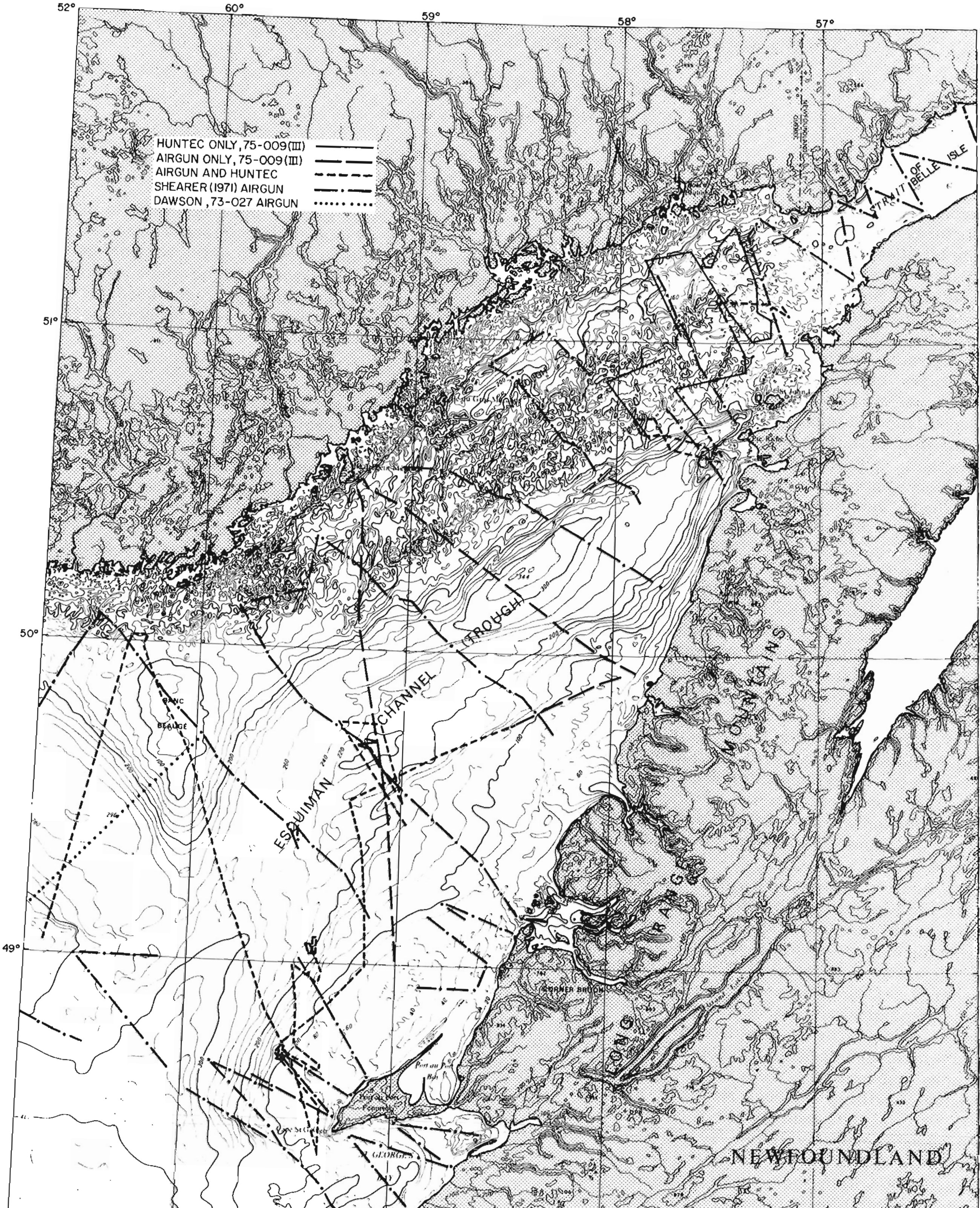

Figure 1.1. Index map of seismic reflection profiles, northeast of Gulf of St. Lawrence, and the morphology of the area profiled. Scale 1: 2 000 000.

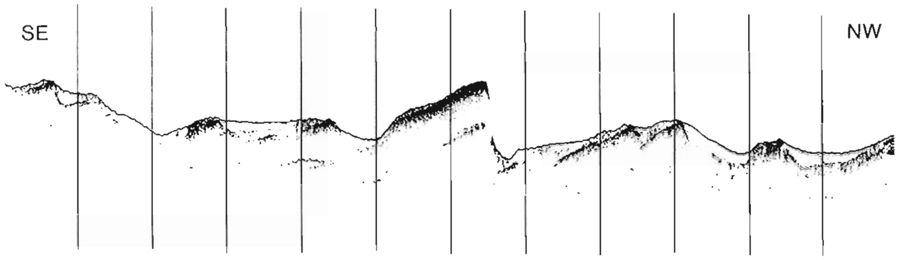

Figure 1. 2.

Huntec high resolution seismic reflection profile of cuestas forming the ridge between Mecatina Trough and Esquiman Trough. Vertical bars are separated by approximately 400 m and have a length corresponding to approximately 40 m.

In the northern part of the survey area, north of the Esquiman Trough, very little seismic penetration was obtained so that morphology of the sea floor provided the best indication of lithological boundaries. A series of Huntec profiles across morphological units (Fig. 1. 1) provided a good indication of potential drilling sites and revealed more clearly than ever before the cuestas that are characteristic of this area. Whereas bathymetric profiles only provide a general indication of this character except where a cuesta has a high and steep scarp face, the Huntec profile is able to see through the surficial material and reveal the clarity in shape of the small cuestas (Fig. 1. 2). Most of the drilling in this area concentrated on sampling the bare crests of those cuestas sufficiently large to provide a good target. The prominent cuesta in Figure 1. 2 would only marginally qualify as a potential drilling target because of the thickness of surficial material, whereas all of the others have too limited an outcrop. In most cases drilling would not be attempted unless the outcrop extended between two fiducial marks on the Huntec record (the equivalent of approximately 400 m), because of the distance it takes to manoeuvre and stop *CSS Hudson* (200-300 m).

The geological map (Fig. 1. 3) prepared on the basis of the seismic data and the rock core samples, is obviously critically dependent upon the correct identification of those samples. Earlier doubts as to whether any sample was an erratic were virtually eliminated through examination of the readout from numerous sensors on the drilling rig. Probability that the cores represent bedrock is given in Table 1. 1. Perhaps the most critical samples are those obtained at sites 7 and 8. Site 7 was situated on a mesa at the northern end of the Mecatina Trough. The first attempt recovered over 1 m of granite of presumed Grenville age, after drilling through 3 m of unconsolidated overburden. However, the drilling log and the presence of a quartz sandstone at the maximum drill extent confirmed that the granite was an erratic. The second attempt yielded a 2. 5 m bedrock core in which there is a very sharp change in lithology. The upper part of the core consisting of algal limestone has been identified tentatively as Forteau, with the lower part possibly being Bradore. This does not seem unreasonable due to the close proximity of rocks of the Forteau Formation overlying the Bradore Formation on the north shore headlands at the gulf end of the Strait of Belle Isle.

The samples from sites 8 and 9 have been tentatively identified as being from the Hawke Bay Formation and St. George Group respectively. The two sites are separated by a pronounced morphological lineament (Fig. 1. 1) that is interpreted as a fault with the eastern

TABLE 1. 1

Lithology of core samples, Gulf of St. Lawrence

CORE	LENGTH (m)	RELIABILITY*	LITHOLOGY
1	0. 4	Good	Coarse grained brown-red sandstone, massive to vague bedded.
2	0. 8	Good	Medium to coarse grained grey sandstone.
3	0. 1	Minimal	Grey shale or shaly mudstone.
4	0. 2	Poor	Grey shale or shaly mudstone and grey fine sandy limestone.
5	1. 8	Excellent	Light brown to grey brown fine grained limestone.
6	1. 1	Very good	Brown medium crystalline dolomite.
7	2. 5	Excellent	Brown coarsely crystalline limestone and grey siltstone overlying and in sharp contact with white massive sandstone.
8	2. 9	Excellent	Light to dark grey finely crystalline dolomite.
9	2. 6	Excellent	Brown, finely crystalline dolomite.
10	2. 6	Excellent	Dark brown, fine grained limestone breccia.

*Probabality that the core represents bedrock at particular locality.

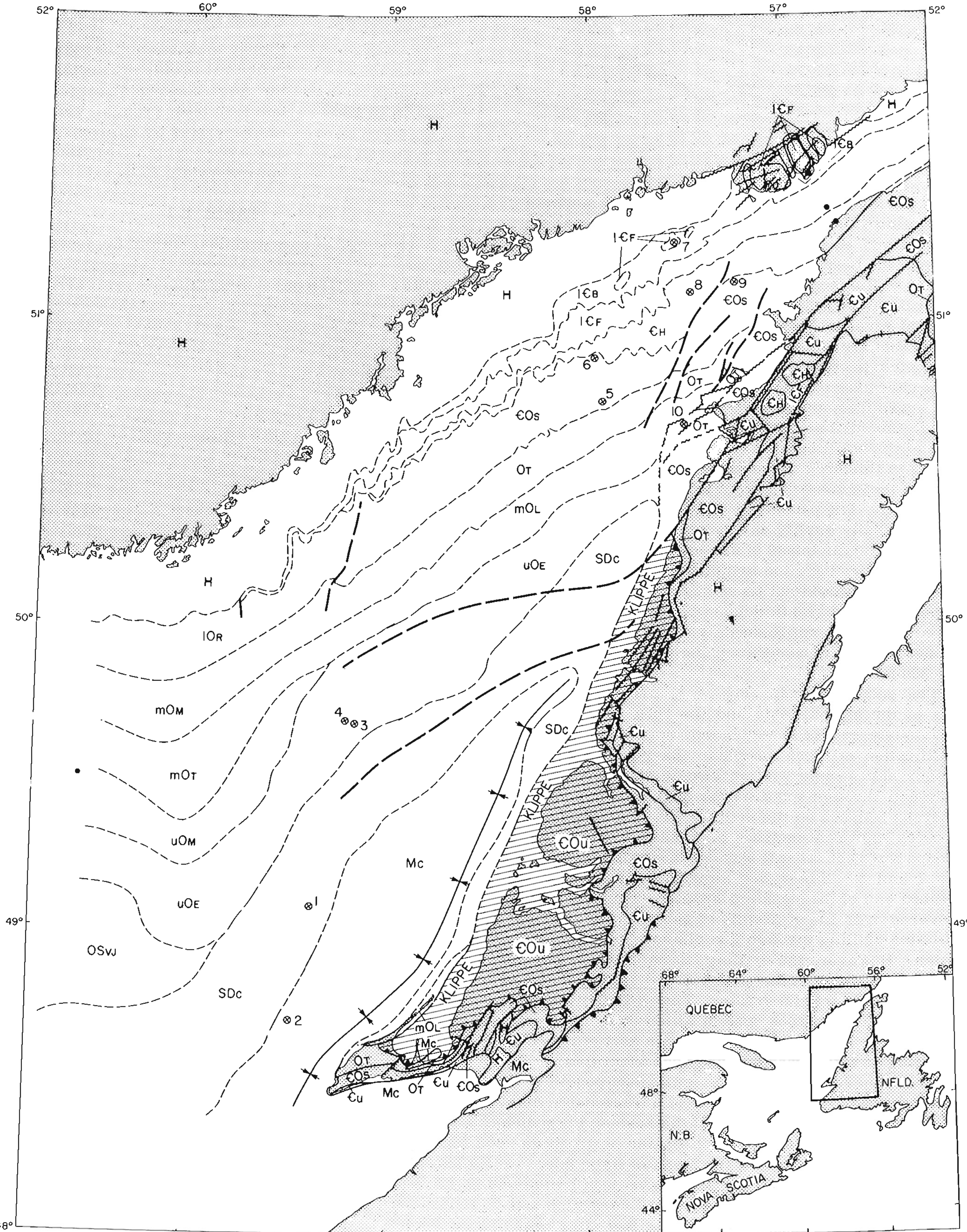

Figure 1.3. Geological map, northeastern Gulf of St. Lawrence. Key to map is on opposite page. Scale 1: 2 000 000.

L E G E N D

MISSISSIPPIAN

Mc	CODROY GROUP: grey and red sandstone, siltstone, shale and conglomerate (may include undifferentiated ANGUILLE FORMATION).

SILURIAN AND DEVONIAN

UPPER SILURIAN AND LOWER DEVONIAN

SDc	CLAM BANK FORMATION: red sandstone, siltstone and shale, succeeded by grey, finely crystalline fossiliferous limestone.

ORDOVICIAN AND SILURIAN

UPPER ORDOVICIAN, LOWER AND MIDDLE SILURIAN

OSvj	VAUREAL, ELLIS BAY, BECSCIE, GUN RIVER AND JUPITER RIVER FORMATIONS undivided: limestones with interbedded shales.

ORDOVICIAN

UPPER ORDOVICIAN

uOe	ENGLISH HEAD FORMATION: grey, micaceous, fissile shale with interbeds of grey argillaceous to silty fossiliferous limestone.
uOm	MACASTY FORMATION: black shale with interbeds of limestone.

MIDDLE ORDOVICIAN

mOt	TRENTON AND BLACK RIVER GROUPS: limestone and shale.
mOl	LONG POINT FORMATION: limestone, succeeded by siltstone and shale.

LOWER AND MIDDLE ORDOVICIAN

mOm	MINGAN FORMATION: limestone with minor sandstone and shale.
Ot	TABLE HEAD FORMATION: brown, argillaceous limestone and brecciated limestone.

CAMBRIAN (?) AND LOWER ORDOVICIAN

lOr	ROMAINE FORMATION: brown and tan, fine to medium crystalline dolomite.
ЄOs	ST. GEORGE GROUP: brown and tan, fine to coarsely crystalline dolomite, and brown micritic limestone (includes Ot southeast of klippe).

CAMBRIAN AND ORDOVICIAN

ЄOu	UNDIVIDED: slate, greywacke, quartzite, limestone conglomerate and ophiolitic rocks.

CAMBRIAN

LOWER AND MIDDLE CAMBRIAN

ЄH	HAWKE BAY FORMATION: grey fine to coarse grained ortho-quartzitic sandstone, succeeded by grey, fine crystalline dolomite and limestone, with interbeds of grey shale.
lЄF	FORTEAU FORMATION: brown, finely crystalline algal limestone with interbeds of grey shale.
lЄB	BRADORE FORMATION: red, medium to coarse grained sandstone and conglomerate succeeded by grey, fine grained sandstone.
Єu	CAMBRIAN UNDIVIDED: sandstone, dolomite, shale and slate.

HELIKIAN OR OLDER

H	Grenville terrain consisting of gneissic and granitic rocks.

	Synclinal axis	
	Lineament	(assumed fault)
	Fault	(defined, approximate)
	Geological boundary	(defined, seismic, assumed)
⊕	Core Hole	(GSC)
●	Core Hole	(Other agencies)

Geology by B.V. Sanford and R.T. Haworth side downthrown. That fault is subparallel to a series of faults between Pointe Riche and the Precambrian northern highland of the Long Range Mountains (Fig. 1. 3), although the throw of the onshore faults is in the opposite sense (Cumming, 1967). It may be possible from the surface dips observed on the Huntec records to put limits on the extra thickness of section present offshore from St. John Island (north of Pointe Riche) in the centre of the minor graben created by the faults.

Summary

The major features of the geological map remain as they were outlined by Shearer (1973), but the details have been modified on the basis of onboard identification of drill hole samples. In addition to more definitive identification and dating of the samples obtained, considerable work remains to be done. From the seismic reflection data better estimates of the thickness of each of the formations and closer identification of their seismic character can be made. These tasks will be better accomplished with additional short seismic refraction profiles, but first the data already available (Hobson and Overton, 1973) need review and integration with the existing reflection data. Although the extent of the surficial geological work (Loring and Nota, 1973) would appear to make further effort in that direciton superfluous, the area is ideal for an evaluation of the success of the Huntec system. If the acoustic response of the system as calibrated on the Scotian Shelf earlier in the cruise can be used to predict the known sedimentary types mapped in the Gulf of St. Lawrence (Loring and Nota, 1973), the success of the system will have been conclusively demonstrated.

References

Bolton, T. E.
1972: Geological map and notes on the Ordovician and Silurian litho - and biostratigraphy, Anticosti Island, Quebec; Geol. Surv. Can., Paper 71-19, 45 p.

Canadian Hydrographic Service
1972: Distribution of surface sediments, Gulf of St. Lawrence (D. H. Loring and D. J. G. Nota); C. H. S. Map 811-G, Dep. of the Environment, Ottawa.

Cumming, L. M.
1967: Platform and klippe tectonics of western Newfoundland: A review; *in* Appalachian Tectonics, ed. T. H. Clark, Roy. Soc. Can., Spec. Pub. 10, p. 10-17.

Fowler, G. A. and Kingston, P. F.
An underwater drill for continental shelf exploration; J. Soc. Underwater Technology (U. K.). (in press).

Harris, I. M.
1973: Underwater, diamond drill trial and support survey northeast of Belle Isle, Newfoundland; *in* Report of Activities, Part A, Paper 73-1A, p. 109-110.

Haworth, R. T. and MacIntyre, J. B.
1975: The gravity and magnetic fields of Atlantic offshore Canada; Marine Science Paper 161, Geol. Surv. Can., Paper 75-9, 22 p.

Hobson, G. D. and Overton, A.
1973: Sedimentary refraction seismic surveys, Gulf of St. Lawrence; *in* Earth Science Symposium on Offshore Eastern Canada, ed. P. J. Hood, Geol. Surv. Can., Paper 71-23, p. 325-336.

Lilly, H. D.
1966: Late Precambrian and Appalachian tectonics in the light of submarine exploration on the Great Bank of Newfoundland and in the Gulf of St. Lawrence — Preliminary views; Am. J. Sci., v. 264, p. 569-574.

Loring, D. H. and Nota, D. J. G.
1973: Morphology and sediments of the Gulf of St. Lawrence; Fish. Res. Board Can., Bull. 182, 147 p.

McKeown, D. L.
1975: Evaluation of the Huntec ('70) Hydrosonde deep tow seismic system; Bedford Inst. Rept. BI-R-75-4, 115 p.

Shearer, J. M.
1973: Bedrock and surficial geology of the northern Gulf of St. Lawrence as interpreted from continuous seismic reflection profiles; *in* Earth Science Symposium on Offshore Eastern Canada, ed. P. J. Hood., Geol. Surv. Can., Paper 71-23, p. 285-303.

MARINE GEOSCIENCE SURVEY NORTHEAST OF NEWFOUNDLAND

Projects 750001 and 730081

R.T. Haworth[1], W.H. Poole[2], A.C. Grant[1], and B.V. Sanford[2]

Introduction

The hydrographic-geophysical surveys carried out off the east coast of Canada as a co-operative project of the Canadian Hydrographic Service, Department of the Environment and the Atlantic Geoscience Centre, together with additional surveys carried out as part of other geoscientific programs, have provided geophysical data adjacent to the entire coast of Newfoundland except for the triangular area inshore of a line between Cape Bauld and Fogo Island (Fig. 2.1). The area was avoided by the 1973 hydrographic surveys because a sufficiently accurate navigation system was not available to provide continuous fixes throughout the whole area. Recognizing the geoscientific importance of the area as the northern extremity of the Appalachian system and having, since 1973, gained considerable experience with the LORAN-C navigation system, it was decided to use the best available combination of navigation systems in order to complete the basic survey of areas adjacent to Newfoundland. Such a survey was carried out during the first three weeks of August 1975.

The bathymetric charts of the area are unreliable, having been produced on the basis of information collected on isolated tracks by a multitude of vessels primarily in the 19th century. Grant (1972) ventured only part way into this area to collect geophysical data.

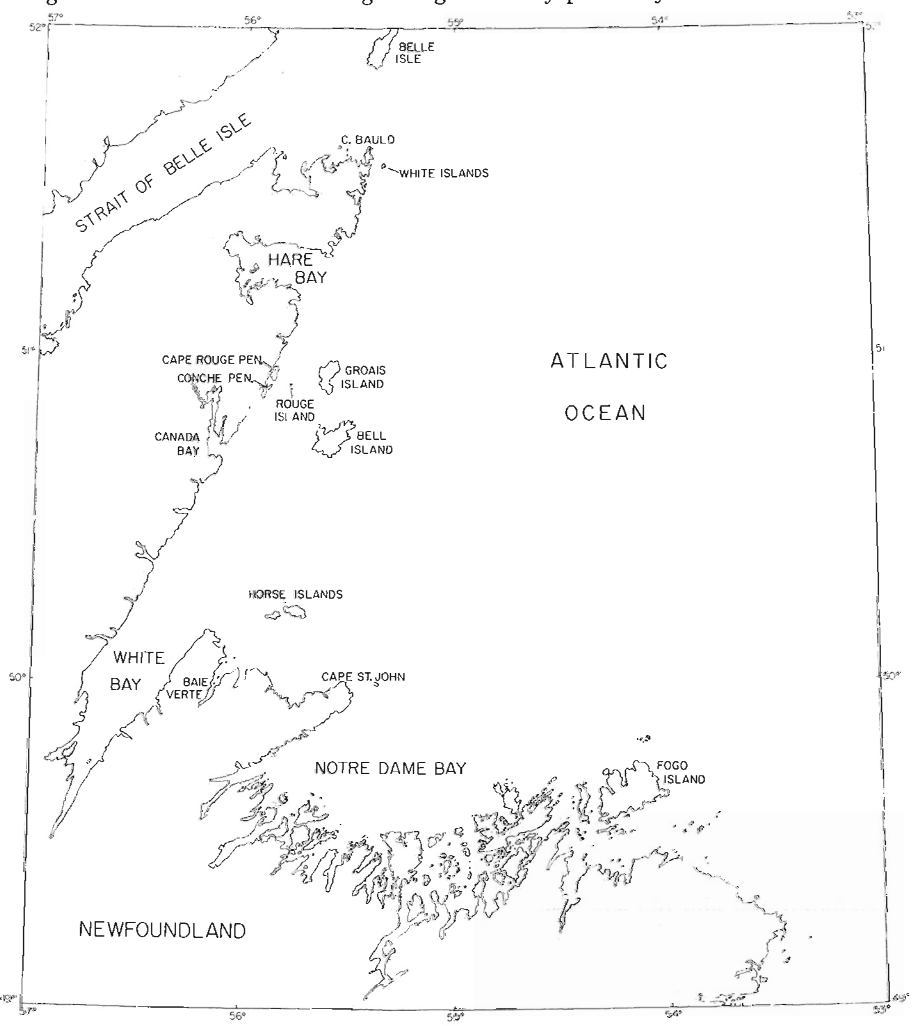

Figure 2.1. Index map showing locations referred to in text.

[1]Atlantic Geoscience Centre, Dartmouth.

[2]Regional and Economic Geology.

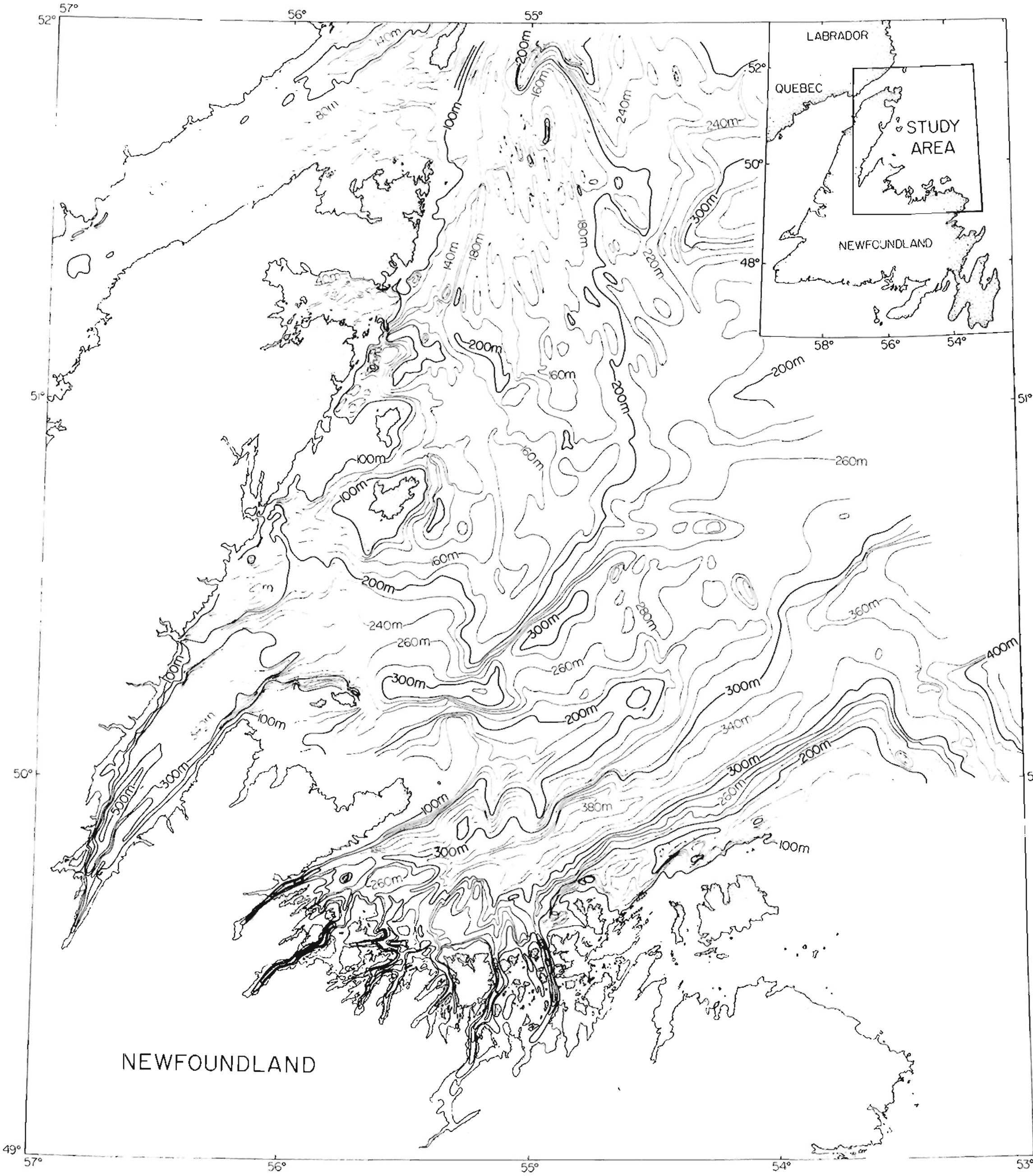

Figure 2. 2. Bathymetry map, northeast of Newfoundland. Scale 1: 2 000 000.

With such a dearth of basic information, a regular pattern survey was deemed necessary to provide the basis for future work in specific areas of importance. Survey lines were planned every 5 minutes of latitude between the coast of Newfoundland and the western extremity of the MINNA 73-019 survey (Macnab, 1974; Grant and Macnab, 1974) in addition to coverage of White Bay and Notre Dame Bay. Bathymetry, gravity and magnetic field data were to be collected on all survey lines, with the addition of seismic reflection data on alternate lines. At the outset it was also hoped to drill one or two rock cores. The good weather experienced throughout the entire operation resulted in the collection of more survey data than anticipated, and the drilling of 13 rock cores. This has enabled us to prepare preliminary maps of all parameters and combine their interpretation in a preliminary geological map of the area.

Navigation

Major problems were experienced with the LORAN-C navigation system. Correction for the overland phase delay of signals received from Cape Race (southeast Newfoundland) had to be continuously computed on the basis of comparison with satellite navigation and radar

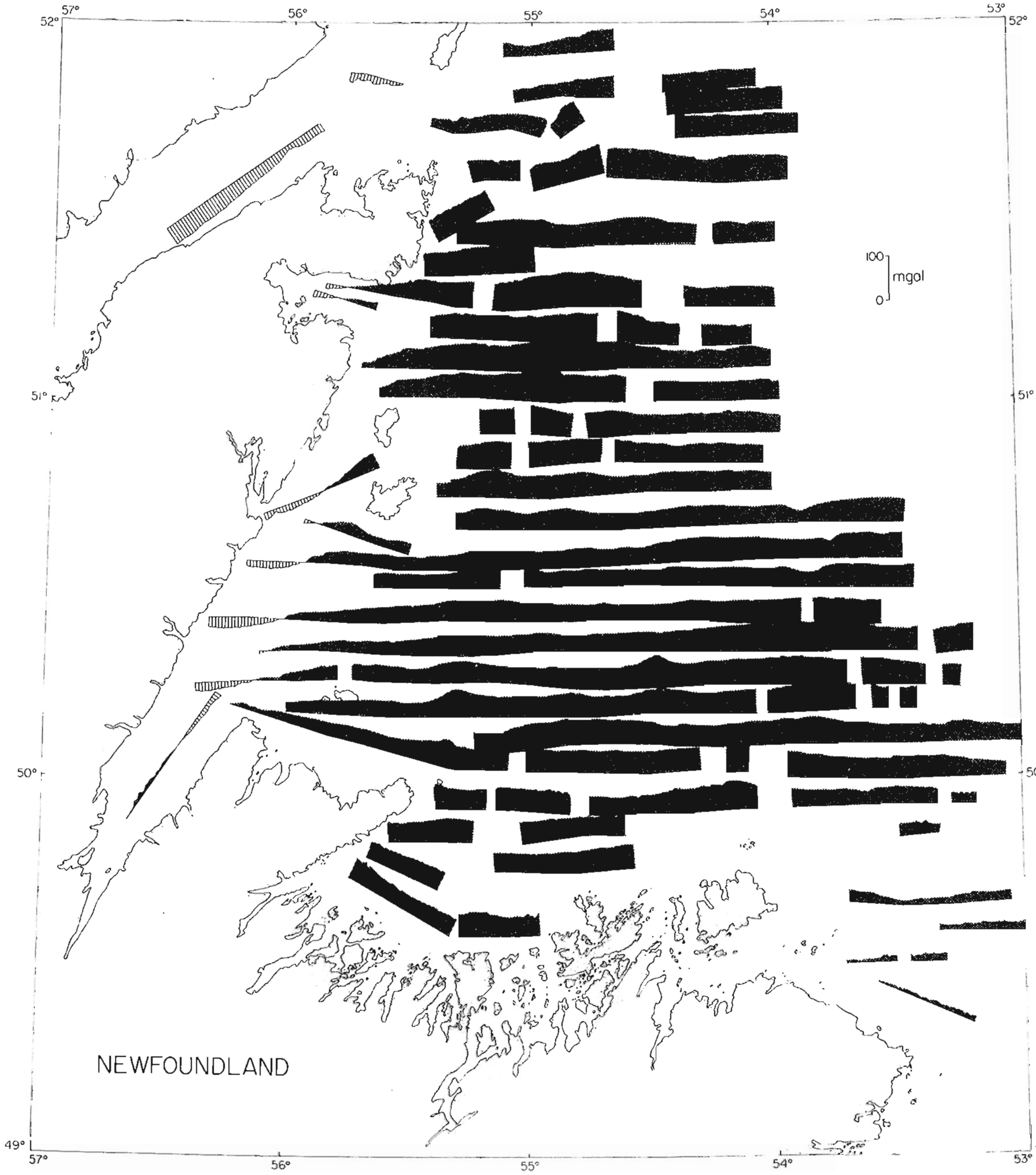

Figure 2. 3. Bouguer gravity anomaly profiles, northeast of Newfoundland. Anomalies are plotted along selected ships tracks, darker shading indicates positive anomalies. Scale 1: 2 000 000.

fixes. Poor reception of the LORAN-C signal on occasions, and continuous problems with the fix calculation program used in conjunction with the Austron receiver made accurate real time navigation impossible. As a result, complete post-cruise processing of the potential field data has been necessary. The effect of this on the anomalies as presented in Figures 2. 2 and 2. 3 will be almost negligible although the spatial relationship between anomalies and seismic events on adjacent lines run at widely different times might be changed slightly.

Data Collection

All data used in the preparation of the bathymetric map (Fig. 2. 4) were collected with a conventional 12 kHz wide-beam echo sounder. Bathymetric values were manually recorded in digital form every 5 minutes of time together with intervening peaks and troughs. Computer profiles of the data and plots of posted values were used in conjunction with pre-existing bathymetric data inshore to produce a contour map of the study area at a scale of 1: 250 000 from which Figure 2. 4 was prepared.

The gravity field was measured with a Graf-Askania s-2 surface ship gravimeter and recorded automatically in digital form every minute. The calm weather ensured good data accuracy with a maximum discrepancy of 4 mgal at track intersections. The gravity values were calculated with reference to base values on the Canadian gravity mapping datum and were reduced to Bouguer anomalies using the International Gravity Formula of 1930 and an infinite slab correction for water depth assuming a crustal density of 2.67 g cm^{-3}. This produces gravity data compatible both with that east of the study area in the Natural Resource Map series (Canadian Hydrographic Service, 1975) and also with that published for Newfoundland (Weaver, 1968; Miller and Deutsch, 1973).

Magnetic field data were collected simultaneously with both a Barringer proton precession magnetometer and a Geometrics magnetic gradiometer. The Geometrics gradiometer was on trial in an attempt to eliminate diurnal variations from magnetic measurements. Although a buoy magnetometer was moored in the centre of the study area to monitor magnetic variations this broke from its mooring and was lost. The variations monitored at St. John's have not been applied to the profiles in Figure 2.3. The Barringer magnetometer, with which no problems were encountered, was used as the primary source of information presented here, although survey lines in the outer portion of Notre Dame Bay run only with the gradiometer have been included. The data from both instruments were logged automatically every six seconds although one minute averages were used in the preparation of Figure 2.3.

Seismic reflection information was collected using a 156 $10^{-6}m^3$ (10 in^3) airgun operating at approximately 10 MPa (1500 psi) with a 2 second repetition rate. The seismic signals were received by a short hydrophone having a 2 m long live section and were recorded without filtering on a Hewlett Packard FM tape recorder. Filtered analog records were made simultaneously. On most of the seismic reflection lines data were also collected with the Huntec high resolution deep tow seismic system (McKeown, 1975). These data were used primarily in the selection of drill sites although there is much potential for sedimentological studies from them.

Geological Interpretation

The interpretation (Fig. 2.5) was prepared during the survey, at which time all the geophysical information was not in a useable form, nor was the navigation information fully corrected for its errors, nor had the drill samples been analyzed except visually as hand specimens. It must therefore be viewed as an initial report for possible revision in the light of further analysis prior to more formal publication in the near future.

The survey proceeded from north to south so that the on-board interpretation developed from concepts originating in the north. Immediately east of Cape Bauld and Belle Isle, the sea floor is very irregular with seismically hard units well exposed. These units are in abrupt, fault(?) contact units in which, particularly

Table 2.1

Core	Length (m)	Reliability*	Lithology
11	1.8	Excellent	Pink, medium to coarse grained granite
12	0.3	Good	Brown-grey, well indurated quartzose sandstone.
13	0.3	Excellent	Dark red and light-grey cross bedded sandstone.
13	0.4	Good	Light greenish grey well compacted sandstone.
14	0.1	Poor	Dark grey-green fine to medium grained, mottled greenstone.
14	0.3	Poor	— ditto —
15	1.0	Good	Yellow and dark green chlorite schist
16	0.2	Poor	Dark grey-green, fine to medium grained, mottled greenstone.
17	0.8	Good	White to grey foliated quartzite and quartz granule conglomerate, schistose.
18	1.3	Excellent	Red brown massive bioclastic limestone.
19	2.0	Excellent	Dark grey massive argillaceous limestone.
20	0.7	Good	Medium brown to grey finely crystalline limestone.
21	1.1	Excellent	Reddish brown acidic to intermediate volcanic.
22	0.4	Good	Reddish brown, fine to medium grained sandstone.
23	0.6	Good	Grey-greeen, fine grained volcanic.

*Probability that the core represents bedrock at particular locality. Lithology of core samples, northeast of Newfoundland.

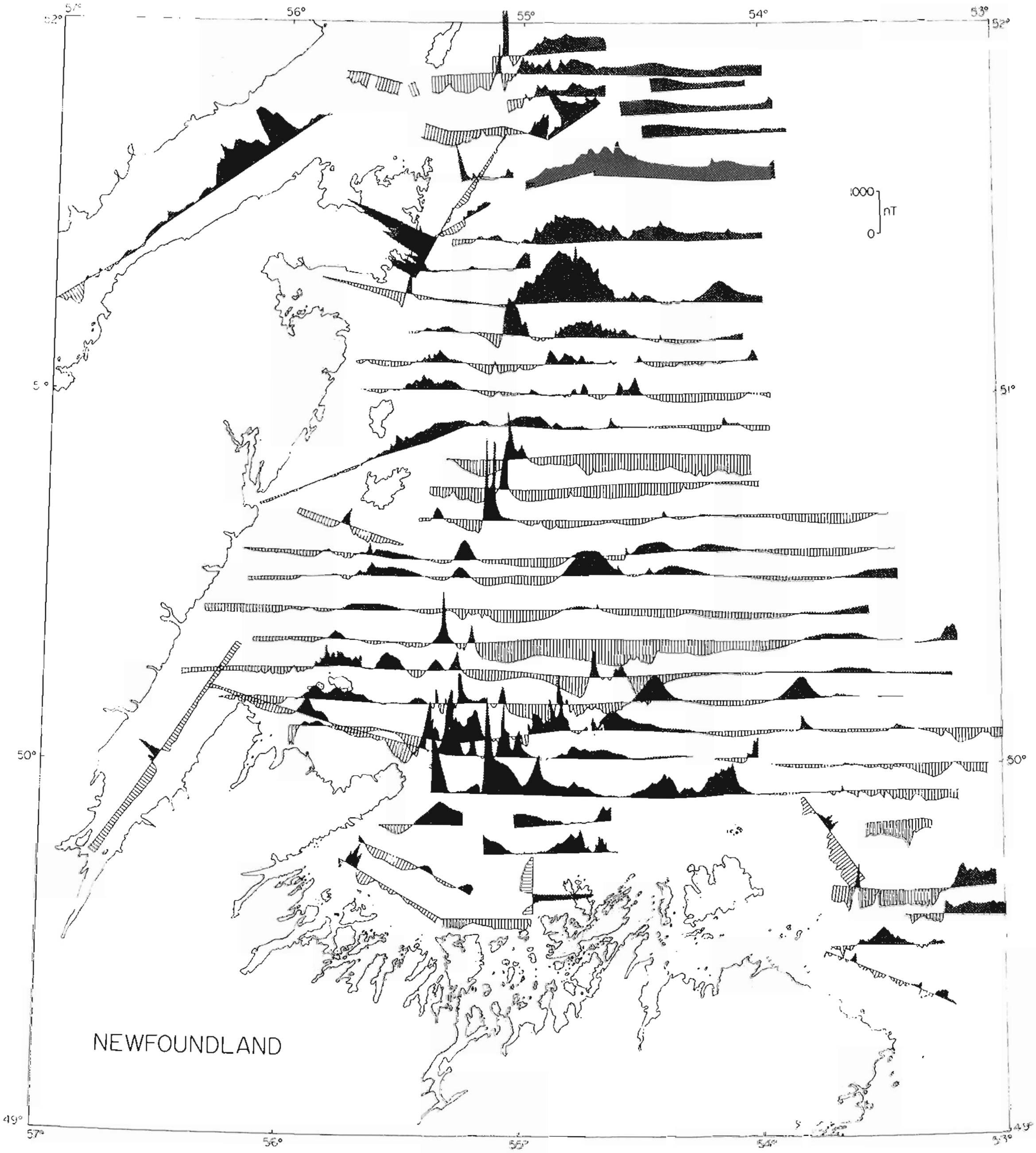

Figure 2. 4. Magnetic anomaly profiles, northeast of Newfoundland. Anomalies are plotted along selected ships tracks, darker shading indicates positive anomalies. Scale 1: 2 000 000.

east of 54½°W, the seismic penetration is sufficient to give a clear indication of the undulations of the bedding planes within it. Drill holes within the eastern transparent units produced cores of sandstone of presumed Carboniferous age. The most transparent unit is continuous south along the eastern edge of the survey area. This unit has been interpreted as the equivalent of the Pennsylvanian Barachois Group in southwestern Newfoundland (Riley, 1962) on the basis of a single core and the fact that it overlies a unit from which three limestone (Mississippian Codroy equivalent?) cores were obtained.

A seismic unit observed only briefly on one line apparently represents a sedimentary rock lying unconformably on the Carboniferous unit, but overlain by more extensive coastal plain type sedimentary rocks. Although neither of the post-Carboniferous units was sampled because of extensive thick surficial cover, the predominant one is designated Cretaceous-Tertiary on the map whereas the limited outlier may be of Jurassic age.

A second seismic unit into which good seismic penetration was obtained was outlined closer to shore south of 51°40'N. This unit was found to be continuous

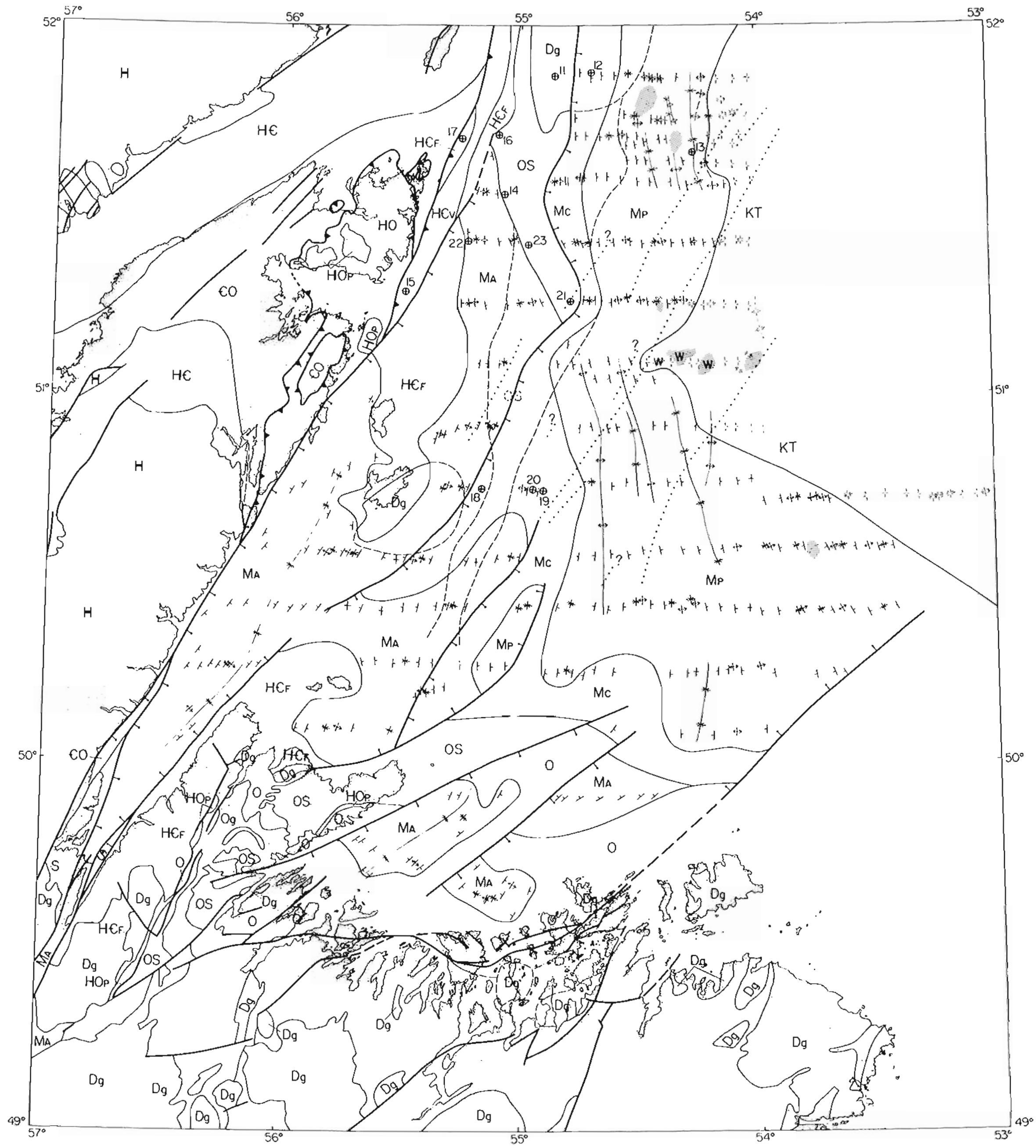

Figure 2. 5. Geological map, northeast of Newfoundland. Key to map is on opposite page. Scale 1: 2 000 000.

throughout the western half of the survey area and was traced southward into White Bay where bedrock exposures on Rouge Island, Cape Rouge Peninsula and Conche Peninsula belong to the Crouse Harbour and Cape Rouge Formations (Baird, 1966), equivalents of the Anguille Group of early Mississippian age. The entire western transparent unit has therefore been designated as early Mississippian on the geological map (Fig. 2. 5).

In the centre of the study area the two transparent units are separated by a unit in which there are only vague, weakly reflecting horizons. Three limestone cores were obtained across the width of this unit and the unit has been interpreted as an equivalent of the Mississippian Codroy and Windsor groups. Magnetic field data were used to delineate the boundaries between this vaguely transparent unit and the neighbouring opaque units because these boundaries are difficult to establish using seismic data alone. This was particularly so in Notre Dame Bay.

The high magnetic anomalies over seismically opaque units trending northeast out of Notre Dame Bay indicate offshore extension of the Ordovician-Silurian formations onshore. The truncation of a seismically transparent unit at the steep western margin of Notre Dame Bay suggests that that margin is fault controlled.

GEOLOGY OF THE NORTHEAST NEWFOUNDLAND SHELF

Geological boundary (defined, assumed or subsurface)

Fault (defined, assumed)

Bedding from seismic reflection records (surface unit, subsurface unit)

Anticline, anticlinal axis (surface unit, subsurface unit)

Syncline, synclinal axis (surface unit, subsurface unit)

Drill core location

Dike interpreted from magnetic field data

Diapiric zone; W indicates collapse/withdrawal feature

WESTERN AUTOCHTHONOUS ROCKS

SILURIAN

S Sedimentary and volcanic rocks southwest of White Bay

UPPER CAMBRIAN TO MIDDLE ORDOVICIAN

€O Limestone, dolomite, and in upper part, shale and slate (includes Doucers, St. George, Table Head and Goose Tickle Formations).

UPPER HADRYNIAN AND CAMBRIAN

H€ Quartzite, slate, limestone, dolomite locally minor basalt (includes Bateau, Lighthouse Cove, Bradore, Forteau and Hawke Bay Formations).

HELIKIAN AND OLDER

H Grenvillian rocks. Gneiss, schist, granitic and gabbroic rocks.

HARE BAY ALLOCHTHON

HADRYNIAN (?) TO MIDDLE ORDOVICIAN

HO Quartzose greywacke and slate; basalt, tuff, schist and amphibolite, structurally beneath peridotite sheets (HOp); Lower Ordovician basalt, tuff and black slate; Middle Ordovician melange locally at base of and within the allochthon (includes strata of Hare Bay Allochthon).

CRETACEOUS(?) AND TERTIARY

KT Siltstone, mudstone, sandstone

MISSISSIPPIAN AND PENNSYLVANIAN

MP Grey, red and brown sandstone, conglomerate and shale (mainly Barachois Group) (may include some Permian and lower Mesozoic strata)

MISSISSIPPIAN

Mc Red and grey conglomerate, sandstone, siltstone, shale, limestone and evaporites (mainly Codroy and Windsor Groups)

MA Grey, green and red conglomerate, sandstone, siltstone, shale (mainly Anguille, Crouse Harbour and Cape Rouge Formations)

DEVONIAN

Dg Granitic rocks (may include some Ordovician granites).

EASTERN AUTOCHTHONOUS ROCKS

ORDOVICIAN OR SILURIAN

OS Felsic and mafic volcanics, minor sedimentary rocks (includes Cape St. John Group)

LOWER ORDOVICIAN mainly

O Basalt, tuff, greywacke, slate, gabbro and peridotite; granites (Og) (includes Baie Verte, Mings Bight and Lushs Bight Groups).

HADRYNIAN AND CAMBRIAN mainly

H€P Psammitic and pelitic schists, locally basic schist and amphibolite derived from basalt (includes undifferentiated plutonic rocks); H€v, mainly chlorite schist derived from basaltic rocks and may include meta-gabbro and meta-peridotite (includes Fleur de Lys Supergroup).

A second sharp boundary between seismically opaque and transparent units within the bay having much the same trend as the western margin fault also correlates with a change in magnetic character and is also interpreted, therefore, as a normal fault although there is no corresponding bathymetric lineament. Within the downfaulted blocks the outlying erosional remnants are interpreted as lower Mississippian since they comprise the oldest transparent unit in the area and directly overlie the presumed Ordovician-Silurian rocks. Northeast from the eastern margin of Notre Dame Bay, the unit interpreted as Pennsylvanian has a sharp faulted boundary. This is an inshore continuation of the fault recognized farther seaward by Grant (1972), and trends towards a bathymetric lineament extending the Lukes Arm Fault seaward. The fault appears to be coincident with a decrease to the southeast in both the magnetic and Bouguer gravity anomalies.

Although there are few major local changes in gravity (Fig. 2. 3), the Bouguer anomalies over Notre Dame Bay and the area to the northeast are generally higher than over adjacent areas as recognized onshore by Miller and Deutsch (1973). Further analysis of these data together with the magnetic anomalies may indicate the offshore extent of the Ordovician unit which includes the Lushs Bight Group possibly representing Paleozoic oceanic crust (Strong, 1973). Initial inspection indicates its extension at least to $50\frac{1}{2}$°N, $53\frac{1}{2}$°W. The Bouguer gravity anomalies east of Fogo Island (Fig. 2. 2) are the lowest in the study area with the exception of the western margin. This indicates an offshore extension of the granites within the pre-Middle Ordovician metasedimentary rocks onshore (zone G of Williams *et al.*, 1972).

North of Baie Verte several trends can be recognized. A seismically opaque unit between Burlington Peninsula and Horse Islands is interpreted as an offshore extension of the Fleur de Lys Supergroup. This unit is overlain to the north of Horse Islands by rocks assigned to the Anguille, Codroy and equivalent units, but re-emerges on Groais and Bell islands. North of the islands, the seismically opaque unit continues towards Hare Bay, but the correlation of its boundaries with magnetic anomalies begins to break down. A prominent positive magnetic anomaly at the eastern boundary of the Fleur de Lys Supergroup runs from the vicinity of Baie Verte in the south, just east of Bell and Groais islands. Its only interruption is on the profile immediately south of Bell Island, coincident with a decrease in gravity, indicating an offshore extension of the Bell Island granite. However, north of Groais Island the magnetic anomaly gradually diminishes. At the southern edge of Hare Bay a sharp magnetic anomaly (GSC aeromagnetic map 7366G) interpreted in terms of an offshore peridotite sheet, part of the Hare Bay allochthon (Williams *et al.*, 1973; Williams and Smyth, 1974) is distinct from a linear magnetic anomaly paralleling the coast between Hare Bay and Cape Bauld (Fig. 2.3). That anomaly is somewhat speculatively interpreted as being continuous with the anomaly east of, and parallel to, Belle Isle, and the entire causative unit is interpreted as being part of the northern extension of the Fleur de Lys equivalent. The chlorite schist core at site 15, therefore, is initially interpreted as equivalent to similar rocks at the eastern boundary of the Fleur de Lys Supergroup southwest of Baie Verte (Neale and Nash, 1963) rather than the equivalent of the Goose Cove Formation (Williams *et al.*, 1973) as part of a steeply dipping submarine allochthonous slice, although the two units may indeed be approximate stratigraphic correlatives.

About 40 km east of the coast between Hare Bay and Cape Bauld the magnetic field is high, as are the Bouguer anomalies. The continuity of positive magnetic anomalies into the area along a trend just east of north from Cape St. John has strengthened the drill hole evidence that an equivalent of the Cape St. John volcanics is part of the seismically opaque unit in the area that also extends to the south beneath the carboniferous cover. The gravity and magnetic character of this area east of the Hare Bay allochthon is similar to that of the Notre Dame Bay region where the Lushs Bight Group has been interpreted in terms of ancient oceanic crust (Strong, 1973). The high gravity and magnetic fields in the northern part of the study area, therefore, may delineate an additional area of ancient oceanic crust as the source of the Hare Bay allochthon. This possibility will be examined during detailed analysis of the gravity and magnetic data.

On the northernmost seismic line east of Belle Isle, a remarkably smooth topped morphological unit yielded a granite drill core although there was only a minimal decrease in gravity over the unit. Visually, the granite appeared to be typically Acadian (Devonian).

Within the Pennsylvanian unit several diapiric structures were found. Those diapirs marked on the geological map with a W (because they are collapse structures from which the salt(?) had presumably been withdrawn) were subject to a detailed seismic survey to determine their shape and tectonic history. The latter remains to be resolved but they were roughly circular in shape. Although these were the only well developed collapse structures, other diapirs also showed some minor signs of similar salt withdrawal.

The most definite feature of the gravity map is the line along which the Bouguer anomalies change from negative to positive. This line does not follow specific geological boundaries because it is dominated by the effect of deeper structures. The line may be related to the eastern edge of the Paleozoic craton upon which the western autochthonous rocks rest, although the trend is deflected by such local effects as the gravity highs over the structurally highest elements of the Hare Bay allochthon.

Localized magnetic highs of 200nT amplitude correlate extremely well between adjacent magnetic profiles to indicate a significant pattern of northeasterly trending features. Although the area is traversed by a number of trans-Atlantic cables (U.S. Naval Ocean. Off. Chart 14023), their trend, number and anticipated anomaly do not coincide with our observations. The magnetic anomalies have been interpreted in terms of dykes having a similar trend to many others offshore around Newfoundland (e.g. Papezik *et al.*, 1975).

References

Baird, D.M.
1966: Carboniferous rocks of the Conche-Groais Island area, Newfoundland; Can. J. Earth Sci., v. 3, p. 247-257.

Canadian Hydrographic Service
1975: Natural resource maps of free-air and Bouguer gravity anomalies, magnetic anomalies and total magnetic field. Map areas 18602, 18600, 15092, 15090. Maps and index available from CHS Chart Distribution Office, Dept. of the Environment, Ottawa K1A 0E6.

Grant, A.C.
1972: The continental margin off Labrador and eastern Newfoundland — Morphology and geology; Can. J. Earth Sci., v. 9, p. 1394-1430.

Grant, A.C. and Macnab, R.F.
1974: Bedrock geology, northeast Newfoundland; Geol. Surv. Can., Paper 74-1B, p. 153.

Macnab, R.F.
1974: Natural resource maps east of Newfoundland; Geol. Surv. Can., Paper 74-1B, p. 156.

McKeown, D. L.
1975: Evaluation of the Huntec('70) hydrosonde deep tow seismic system; Bedford Inst. Rept. BI-R-75-4, 115 p.

Miller, H. G. and Deutsch, E. R.
1973: A gravity survey of eastern Notre Dame Bay, Newfoundland; *in* Earth Science Symposium on Offshore Eastern Canada ed. P. J. Hood, Geol. Surv. Can., Paper 71-23, p. 389-406.

Neale, E. R. W. and Nash, W. A.
1963: Sandy Lake (east half), Newfoundland; Geol. Surv. Can., Paper 62-28, 40 p.

Papezik, V. S., Hodych, J. P., and Goodacre, A. K.
1975: The Avalon magnetic lineament: a possible continuation of the Triassic dike system of New Brunswick and Nova Scotia; Can. J. Earth Sci., v. 12, p. 332-335.

Riley, G. C.
1962: Stephenville map-area, Newfoundland; Geol. Surv. Can., Mem. 323, 72 p.

Strong, D. F.
1973: Lushs Bight and Roberts Arm Groups of central Newfoundland: possible juxtaposed oceanic and island-arc volcanic suites; Geol. Soc. Am., Bull., v. 84, p. 3917-3928.

Weaver, D. F.
1968: Preliminary results of the gravity survey of the Island of Newfoundland; Dominion Observatory Gravity Map Series, nos. 53-57.

Williams, H., Kennedy, M. J., and Neale, E. R. W.
1972: The Appalachian structural province; *in* Variations in Tectonic Styles in Canada ed. R. A. Price and R. J. W. Douglas, p. 181-261.

Williams, H. and Smyth, W. R.
1974: The Hare Bay allochthon, northern Newfoundland; Geol. Surv. Can., Paper 74-1A, p. 3-6.

Williams, H., Smyth, W. R., and Stevens, R. K.
1973: Hare Bay allochthon, northern Newfoundland; Geol. Surv. Can., Paper 73-1A, p. 8-14.

Project 680047

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The existence of permafrost in the geologic past generally is deduced from "fossil" evidence of permafrost ice features, such as ice-wedge pseudomorphs. The purpose of this note is to report on relict ice wedges, which still preserve the original Pleistocene ice, in the glacially deformed sediments of Hooper Island, N.W.T. The age of permafrost may have a bearing on the occurrence of subpermafrost gas hydrates.

Hooper Island, N.W.T.

Hooper Island, 160 km north of Inuvik, N.W.T., is about 5 km long, 2 km wide, and the highest point reaches 50 m above sea level. Till overlies marine and fluvial sands, beyond the age limit of radiocarbon dating. The sediments have been glacially deformed (Mackay, 1971, Fig. 15; Rampton, 1971). Hooper Island lies beyond the limit of the "classical" Wisconsin. Glacial deformation took place more than 40 000 years ago (Mackay *et al.*, 1972).

Ice Wedges

In March 1974 the steeply eroded north bluff of Hooper Island was examined for permafrost structures which commonly are obscured by summer thaw and slumping. At an altitude of 7 to 10 m above sea level and for a horizontal distance of 10 m, several frost crack structures were found on the north bluff (Fig. 3.1). The site was revisited in early June 1975 when it was possible to expose a fresh section by digging. Four of the features were excavated; others were seen, but were inaccessible. In each of the excavated structures, ice veins were still preserved at the bottom (Fig. 3.2) thus showing that the frost crack structures were relict Pleistocene ice wedges. The ice veins numbered from 3 to 8; they were from a few millimetres to one centimetre in width. Each vein was separated from adjacent veins by a thin sand vein of about the same thickness as the ice veins. The ice veins tapered downwards and were subparallel to the sides of the structure. Some of the ice veins had horizontal bubbles and a central seam, both typical of ice vein substructures formed by infreezing of water between two parallel, closely spaced vertical freezing planes. The tops of the veins were convex, just as similar present day ice veins are convex when they thaw. The stratigraphic succession from top to bottom showed: an olive brown sand exceeding 5 m in thickness (a of Fig. 3.1); sharp erosional (?) contact; 1 m of light grey sand (b of Fig. 3.1); 1 to 1.5 m of organic rich lacustrine (?) silts and sands (c of Fig. 3.1); segregated ice, with an ice content (weight basis) exceeding 500 per cent. The lower portions of the ice-wedge structures occurred both in the grey sands (b of Fig. 3.1) and lacustrine (?) deposits (c of Fig. 3.1). The upper portions, with the subsidence structures, were primarily in the olive brown sands (a of Fig. 3.1). The subsidence structures, which are not easily seen in Figure 3.1 because of lack of contrast, are graben-like faulting which results when an ice wedge thaws.

Interpretation

The preservation of the original ice veins in the frost structures shows, conclusively, that the features are relict ice wedges. It is obvious that the ice wedges

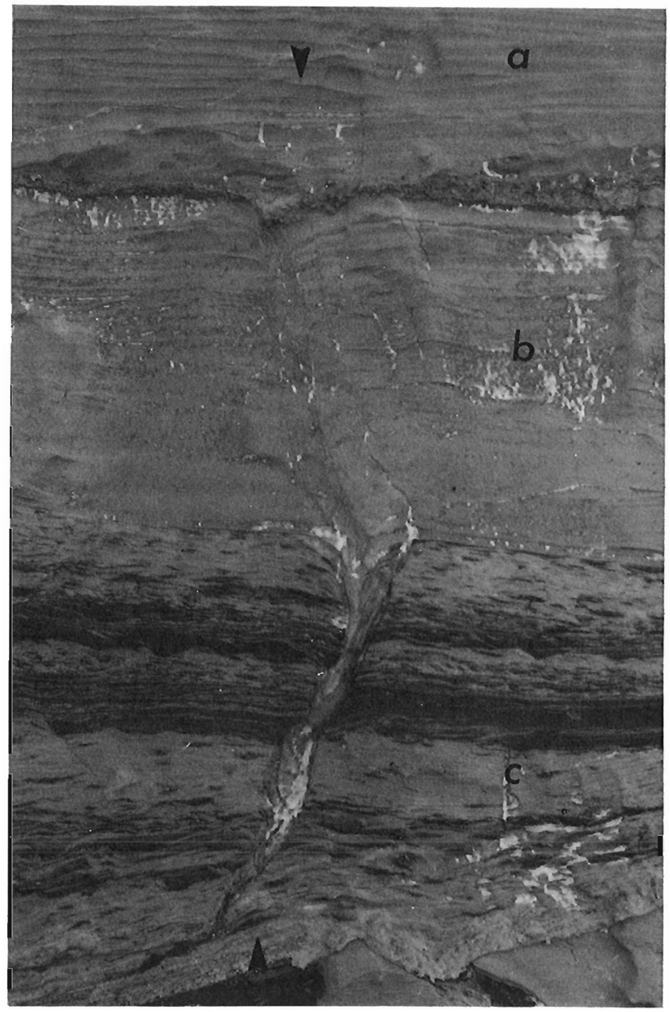

Figure 3.1. Relict Pleistocene ice wedge, Hooper Island, N.W.T. Ice-wedge ice is still preserved at the bottom. The height of the structure, between the two arrows, is 3 m (see text).

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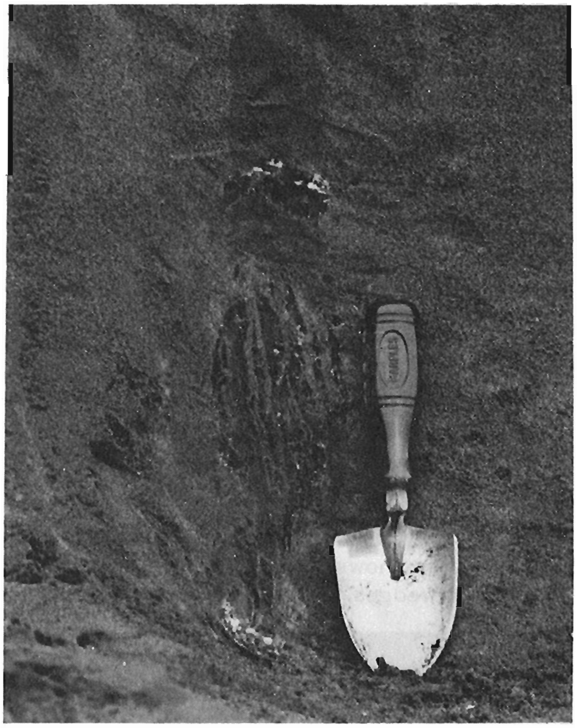

Figure 3. 2. Ice veins (dark) between sand veins (light) at the bottom of a relict ice wedge. The trowel is about 20 cm long. The veins are to the left of the trowel.

are of syngenetic origin, having grown concurrent with sedimentation. It is also apparent that permafrost was present in the sediments during sedimentation and has been present ever since then. The subsidence structures resulted when slight thaw of the upper parts of the wedges occurred subsequent to the deposition of the grey and olive sands (Fig. 3. 1).

Conclusion

Relict syngenetic ice wedges are present at Hooper Island, N. W. T. Similar features, usually in pseudomorph form, have been observed in many parts of the world (e. g. Matthews, 1974). The minimum age for the preservation of permafrost appears to date back to at least early Wisconsin, because Hooper Island lies beyond the limit of the "classical" Wisconsin glaciation, and the sediments were glacially deformed prior to 40 000 years ago. By geomorphic inference, it seems likely that permafrost has been present on much of Richards Island, the Pleistocene offshore islands, and Tuktoyaktuk Peninsula for the same period.

References

Mackay, J. R.
1971: The origin of massive icy beds in permafrost, Western Arctic Coast, Canada; Can. J. Earth Sci., v. 8, p. 397-422.

Mackay, J. R., Rampton, V. N., and Fyles, J. G.
1972: Relic Pleistocene permafrost, Western Arctic, Canada; Science, v. 176, p. 1321-1323.

Matthews, J. V. Jr.
1974: Quaternary environments at Cape Deceit (Seward Peninsula, Alaska): Evolution of a tundra ecosystem; Geol. Soc. Am., Bull., v. 85, p. 1353-1384.

Rampton, V. N.
1971: Quaternary geology, Mackenzie Delta and Arctic Coastal Plain, District of Mackenzie; *in* Rept. of Activities, Part A, Geol. Surv. Can., Paper 71-1A, p. 173-177.

4. SUSPENDED MATTER IN COASTAL AND SHELF WATERS, SOUTHWESTERN VANCOUVER ISLAND, BRITISH COLUMBIA

Project 750108

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The distribution, composition, and ultimate fate of suspended matter in the world oceans have been of increasing interest in recent years. Not only does a knowledge of particulate solids help in defining the sedimentary and biologic budgets within the ocean, but these data also can aid in understanding the transport and effect of various pollutants.

An overwhelming amount of the research on

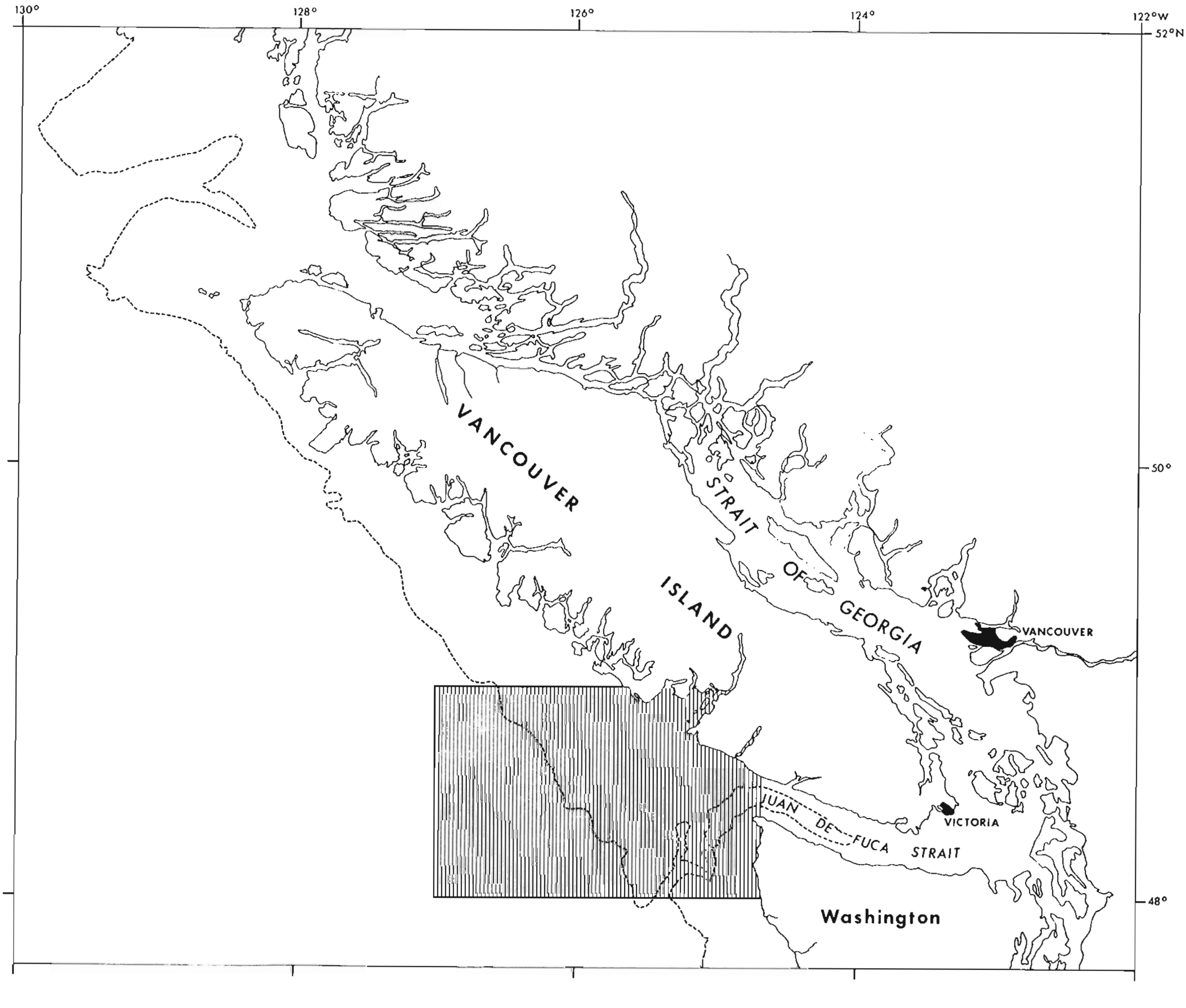

Figure 4.1. Southwest British Columbia: shaped area represents the survey site down in Figs. 4.2 and 4.3. Lightly dashed line is the 100-fathom (180-m) depth contour, and indicates the shelf and Juan de Fuca Canyon.

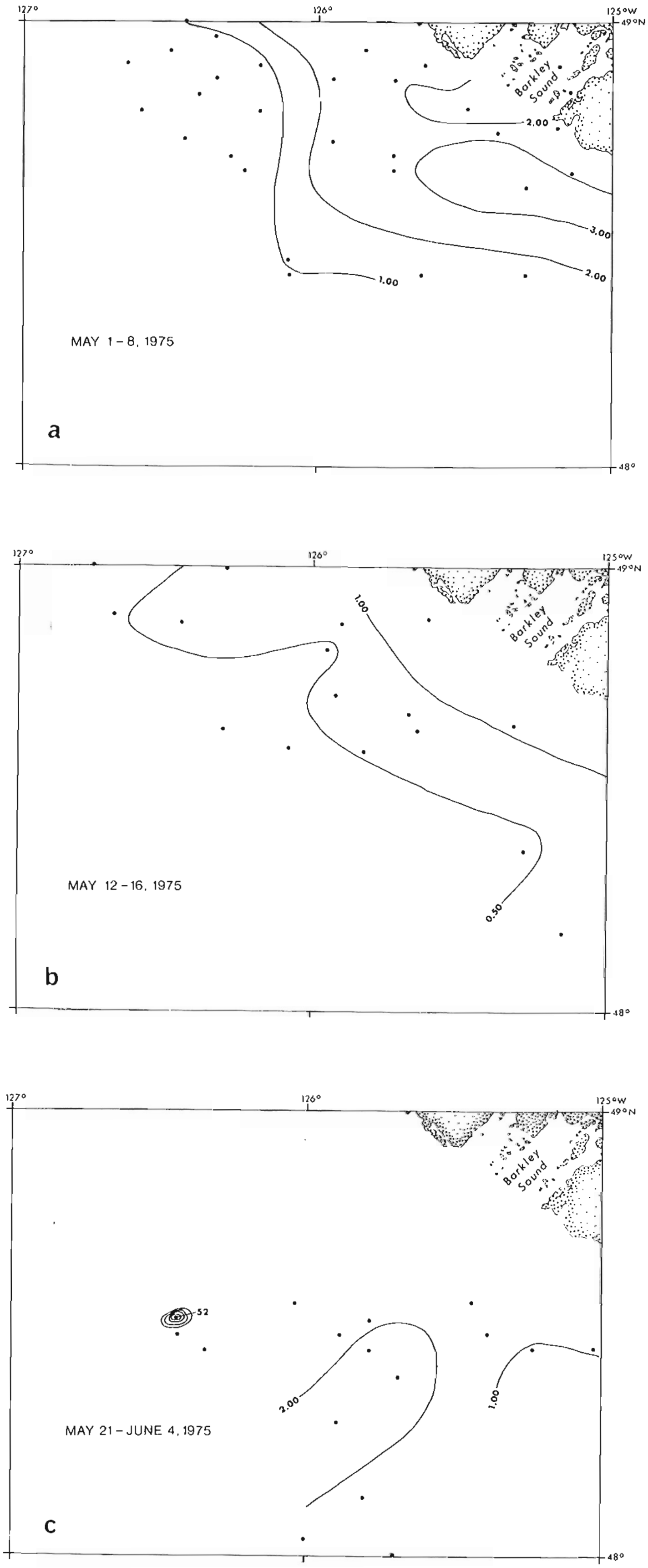

Figure 4. 2

Distribution of suspended matter in the surface waters in May 1-8 (a), May 12-16 (b) and May 21-June 4 (c), 1975. Values are expressed in mg/l.

suspended solids has been done in the open ocean (e. g., *see* Lisitzin, 1972). Even continental margin studies (e. g., Emery and Milliman, in press) have emphasized distributions on the continental slope and rise. Clearly, however, the most important area for particulates is coastal and shelf waters, these being nearest terrestial sources as well as the site of high biologic activity. Many suspended matter studies have been limited to a single series of observations, commonly made during a single cruise; conclusions drawn from such studies necessarily negate seasonal or shorter variations, which, as will be seen, can be far greater than spatial differences.

This paper presents initial results in the study of the sedimentary regime in the coastal and shelf waters off southern British Columbia. Data were collected seaward of southwestern Vancouver Island (Fig. 4. 1) during two periods in 1975 (May and late August). Particulate samples in May were collected during three consecutive cruises, while those in August were taken within a 24-hour period.

Water samples were collected from the sea surface using a polyethylene bucket. In addition, at six stations in August, water samples were taken at various levels within the water column using van Doren bottles. Temperature and salinity of the samples were measured upon recovery. Water samples collected in May were filtered through pre-weighed Nucleopore filters (openings of 0. 45 micron), while August samples were filtered through pre-weighed paired Millipore filters (nominal openings of 0. 45 micron). Upon return to the laboratory, all samples were washed, air-dried, and weighed for total suspended content. Subsequent analyses will include measurement of combustible organics, noncombustible organics (such as opaline frustules), and terrigenous grains; these results will be reported at a later time.

May 1975

Each of the three May cruises defined a unique pattern of suspended particle distribution. During the first cruise (May 1-8), concentrations in the surface waters were highest on the shelf (particularly north of Juan de Fuca Strait, where concentrations reached 3. 49 mg/l). In contrast, concentrations in the nearshore waters (such as Barkley Sound) generally were less than 2 mg/l, while those on the slope ranged from 0. 6 to 0. 9 mg/l (Fig. 4. 2a). One week later (May 12-16), concentrations were strikingly depleted (Fig. 4. 2b) with maximum values now occurring in the inner shelf seaward of Barkley Sound (1. 42 mg/l). Shelf concentrations were generally between 0. 6 and 1. 0 mg/l,

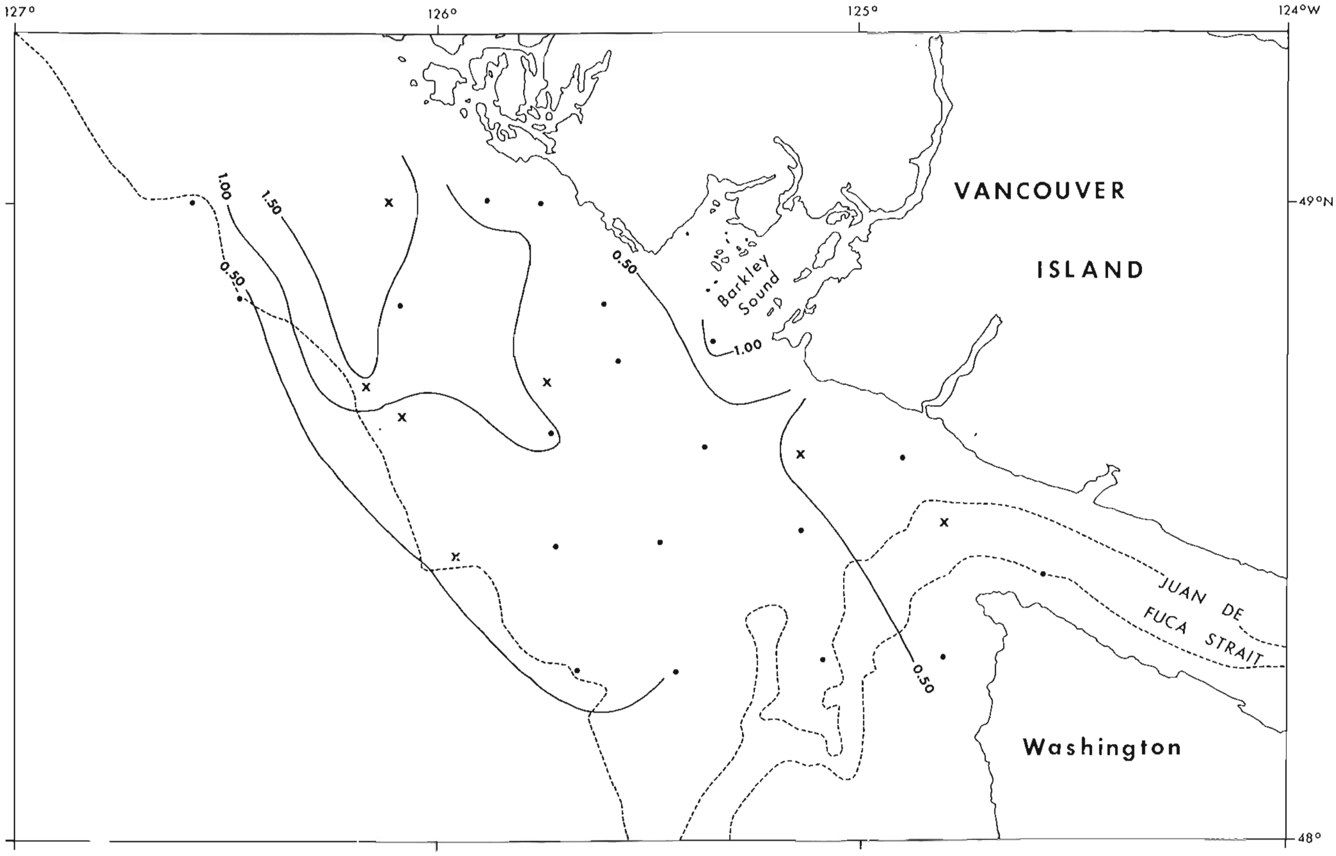

Figure 4. 3. Distribution of suspended matter in the surface waters during August 24-25, 1975. X's mark stations in which suspended samples were taken within the water column. Contour intervals are the same as in Fig. 4. 2. Dashed line is the 180-m isobath roughly defining the shelf-slope break.

while slope contents ranged from 0. 35 to 0. 64 mg/l. Subsequently (May 1-June 4), however, total suspended content on the shelf increased to 1 to 2 mg/l (Fig. 4. 2c); the general lack of overlap between this and the prior two cruises, however, precludes a more thorough comparison. One sample collected on the last cruise contained 52. 45 mg/l of suspended matter; this sample was found to be composed almost entirely of copepods.

August 1975

During August 24-25 concentrations were highest on the outer shelf west of Barkley Sound (highest value — 1. 68 mg/l) and lowest in Juan de Fuca Strait and the coastal waters off Washington (0. 23 and 0. 34 mg/l, respectively). Most of the area, however, displayed values between 0. 5 and 1. 0 mg/l (Fig. 4. 3).

Suspended matter concentrations remained relatively constant in the upper part of the water column; decreases in the upper 20 to 30 m seldom exceeded 20 per cent. Apparently these waters are well mixed (as indicated by temperature and salinity data), thus explaining the relatively even distributions of particulates. Although available data are sparse, particulates apparently decrease markedly between 30 and 50 m and presumably remain low below this depth (Fig. 4. 4). At one station (at the mouth of Juan de Fuca Strait), however, the concentration increased near the bottom, possibly indicating reworking of bottom sediment by current activity.[1]

Preliminary Conclusions

Judging from the green to brownish-green colour of the material collected on the filters, most of the suspended matter in the surface waters off Vancouver Island in May and August was organic; subsequent analyses will confirm (or deny) this interpretation. Similarly, the marked and relatively rapid fluctuations in suspended matter noted in the three May cruises are suggestive of a nonconservative process, such as biologic activity. Apparently such rapid turnovers are required to obtain the high biologic productivity noted in these waters (Anderson, 1964). Short-term

[1] Although echo sounding capabilities on this cruise were minimal, extreme care was taken to prevent the water bottles from striking the bottom. Thus the high near-bottom concentration of suspended matter at the entrance to Juan de Fuca Strait appears to be real and not the result of sampling.

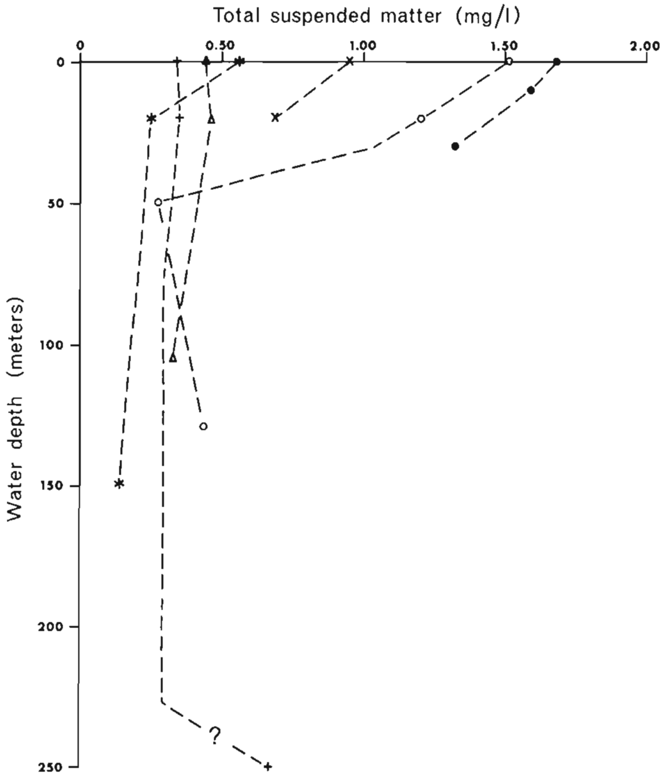

Figure 4. 4. Distribution of suspended matter within the water column, August 1975.

variations, however, seriously limit the ability to integrate widely spaced observations in obtaining a suspended matter budget; clearly, the picture of particulate distribution gained from the first cruise in May would have differed greatly from that based only on the second cruise. Hopefully subsequent cruises can investigate further both short-term and seasonal fluctuations, as well as define subsurface distributions.

Acknowledgments

I thank the crews of the C. S. S. Vector and C. F. A. V. Laymore for their help in collecting the May and August samples. Numerous individuals aided in the collection of the suspended matter samples. I thank particularly B. D. Bornhold (Ottawa) and Wendy Symonds (Vancouver) for their help in preparing and weighing the filters.

References

Anderson, G. C.
1964: The seasonal and geographical distribution of primary productivity off the Washington and Oregon coasts; Limnol. Oceanogr., v. 9, p. 284-302.

Emery, K. O. and Milliman, J. D.
Suspended matter in surface waters: Influence of river discharge and of upwelling; Orville Bandy Memorial Volume (in press).

Lisitzin, A. P.
1972: Sedimentation in the world ocean; Soc. Econ. Paleontol. Mineral., Spec. Publ. 17, 218 p.

5. # SAND AND GRAVEL DEPOSITS OFF SOUTHWESTERN VANCOUVER ISLAND, BRITISH COLUMBIA

Project 750108

John D. Milliman
Terrain Sciences Division, Vancouver

Much has been written about the economic potential on continental shelves: oil, heavy minerals, even diamonds (Emery and Noakes, 1968). The sedimentary deposit with perhaps the greatest economic value, however, may well be sand (0.06 to 2 mm) and gravel (coarser than 2 mm). This material is basic to the construction industry, being used predominantly in road construction, concrete and asphalt aggregates, and as fill. In Canada alone, 228 million short tons of sand and gravel were used in 1973 valued at 187 million dollars (Stonehouse, in press).

Because of the volume and weight of the material involved, sand and gravel are best "mined" from near the sites of their ultimate use. Since the sites of construction in British Columbia (which ranks third among the provinces in use of sand and gravel) are mostly coastal, a major source of the sand and gravel is Pleistocene coastal deposits. Unfortunately, urban "development" commonly spreads over these deposits, thus preventing continued utilization. Since it is not realistic to move or alter sites of construction or habitation, it is necessary to look for other nearby areas which might serve as future sources of sand and gravel. One obvious site is the offshore.

Particularly high concentrations of sand and gravel occur on the continental shelf off southwestern Vancouver Island (Fig. 5.1); this area was studied in May and June 1975 when more than 400 sediment samples were taken on a two-mile grid, using a Shipek grab-sampler. For the purpose of this report, only sand and gravel are reported although subsequent reports will discuss other aspects of these sediments.

The shelf off southwestern Vancouver Island is marked by several prominent topographic basins on the inner shelf, as well as Juan de Fuca Canyon (Fig. 5.2). On the seaward side of the basins, the shelf is particularly shallow, 60 to 90 m in most places. Seaward and south of these shoals, the shelf generally averages 100 to 150 m in depth (Fig. 5.2).

Sand and gravel are the major components throughout the shelf; the only exceptions are in the basins and in Juan de Fuca Canyon where muds predominate (Fig. 5.3). In fact much of the middle shelf is covered with deposits in which sand and gravel total more than 98 per cent of the surficial sediment. High concentrations also occur in coastal areas just north of Juan de Fuca Strait.

Although much of this coarse material is sand, sediments containing more than 50 per cent gravel occupy much of the middle shelf, particularly just seaward of the inner shelf basins (cf. Figs. 5.2 and 5.4). Another gravel-rich area occurs on the outer shelf and upper-

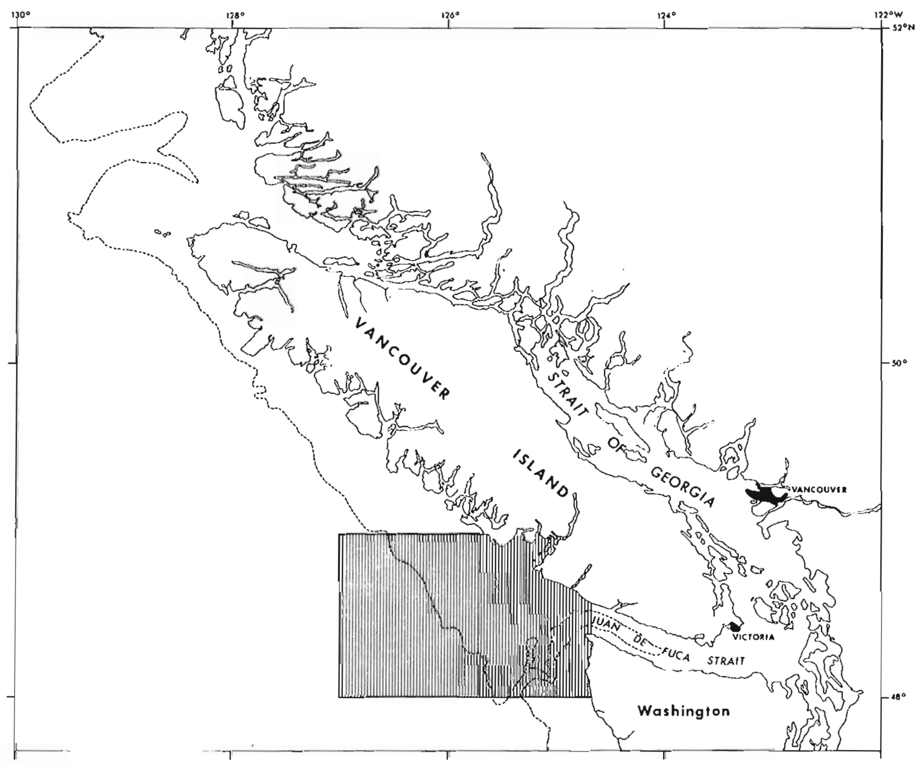

Figure 5.1

Area off southwestern Vancouver Island studied in May and June, 1975. Dashed line is the 180-m isobath roughly defining the shelf-slope break.

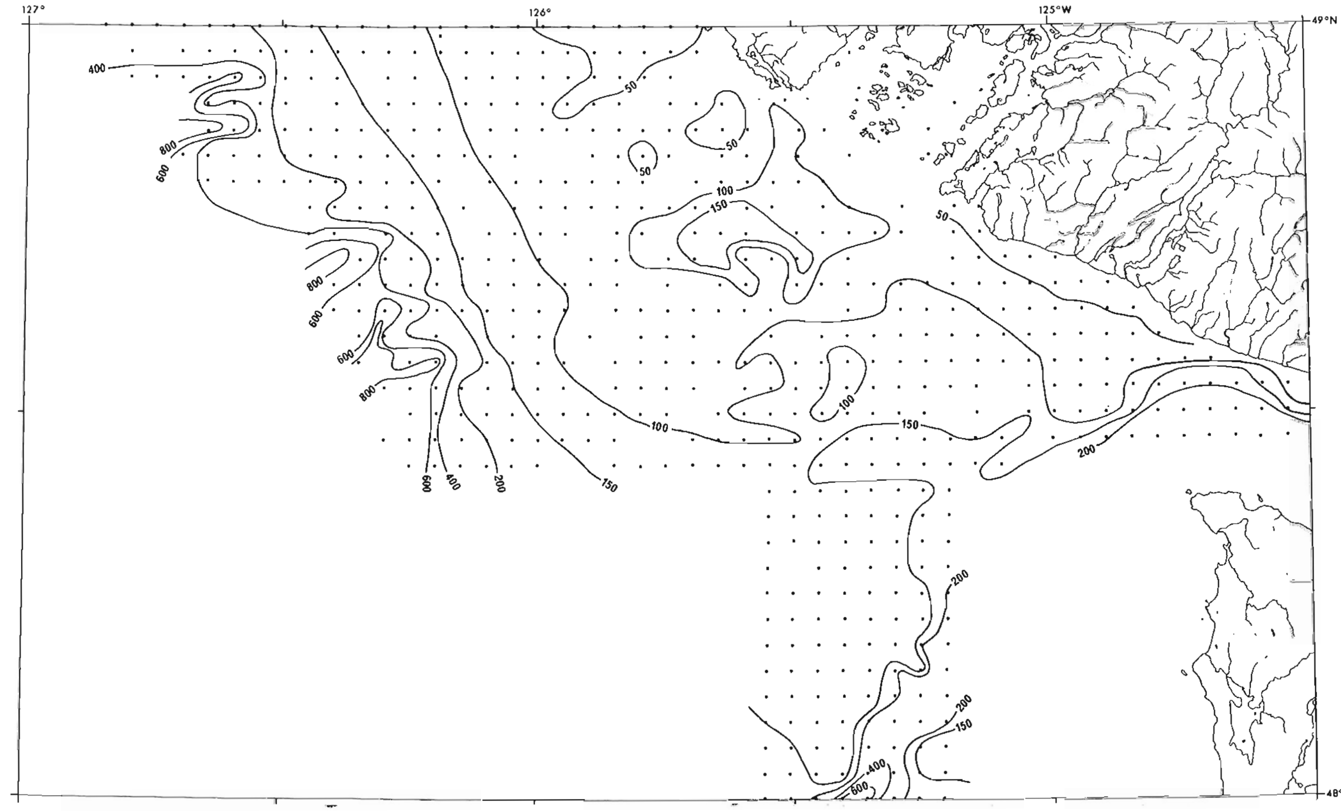

Figure 5. 2. Preliminary bathymetric chart of the shelf and upper slope off southwestern Vancouver Island based on echo soundings made in May and June, 1975. A far more detailed chart will be available in 1976 from the Canadian Hydrographic Service. Isobaths are in metres; dots indicate sample sites.

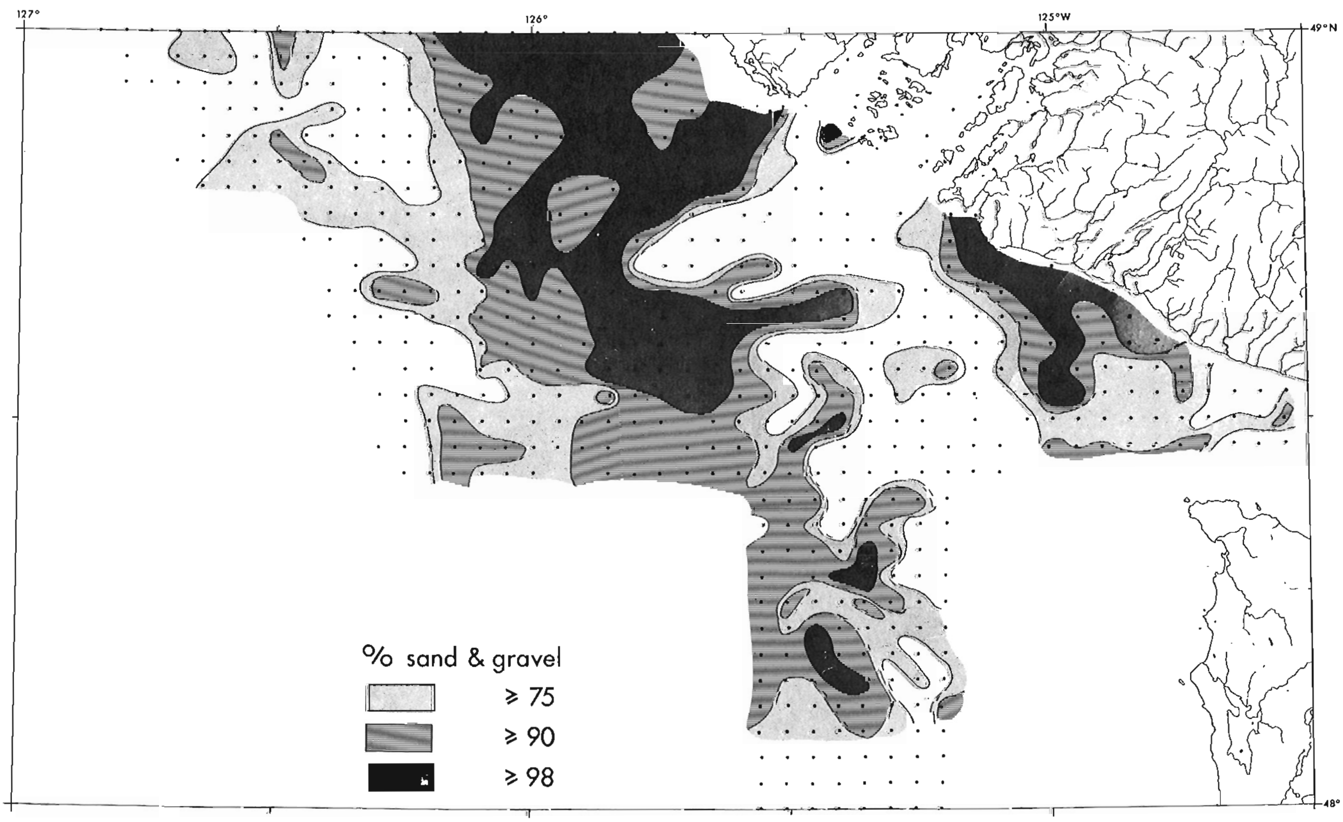

Figure 5. 3. Distribution of sand and gravel on the continental shelf off southwestern Vancouver Island.

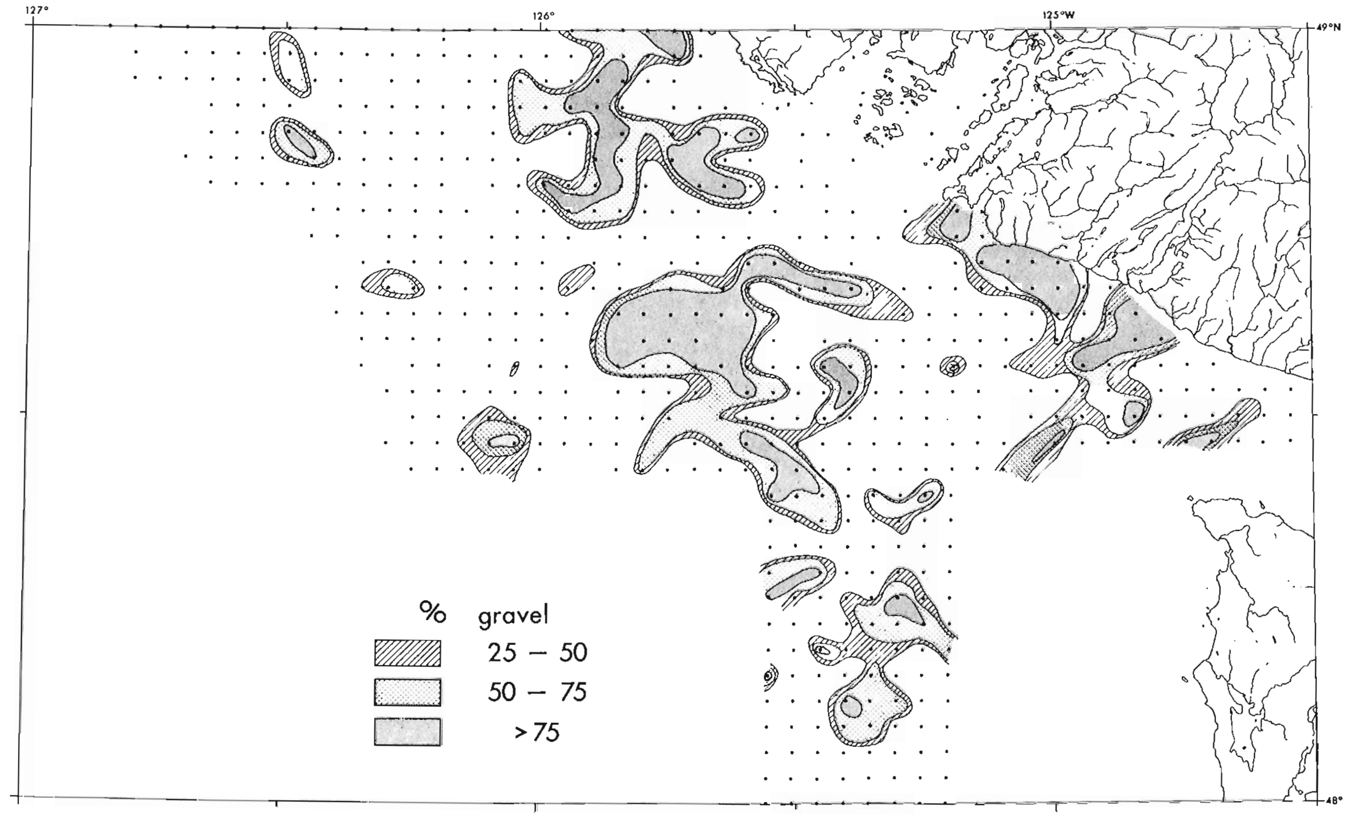

Figure 5.4. Distribution of gravel (coarser than 2 mm) on the continental shelf off southwestern Vancouver Island.

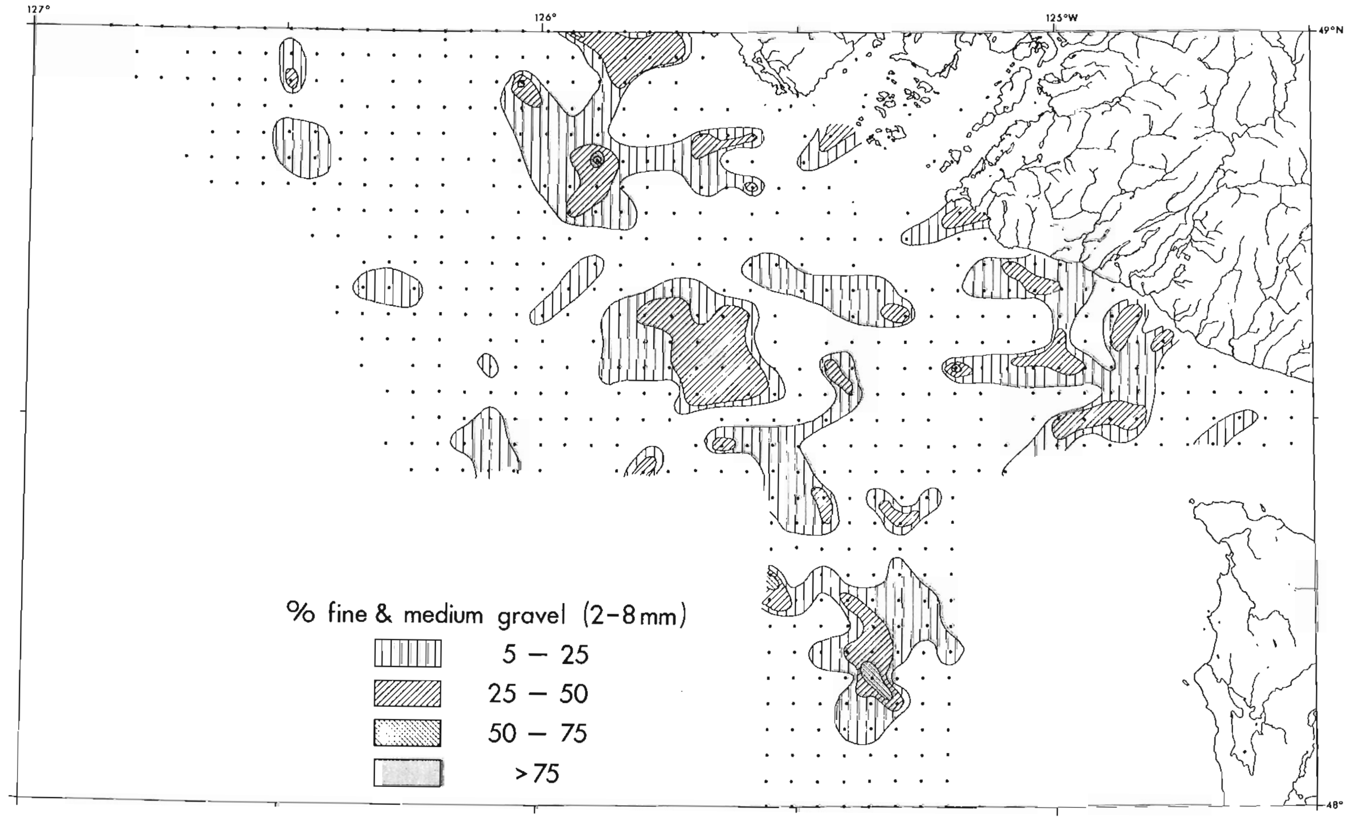

Figure 5.5. Distribution of fine gravel (2 to 8 mm) on the continental shelf off southwestern Vancouver Island.

west slope. Much of the shelf gravel, however, is composed of cobble-size material (coarser than 16 mm). The finer gravel, which undoubtedly has greater economic value, occurs in distinct patches on the middle shelf as well as in nearby coastal areas, with local concentrations reaching 40 per cent (Fig. 5.5). This material is much purer than many land-based commercial pits or offshore deposits which have been suggested as possible future sources (e.g. Schlee, 1968). Assuming a thickness of 5 m (which probably is an underestimation), the volume of sediment containing more than 25 per cent fine gravel alone exceeds 3 billion m^3, or more than enough to supply all of Canada for 15 years; using 1973 prices, this would be worth nearly 3 billion dollars!

Preliminary data suggest that gravels on the outermost shelf and uppermost slope are derived from nearby outcrops (and therefore can be classified as residual). In contrast, the shelf gravels are relict in age and were deposited during the last lower stand of sea level. The more poorly sorted gravels (particularly those containing large cobbles) probably were deposited as glacial till and subsequently winnowed by bottom currents. In contrast, many of the sand and fine gravel sediments may have been deposited or reworked by glacier-fed rivers, thus explaining their generally finer size and better sorting.

Subsequent studies on the texture and composition of these sediments should help clarify further their origin and economic value.

I thank Wendy Symonds and Greg Seid for their help with the sediment analyses and the crew of the C.S.S. Vector for aid in sample collection.

References

Emery, K.O. and Noakes, L.C.
1968: Economic placer deposits of the continental shelf; Tec. Bull. ECAFE, v. 1, p. 95-111.

Schlee, J.
1968: Sand and gravel on the continental shelf off the northeastern United States; U.S. Geol. Surv., Circ. 602, 9 p.

Stonehouse, D.H.
Sand and gravel; *in* Canadian Minerals Yearbook, 1973 (in press).

6. INVENTAIRE DES DEPOTS DE SURFACE DE L'ILE BANKS, DISTRICT DE FRANKLIN

Projet 740065

J.S. Vincent et R.C. Gauthier
Division de la science des terrains

Les études de la répartition et des caractéristiques des dépôts de surface de l'île Banks, entreprises au cours de l'été 1974, (Vincent *et al.*, 1975) se sont poursuivies au cours de l'été suivant. Le soutien logistique, soit l'utilisation d'un hélicoptère Bell 206B pour la durée du séjour sur l'île et le ravitaillement par un avion Twin Otter, a été fourni par l'Etude du plateau continental polaire (EPCP).

Le programme d'identification des dépôts meubles, au niveau de la reconnaissance, a été complété pour l'ensemble de l'île. Suffisamment de données ont été recueillies pour subdiviser les différents types de sédiments en unités lithologiques homogènes qui tiennent compte des caractères génétiques, texturaux et morphologiques de dépôts. On a concentré les efforts pour identifier et distinguer, dans le secteur ouest de l'île, les sédiments de la formation de Beaufort de ceux plus récents (tills, alluvions) dérivés de cette formation.

On a recueilli systématiquement des renseignements sur la morphologie, la topographie, la texture, l'épaisseur, les conditions de drainage, le microrelief, les processus actifs et stables et le pergélisol, afin de caractériser les différentes unités lithologiques.

Un programme de forage à faible profondeur a permis de recueillir une collection d'échantillons *in situ*, à l'intérieur de la zone de pergélisol. Des foreuses Haynes, pour les sédiments fins et Winkie, à tête de diamants, pour les sédiments plus grossiers, ont été utilisées. La granulométrie, le type de glace et la teneur en eau de chaque échantillon ont été déterminés au cours des opérations, dans un laboratoire monté au camp de base. Une légende détaillée, accompagnée de la synthèse des données recueillies pour chaque unité lithologique, s'associera aux cartes des dépôts meubles de l'ensemble de l'île. Le tout devrait être disponible avant la fin de 1976.

Une partie des travaux d'été ont porté à définir le mode de déglaciation de la dernière avancée wisconsinienne, qui recouvrait le secteur est et sud de l'île. Bien que cette glace n'ait recouverte qu'une fraction de l'île, des surfaces importantes on été modifiées par l'écoulement des eaux de fonte glaciaires.

Des coupes dans les dépôts pléistocènes en bordure des côtes est et ouest ont montré jusqu'à quatre séquences distinctes de tills, séparés par des intervalles aquatiques et aériens. Par un contrôle des datations isotopiques au carbone 14 et de la description détaillée de chaque unité, on tentera d'élaborer une séquence générale de l'histoire glaciaire de l'île.

Un programme d'échantillonnage des tills se trouvant en surface ou à partir de coupes permettra de décrire les divers faciès des tills et d'élaborer des relations d'âge entre différents tills. L'étude de l'altération chimique des gabbros permettra également de compléter ces relations.

Référence

Vincent, J.S., Tucker, C.M., et Edlund, S.A.
1975: Surficial geology inventory, Banks Island, District of Franklin; *dans* Report of Activities, Part A, Comm. géol. Can., Etude 75-1A, p. 431-434.

Project 730031

Brian D. Bornhold
Terrain Sciences Division

Introduction

Grab sampling and gravity coring were carried out in the central and eastern Arctic (Fig. 7.1) on an opportunity basis from the Canadian Coast Guard icebreakers Labrador and D'Iberville between September 4 and 29, 1975. In eastern Viscount Melville Sound, off southern Bathurst Island, 91 Shipek grab samples were obtained (Fig. 7.2); 38 grab samples and one gravity core were collected in Strathcona Sound on northern Baffin Island. Temperature and salinity profiles also were obtained at 33 stations in Strathcona Sound.

Eastern Viscount Melville Sound

Sediments collected in water depths exceeding approximately 140 m in general are characterized by a 1- to 2-cm surface layer of soft, olive-brown clay overlying a compact, olive-grey clay. Pebbles up to 10 cm in length are common in these offshore samples with sand increasing in abundance near the islands. Ophiuroids and a variety of polychaetes are the most common biological components in these sediments.

Nearshore samples are composed largely of coarse sand, pebbles, and some shell fragments. In places nearshore samples consisted of a layer of coarse clean sand or pebbles overlying a sandy, pebbly mud. This surface layer is interpreted as a lag deposit, the result of winnowing by bottom currents.

Pebbles in nearshore samples invariably are covered with a red alga, a brown-black manganese oxide (?) coating, a variety of Bryozoa, serpulid polychaete tubes, and commonly barnacles. Bivalve molluscs and brachiopods are common constituents of nearshore samples.

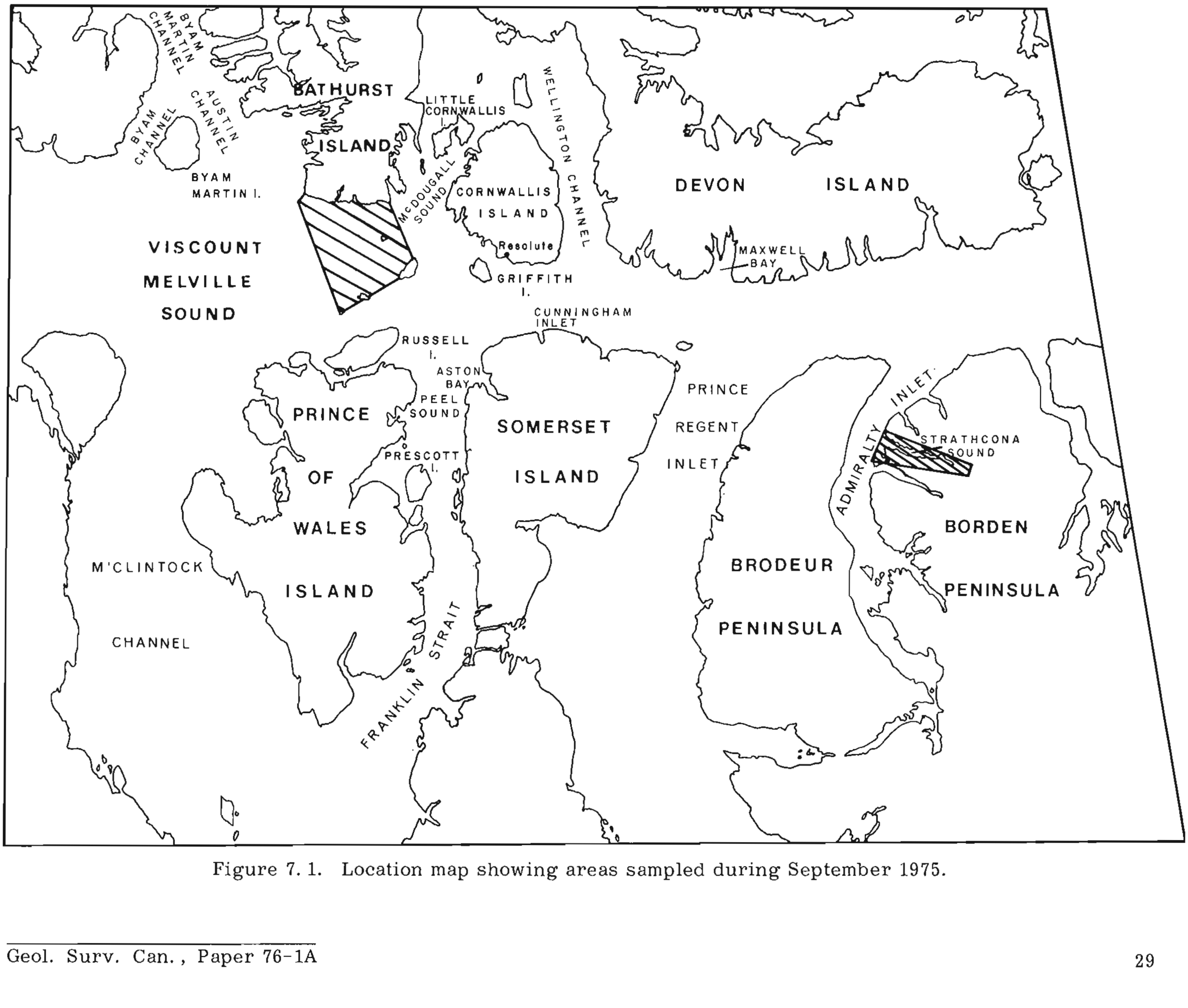

Figure 7.1. Location map showing areas sampled during September 1975.

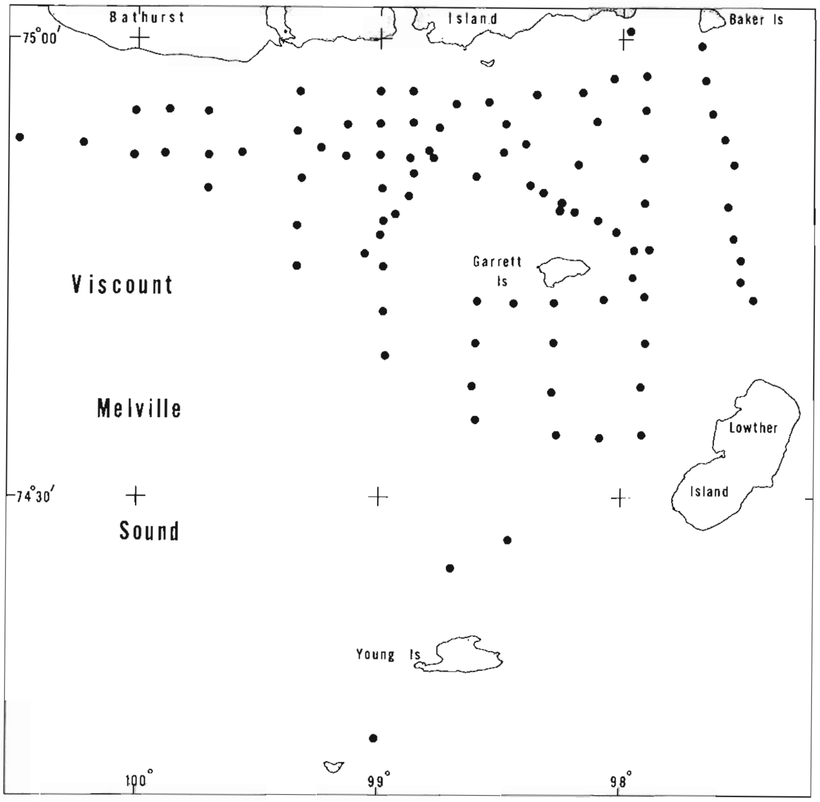

Figure 7. 2.

Grab sample locations in eastern Viscount Melville Sound.

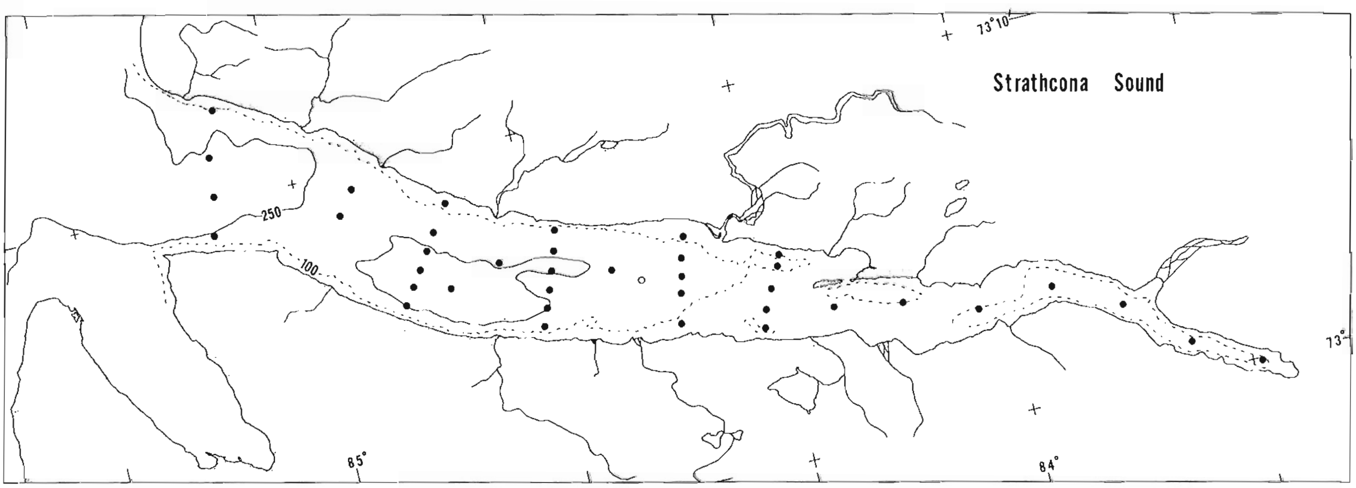

Figure 7. 3. Grab sample locations and gravity core station (open circle) in Strathcona Sound; 100-m and 250-m Bathymetric contours are indicated.

Strathcona Sound

Strathcona Sound is a narrow, deep fiord extending over 50 km into Borden Peninsula on Baffin Island from Admiralty Inlet. It is bounded on both sides by precipitous cliffs over 300 m in height. The sound is divided into two primary basins, one extending 23 km from a narrow constriction to the head of the sound, and the other occupying the central part of the inlet (Fig. 7. 3). Maximum water depths exceed 175 m in the upper basin and 320 m in the main basin.

Sediments in the upper basin consist of a soft brown clay overlying intensely mottled, compact grey clay with abundant black streaks throughout. Macroscopic fauna was restricted to a few species of polychaetes.

Sediments in the main basin and at the entrance to the sound are similar to those in the upper basin but are lighter brown in colour and contain a somewhat more diverse fauna. Samples taken on the sills between the basins and in the mouth of the sound contained a significantly higher concentration of sand and pebbles than those from within the basins.

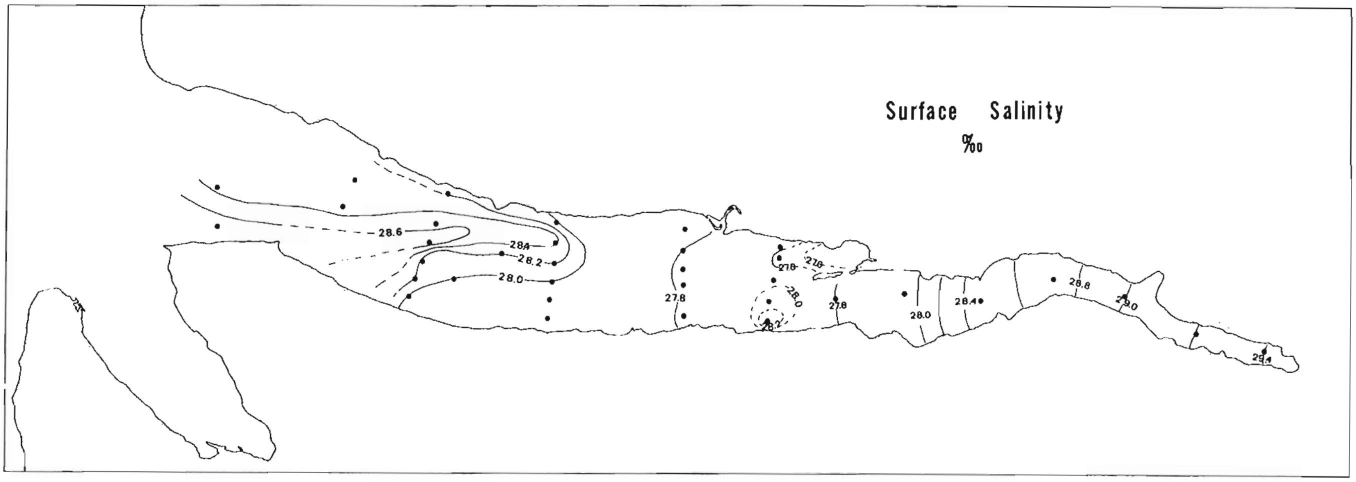

Figure 7. 4. Distribution of surface water salinity, Strathcona Sound.

One 63-cm gravity core was obtained near the centre of the main basin. It penetrated 2 to 3 cm of soft, light brown clay and 60 cm of brown-grey, mottled, very compact clay.

Temperature and salinity were measured at five-metre intervals in the upper part of the water column throughout the sound using a YSI salinometer. Each station displayed a marked temperature and salinity discontinuity between 15 and 20 m.

Surface water temperatures were generally low (-0. 9 to 0. 0°C) at the mouth of the sound and high (1. 0°C) in the upper part. Surface salinities displayed a more complex distribution (Fig. 7. 4). A tongue of high salinity (up to 28. 7 ‰) enters the south side of the inlet and extends towards the north coast. The central part of the sound is characterized by salinities of approximately 28. 0 ‰ . Beyond the narrow constriction, towards the head of the sound, salinity values increase steadily to more than 29. 4 ‰ . The lack of significant interchange of surface waters between the upper and lower parts of the sound is inferred from the markedly higher salinities and temperatures near the head of the sound.

Acknowledgments

The excellent co-operation of Mr. M. A. Hemphill, Hydrographer-in-Charge, Eastern Arctic Surveys, enabled us to carry out the sampling program. The assistance and enthusiasm of Mr. Hemphill and his survey team are greatly appreciated. I wish to thank the Canadian Coast Guard for permitting us to carry out this work from their icebreakers. In particular, I should like to thank Captain Paul W. Tooke of the C. C. G. S. Labrador and Captain Henri St. Pierre of the C. C. G. S. D'Iberville and their officers and crews for their co-operation and help. Jim Savelle assisted in the collection of samples and his help is greatly appreciated.

8. LE QUATERNAIRE DE LA COTE-NORD DE L'ESTUAIRE MARITIME DU SAINT-LAURENT: SECTEURS DE RIVIERE-AUX-GRAINES, SHELDRAKE ET MINGAN

Convention de recherche E.M.R. 1135/D13-4-172/75

Jean-Marie Dubois[1]
Division de la science des terrains

Localisation

Pendant l'été 1975 environ 700 milles carrés (1800 km) de terrain ont été couverts sur la partie sud des cartes de Rivière-aux-Graines (22 I/6 E), de Sheldrake (22 I/7) et de Mingan (22 I/8). Ce relevé fait suite à celui de l'été 1974 alors que les secteurs de Rivière-aux-Graines (22 I/6 W) et de Matamec (22 I/5) avaient été cartographiés (Dubois, 1975). Les 60 milles (97 km) de côtes ainsi représentés se localisent dans les cantons de Coopman, Bailloquet, Touzel, Margane, Fornel, Rocamadour, Mingan et l'ouest de Cugnet. Six petites agglomérations sont échelonnées le long de la côte: Sheldrake, Rivière-au-Tonnerre, Magpie, Rivière-Saint-Jean, Longue-Pointe-de-Mingan et Mingan (fig. 8.1).

Accessibilité

Les villages sont reliés entre eux par une section de la route 138 qui n'est pas encore intégrée au réseau routier provincial. Cette route de gravier se prolonge et vers l'est et vers l'ouest; vers l'est une mauvaise route asphaltée rejoint Havre-Saint-Pierre tandis que vers l'ouest une quinzaine de milles (24 km) de bonne route de pierre concassée a été terminée jusqu'à la rivière Manitou à l'été 1974. Six chemins d'exploitation forestière pénètrent de 2.5 à 6.5 milles (4 à 10.5 km) à l'intérieur des terres: chemin de Rexfor à l'ouest de la rivière au Tonnerre, chemin du lac Maloney (Soc. de Conservation de la Côte-Nord), chemin de Rexfor à l'ouest de la rivière Saint-Jean et chemins à l'est de la rivière Magpie, à l'est de la rivière Saint-Jean et à l'ouest de la rivière Petit Manitou. La majorité des cours d'eau sont difficilement portageables à cause du grand nombre de rapides et de chutes dès l'embouchure. Par contre, la rivière Saint-Jean est canotable sur la majeure partie de son cours et les rivières Mingan et Manitou le sont aux hautes eaux. Une excellente piste bétonnée construite par l'USAF existe à Longue-Pointe-de-Mingan ainsi que des quais pour hydravions aux lacs des Eudistes, Maloney et Patterson.

Géologie et physiographie

La presque totalité du territoire se situe dans la province structurale de Grenville; seulement l'extrême sud-est fait partie de la plate-forme du Saint-Laurent. Deux grands secteurs physiographiques et altimétriques peuvent y être retenus: le plateau Laurentien et son piedmont.

Le plateau passe d'une moyenne de 500 pieds (150 m) d'altitude au sud à une moyenne de 1000 pieds (300 m) au nord sur une distance de 8 à 11 milles (13 à 18 km). Il est principalement constitué d'anorthosite et d'un peu de mangérite (Sharma et Franconi, 1973) et l'énergie du relief est d'environ 300 pieds (90 m).

Le piedmont s'élève à 500 pieds d'altitude sur une distance de 3.5 à 8 milles (5.5 à 13 km) et l'énergie du relief y est généralement de moins de 100 pieds (30 m). A l'ouest de la rivière à la Chaloupe il est constitué de granite et à l'est de mangérite; aucun affleurement de sédiments paléozoïques (calcaire ou grès calcareux de la formation de Mingan) n'a été relevé sur la terre ferme mais l'île du Havre de Mingan à peine à 1000 pieds (300 m) au large en est constitué. Cette île est d'ailleurs un élément de trois alignements de cuestas à pendage sud entre la baie Saint-Laurent et Longue-Pointe-de-Mingan; l'alignement central semble même se prolonger sous les eaux vers les Sept-Iles.

Répartition des formations meubles

La majeure partie du plateau est formée d'affleurements rocheux surmontés d'un mince sol; des placages de till ou de till remanié peuvent parfois être visibles à l'est de Sheldrake. Les plus grandes formations meubles sont contiguës aux cours d'eau principaux. Comme stratigraphie on retrouve souvent trois combinaisons:

1) sédiments fluviaux et diamicton d'éboulis (vers les versants des vallées)/sédiments lacustres ou estuariens
2) sédiments d'épandages fluvioglaciaires et diamicton d'éboulis/sédiments lacustres ou estuariens
3) sédiments littoraux/sédiments estuariens et (ou) diamicton de glissements dans ces mêmes sédiments/sédiments marins/roche en place. Cette troisième combinaison se localise à la sortie des rivières sur le piedmont.

Sur le piedmont alternent deux types de stratigraphie selon que le substratum rocheux est proéminent ou en dépression:

1) lorsque le substratum rocheux est proéminent: dépôts littoraux associés avec des dépôts organiques de tourbière/roche en place. Ces dépôts sont généralement assez minces (moins de 10 pieds — 3 m) et localisés sous forme de petites terrasses ou de surface de recouvrement entre les affleurements; parfois ils représentent des vastes terrasses reposant sur des sédiments

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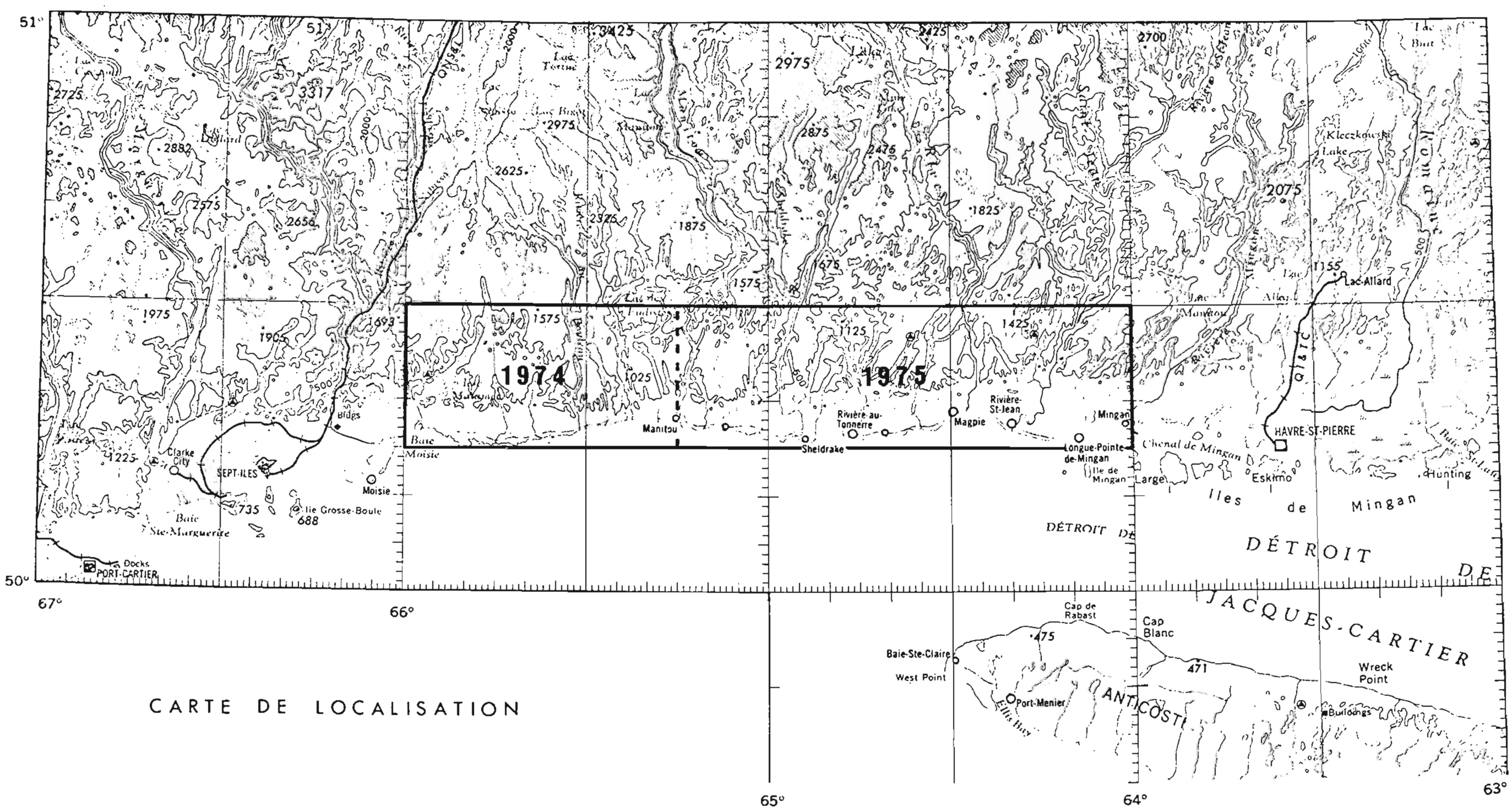
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1155
Lac Tortue
Lac-Allard
Kleczkowski Lake
Clarke City
SEPT-ILES
Baie Ste-Marguerite
735
Ile Grosse-Boule
688
Docks
PORT-CARTIER
Moisie
Baie Moisie
Bldgs
Manitou
Sheldrake
Rivière-au-Tonnerre
Magpie
Rivière-St-Jean
Mingan
Longue-Pointe-de-Mingan
Ile de Mingan
Large
Chenal de Mingan
Eskimo
HAVRE-ST-PIERRE
Hunting
Iles de Mingan
DÉTROIT DE
DÉTROIT
DE
JACQUES-CARTIER
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Baie-Ste-Claire
West Point
Port-Menier
Ellis Bay
ANTICOSTI
Wreck Point
Buildings

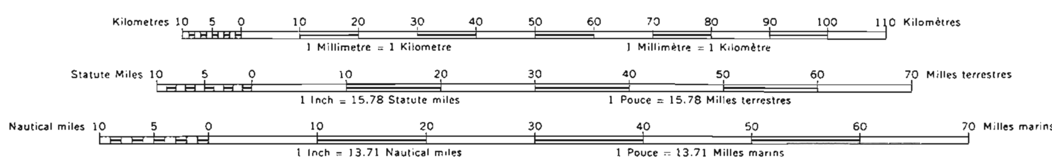
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Figure 8. 1.

Figure 8. 2

Moraine à blocs sur le versant est de la rivière Matamec.

Figure 8. 3

Section morainique au sud du lac Manitou.

estuariens (entre les rivières au Tonnerre et Chipitagun par exemple);

2) lorsque le substratum rocheux est en dépression surtout vers l'embouchure des grands cours d'eau comme les rivières Manitou, à la Chaloupe, Sheldrake, au Tonnerre, Magpie, Saint-Jean et Petit Manitou — Mingan: sables éoliens parfois/sédiments littoraux — dépôts organiques de tourbière/sédiments sableux et graveleux deltaïques/sédiments fins estuariens/sédiments fins marins/roche en place (parfois visible). L'ensemble de ces formations peut atteindre facilement plus de 150 pieds (46 m) de puissance (Rivière-Saint-Jean) et couvrir des superficies de plus de 45 milles (115 km) carrés. Les cours d'eau y ont fortement divagué en y laissant d'impressionnants réseaux de méandres. Les talus d'érosion fluviale ainsi mis en place de même que les talus d'érosion marine furent et sont encore le siège de glissements soit en paquets, soit de solifluxion, soit à ravins (gully flows).

Le littoral actuel à l'ouest de la rivière Magpie ressemble à celui qui a été décrit précédemment (Dubois, 1975). A l'est de cette rivière, on est en présence d'une plage de sable moyen qui s'étale sur plus de 24 milles (40 km) jusqu'à la rivière Mingan; c'est le secteur des deltas juxtaposés des rivières Magpie, Saint-Jean et Manitou-Mingan.

Complexe morainique de Manitou-Matamec

Par photo-interprétation nous avons observé ce qui semble être un vaste complexe morainique à la limite nord de la zone d'étude. Ce complexe formé de minces bourrelets discontinus et parfois juxtaposés s'étire sur au moins une centaine de milles (160 km) entre les rivières Matamec et Romaine. Il est associé à un grand nombre de plaines d'épandages fluvio-glaciaires et aussi à quelques compartiments de fusion surtout vers le lac Manitou (63°55' — 50°27').

Etant peu accessible, nous n'avons pu le rejoindre qu'à deux sites très différents:

1) rivière Matamec entre la rivière Tchinicaman et le lac Key. A cet endroit il est généralement constitué d'un chaos de galets sub-anguleux et de blocs anguleux pouvant atteindre une dimension de 5 x 10 pi.

(1.5 x 3.0 m) (fig. 2). Parfois il présente une stratigraphie complexe de granulométrie variant des blocs aux sables fins.

2) entre les lacs Manitou et Gros-Diable sur le cours de la rivière Manitou. A cet endroit nous sommes en présence d'un bourrelet associé avec un complexe de kames et kettles (fig. 3).

Une exploration plus poussée devra être faite de ce complexe morainique que nous proposons de nommer moraine de Manitou-Matamec.

Epandages fluvio-glaciaires

Tous les épandages fluvio-glaciaires qui ne sont pas associés directement au complexe morainique et parfois même ces derniers se situent vers les 350 pieds (107 m) d'altitude, altitude qui correspond approximativement avec la limite supérieure des sédiments fins estuariens. De tels épandages ont été localisés principalement près des rivières au Tonnerre, Saint-Jean et Petit Manitou.

Glissements de Rivière-Saint-Jean

A l'embouchure de la rivière Saint-Jean, sur le côté du village, nous avons localisé un important secteur de glissements anciens et actuels couvrant environ 6 milles (15.5 km) carrés. Une partie du village lui-même serait assis sur des sédiments glissés. Dans les sédiments estuariens on retrouve de temps en temps jusqu'à Magpie des bancs de galets provenant de glissements anciens de ce secteur. De telles unités ont aussi été trouvées à l'embouchure de la rivière Tortue et sur le cours moyen de la rivière Saint-Jean.

Carapace ferrugineuse

Une carapace ferrugineuse continue de 3 à 5.5 pieds (0.9 à 1.7 m) d'épaisseur a été observée à la surface de toutes les formations sablonneuses et graveleuses fines ou moyennes, à la surface des terrasses marines ou fluviales. Nous supposons que la formation de cette carapace est assez rapide puisqu'elle peut être trouvée sur des terrasses qui étaient à peine à une vingtaine de pieds d'altitude (6 m).

Sols polygonaux de l'île Nue (île de Mingan)

Des structures géométriques ont été découvertes à un seul endroit de la côte, sur l'île Nue. Ces figures couvrent la majeure partie de l'île jusque sur les basses terrasses littorales et se sont formées sur les levées de plage. L'île Nue est la seule île dénudée, ce qui suppose des conditions climatiques particulières à cet endroit.

Références

Dubois, J.M.M.
1975: Le Quaternaire de la Côte-Nord de l'estuaire maritime du Saint-Laurent; *dans* Rept. of Activities, Pt. A, Comm. géol., Can., Etude 75-1A, p. 403-405.

Sharma, K.N.M. and Franconi, A.
1973: Grenville Project 1970; Geology of the rivière Magpie, rivière Saint-Jean and rivière Romaine area, Duplessis County, Quebec; min. Rich. nat., Québec, rapp. GM-28405, 66 p.

9. GEOLOGICAL VARIABILITY OF MARINE DEPOSITS, OTTAWA-ST. LAWRENCE LOWLANDS

Project 740054

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Terrain Sciences Division

Introduction

As a background for continuing studies of geologic and geotechnical variability of Pleistocene marine deposits, field work carried out during the summer of 1975 was aimed at publishing an inventory of clay deposits and landslides of a major part of the Ottawa Valley. To this end an area was selected about which much information was already available and within which unmapped areas might be examined within a single field season. In addition to geological mapping and other studies within this region, there has been a great deal of interest among geotechnical engineers and planners because of the known instability and the frequency of engineering problems associated with the so-called "Leda Clay". As urban expansion continues in the National Capital Region, the importance of the problems involving landslides, as well as potential embankment and foundation failures, continues to grow. The maps produced in this study give a regional view that will enable users to recognize those areas where Leda Clay is present, where large concentrations of landslides have occurred in the past, and where special precautions ought to be taken in planning for future development.

The project consisted of two parts. The first part was a compilation of existing 1: 50 000 scale surficial maps and the correlation of map-units across map boundaries. The source material was chiefly published and unpublished maps of the Geological Survey of Canada and of the Ontario Ministry of Natural Resources. The second part of the project was the mapping at 1: 50 000 scale of principally Leda Clay in areas where no information was available. Roughly one-third of the area shown in Figures 9. 1a and 9. 1b was mapped by the authors during the summer of 1975. In the western and northern parts of the region, mapping was carried out within the limit of marine submergence as defined by Gadd (1973); maps are in preparation. The large end moraine trending south of Hawkesbury (Logan, 1863) was chosen arbitrarily as the eastern boundary of mapping, because it provides a convenient break in the distribution of Leda Clay. The southern boundary selected was latitude 45°15', for south of this latitude Leda Clay occurs in thin discontinuous patches and there are fewer landslides.

A drilling program carried out in conjunction with the mapping was used to give support information on the stratigraphy and geotechnical properties of the fine grained deposits.

Regional Compilation

The results of this twofold operation are illustrated in a greatly reduced and simplified map (Figs. 9. 1a and 9. 1b) that provides a general overview of the surficial materials in the Ottawa Valley. The approximate marine limit and major areas of sensitive clay and silt are shown, along with the geological materials that may underlie and overlie them. Also, the areas where the most extensive landslides occur within this sensitive material are indicated.

Marine Limit

The marine limit has been used as a guide to show where fine grained deposits might be expected. Differential depression and uplift during and following the retreat of the ice sheet has made it impossible to assume a single elevation for the maximum level of submergence throughout the Ottawa Valley. Deltas in valleys open to the Champlain Sea basin were used to define the marine limit. The elevation of the delta nearest to the upper limit of marine erosion and deposition was chosen as the maximum level of submergence. The variation in height above present sea level of the marine limit is from 175 m near Pembroke, to 210 m north of Ottawa, to 228 m in the valley of Riviére du Nord near Montreal (Gadd, in press). The control level southwest of Ottawa of 168 m is based on marine fossils found in a beach deposit at this elevation (Richard, 1974).

Map-units

Rock, till, and gravel. In the compilation of Figure 9. 1, geologic deposits that are older than Leda Clay are grouped into a single unit. This unit includes the materials that have no association with the marine submergence, specifically bedrock, till, and gravel.

Throughout the region the composition of the till is in the range of sandy to silty till. Very poorly sorted, bouldery lag gravel occurs where till has been modified by wave action. Other gravels, mainly of glaciofluvial origin are characterized by poor sorting and abrupt changes in grain size within and between strata. Till and gravel units within the region studied are generally thin (less than 3 m); however, some gravels are thick enough to have been quarried extensively for construction material. At and below the elevation of the marine limit, wave-modified deposits of gravel, till, and even of shaly bedrock, may contain fossil shells of the littoral environment of the Champlain Sea.

Fine grained deposits. The fine grained deposits include the clay to silt facies deposited during the transgression and regression of the Champlain Sea and the fluviodeltaic clay and silt subsequently deposited in fresh water. The major part of the material shown in Figure 9. 1, however, is Leda Clay.

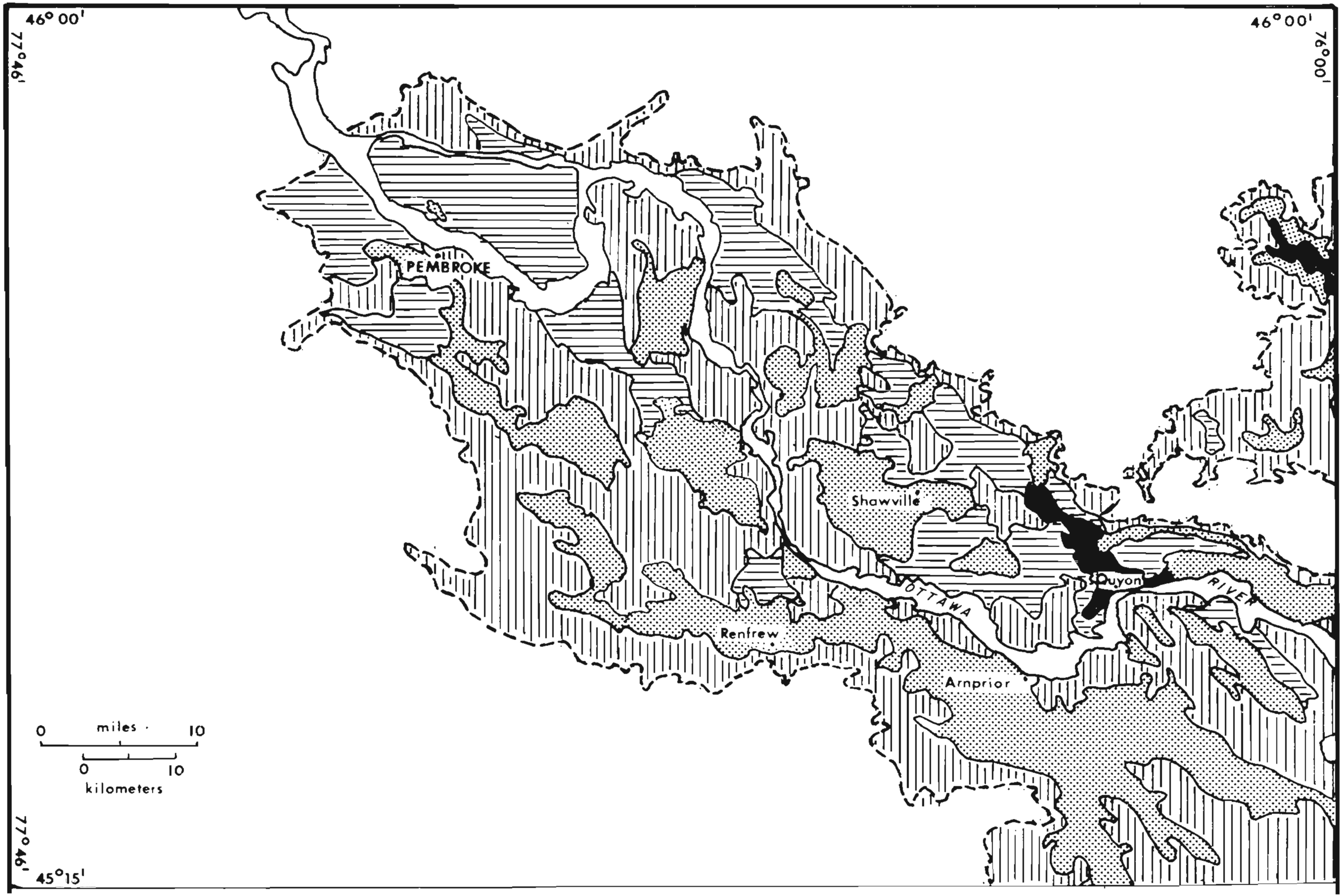

Figure 9. 1a. Distribution of sensitive clay and associated landslides Pembroke — Ottawa.

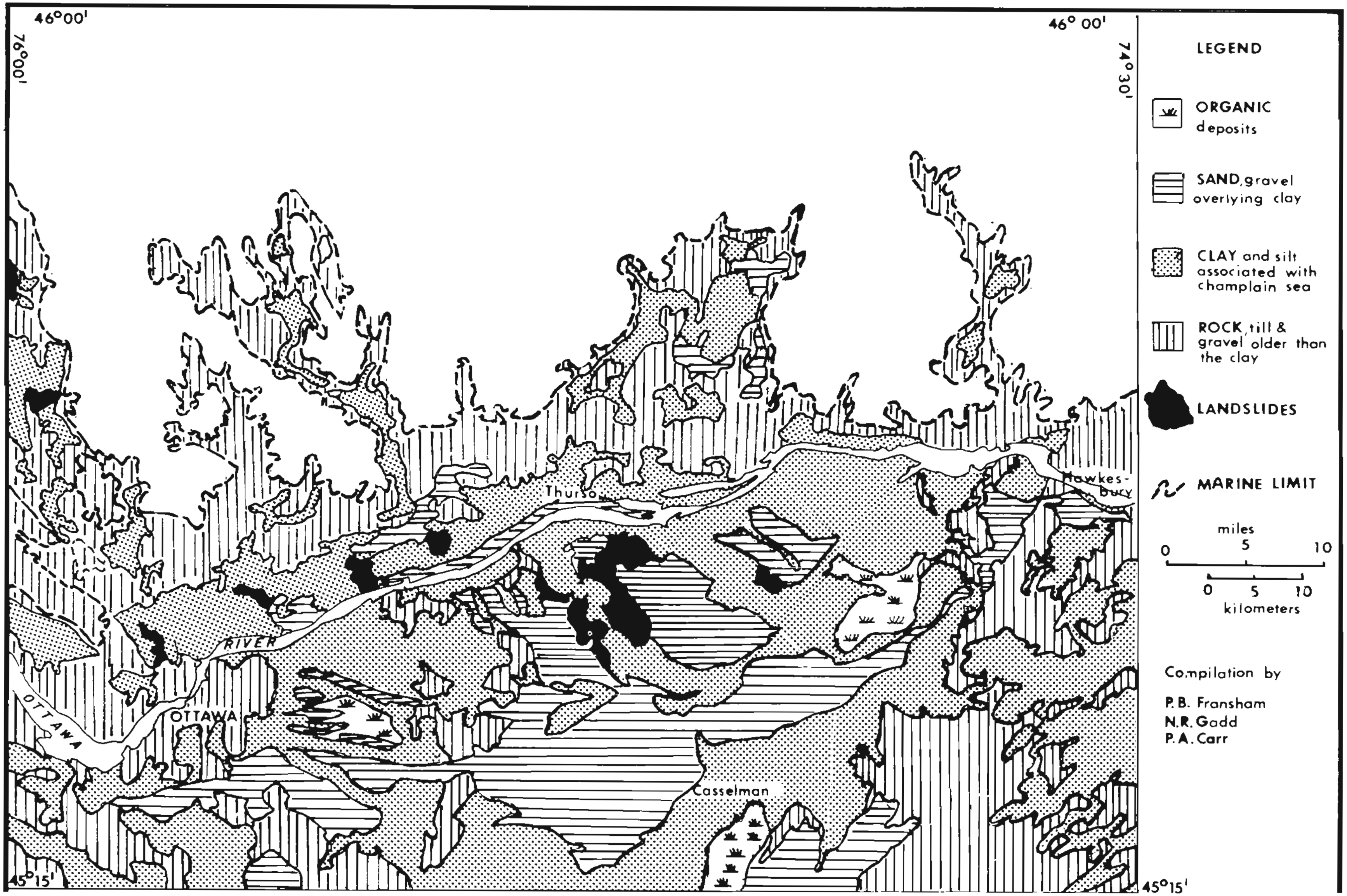

Figure 9. 1b. Distribution of sensitive clay and associated landslides Ottawa — Hawkesbury.

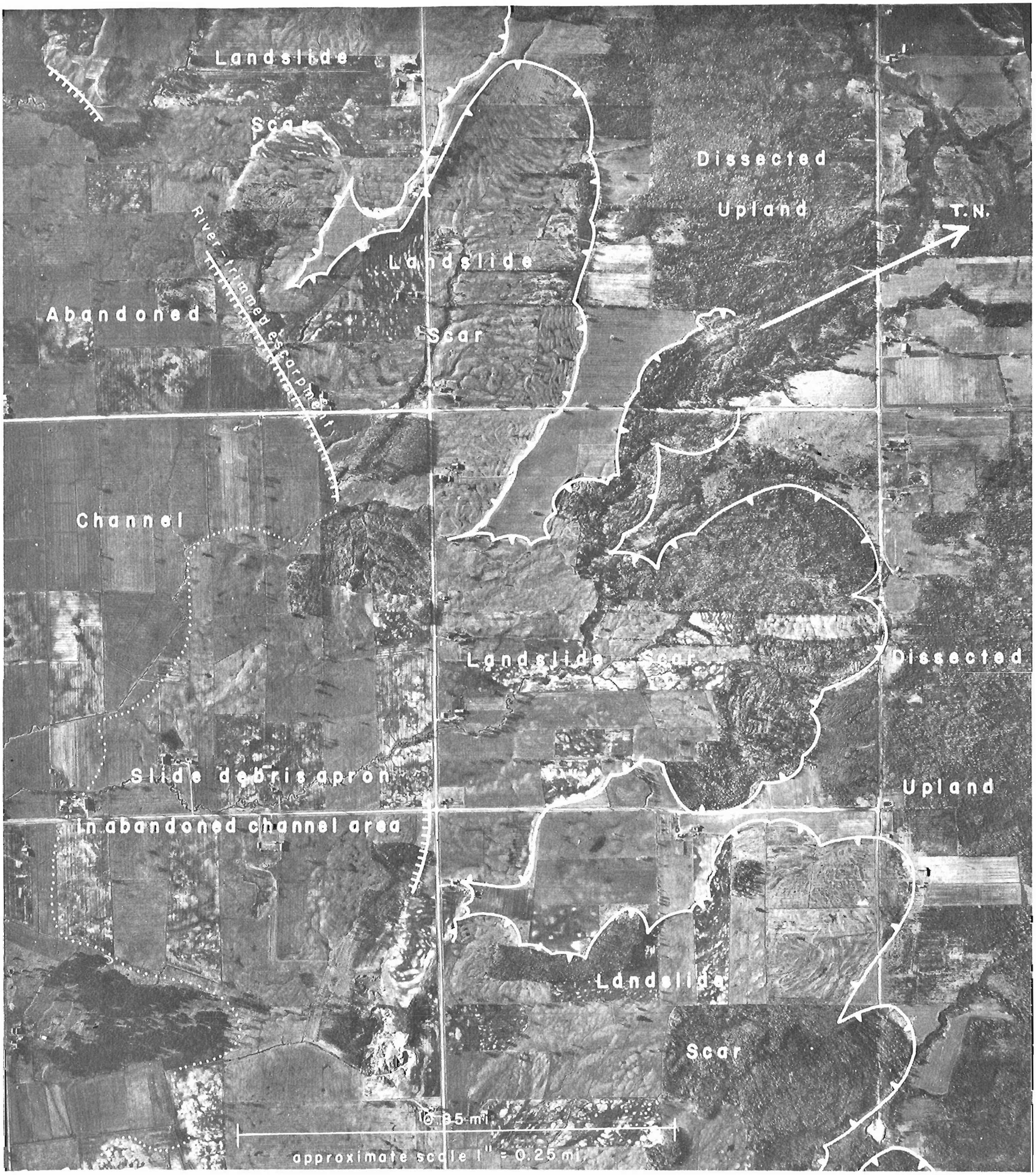

Figure 9. 2. Landslide scars adjacent to an abandoned river channel five miles south of Thurso.

There is good reason to believe that in certain areas a clear distinction between the marine and freshwater facies of silt and clay deposits may be made (Gadd, in press). Marine units of typical fossiliferous, blue-grey, silty clay may be overlain in places by non-calcareous brownish-grey, silty clay. This upper unit in many places is of freshwater origin rather than a "weathered or desiccated crust" as it is sometimes identified. For the purposes of this compilation, because of the great similarity of physical properties of the sediments, no attempt has been made to show these deposits separately on the map.

The thickness of the fine grained deposits is highly variable, depending in part on the bedrock topography. Two deep sections are known to exist. The first is near Arnprior where nearly 91.5 m of clay were encountered during the site evaluation for an earthfill dam (Acres, 1975). The second was at Treadwell, Ontario (5 miles southeast of Thurso), where drilling encountered 104.7 m of clay. The drill log indicated clay with sand lenses to a depth of 39.6 m and black mottled clay from 39.6 to 104.7 m. Indications are that thick deposits of clay are likely to be encountered in some of the abandoned postglacial river channels and along sections of the present Ottawa River.

Sand and gravel overlying the fine grained deposits. Sand and lesser amounts of gravel overlie large areas of the fine grained deposits. The sand comprises fluvial and deltaic deposits associated with the proto-Ottawa River. The sand is remarkably uniform, fine, and has a grey to buff colour. Only the sand and gravel known or suspected to overlie the fine grained deposits are illustrated in Figure 9.1. Similar materials that are laterally contiguous with the clay and are situated around the margins of the basin are not shown in Figure 9.1 for they are not indicative of engineering problems similar to those encountered when the sand is overlain by the fine grained deposits.

Organic deposits. Large organic deposits of peat and muck that overlie the fine grained deposits also are illustrated in Figure 9.1. They occur in the abandoned channels of the proto-Ottawa River at Moose Creek, Alfred, Mer Bleue, and elsewhere. The thickness of these deposits is unknown and is likely to be variable, but they must be considered as areas where engineering problems may occur.

Slope failures. The identification and plotting of landslides and their compilation at the scale of 1:50 000 was preparatory work for the study. It gives a regional view of the distribution of slides, from which it is apparent that most slides occur in the central part of the basin and along some of the northern tributary valleys leading into the Ottawa River. Only a few slides have occurred northwest of Shawville, Quebec and Renfrew, Ontario.

The two most extensive areas of landsliding have developed in the Quyon area of Quebec and near Rockland, Ontario (5 miles south of Thurso) (Figs. 9.1a and 9.1b). Both of these slide areas have developed along active or abandoned channels where extensive sand plains overlie the clay. It is assumed that the sand acts as a groundwater reservoir, thus maintaining high pore pressures in the underlying clay.

An inventory of landslides in part of the Ottawa Valley by Hanley (written comm., 1972) has shown that landslides have occurred in three, well defined geomorphic settings:

1. terrace bluffs and abandoned channels of the proto-Ottawa River;
2. banks of actively eroding streams which are tributaries of the Ottawa and Gatineau rivers;
3. high and steep banks of the present day Ottawa and Gatineau rivers.

Little is known about the age of landslide activity in the Ottawa Valley; however, in some cases relative ages can be determined. An aerial photo (Fig. 9.2) illustrates a series of landslides that occurred along the banks of an abandoned channel of the proto-Ottawa River situated south and east of Rockland. The northernmost slides are truncated by a river-trimmed escarpment and the aprons of slide debris have been removed by river erosion, whereas the landslides to the south have their debris aprons preserved. Thus the latter must have occurred sometime after the channel was abandoned and are younger than the northern slides.

References

Acres Ltd.
1975: Engineering problems in marine clay; Geotech. Seminar held at Arnprior Generating Station, June 13 and 14.

Gadd, N.R.
Geology of Leda Clay; Proc. 4th Guelph Symposium on Geomorphology, Guelph, Ontario, April 23-25 (in press).

1973: Distribution of marine deposits, Ottawa-St. Lawrence Basin; *in* Report of Activities, Part A, Geol. Surv. Can., Paper 73-1A, p. 183.

Logan, W.E.
1863: Geology of Canada, Report of Progress; Geol. Surv. Can., Dawson Bros., Montreal, p. 895-896.

Richard, S.H.
1974: Surficial geology mapping: Ottawa-Hull area (Parts of 31 F, G); *in* Report of Activities, Part B, Geol. Surv. Can., Paper 74-1B, p. 218-219.

10. TAKLA PROJECT (Reports 10-16): McCONNELL CREEK MAP-AREA (94 D, EAST HALF) BRITISH COLUMBIA

Project 750016

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Introduction

McConnell Creek map-area was remapped during the 1975 field season in conjunction with detailed studies of specific geological units in the map-area and in adjacent regions. Responsibilities for the various studies, grouped informally as 'The Takla Project', were as follows:

1. Ingenika Group — C.J. Dodds and J.L. Mansy;
2. Upper Paleozoic and part of Lower Mesozoic assemblages — J.W.H. Monger;
3. Granitic rocks — G.J. Woodsworth;
4. Ultramafic rocks — T.N. Irvine;
5. Mesozoic biostratigraphy — H.W. Tipper;
6. Middle and upper Jurassic sedimentology — O.L. Jeletzky.

Reports of the individual studies are contained elsewhere in this report.

The Takla Project was designed to focus on the plutonic, volcanic, sedimentary and structural history of the late Triassic and early Jurassic in view of numerous copper showings that had been discovered in rocks of these ages, not only in north-central British Columbia but elsewhere in the Cordillera. Future projects will continue these studies to the northwest in northwestern British Columbia and southwestern Yukon Territory.

Earlier work by Lord (1948) provided an excellent background for the current investigations. Recent published and unpublished data by N.B. Church (1974) was invaluable as a basis for detailed and regional correlation. In addition, maps provided by Falconbridge Nickel Mines Limited and British Petroleum Minerals Limited greatly aided specific studies.

Excellent assistance in the field was provided by R.G. Anderson, G.B. Mitchell, J. Dudley, T.E.T. Eadie, P.J. Proudlock, C.F. Roots, S.J. Hills, L.J. Werner and R.J. Wahlgren. Outstanding support was given by pilots Toni Okusaku of Transwest Helicopters Limited and R. Diston of Northern Mountain Helicopters Limited and by cooks J.E. Green and J.M. Wheeler.

Proterozoic

Ingenika Group (map-unit 1)

The Ingenika Group consists of regionally metamorphosed quartzofeldspathic gritty sandstone, siltstone, shale and minor conglomerate and limestone with an aggregate thickness exceeding 2000 metres (Mansy and Dodds, 1976). The most distinctive lithology is opalescent blue-eyed-quartz grit, locally conglomeratic. The rocks range in metamorphic grade from chlorite to at least kyanite. The Ingenika Group is in fault contact with the younger Paleozoic Lay Range Assemblage to the southwest.

Upper Paleozoic

Cache Creek Group (map-unit 2)

The Cache Creek Group consists mainly of dark grey quartzose phyllite (in places to biotite grade), foliated metachert, crystalline carbonate, amphibolite and associated gabbro, greenstone, serpentinite and serpentinized harzburgite (map-unit A). Carbonate of this unit, 20 km south-southeast of the map-area contain mid-Permian fossils (Paterson, 1974).

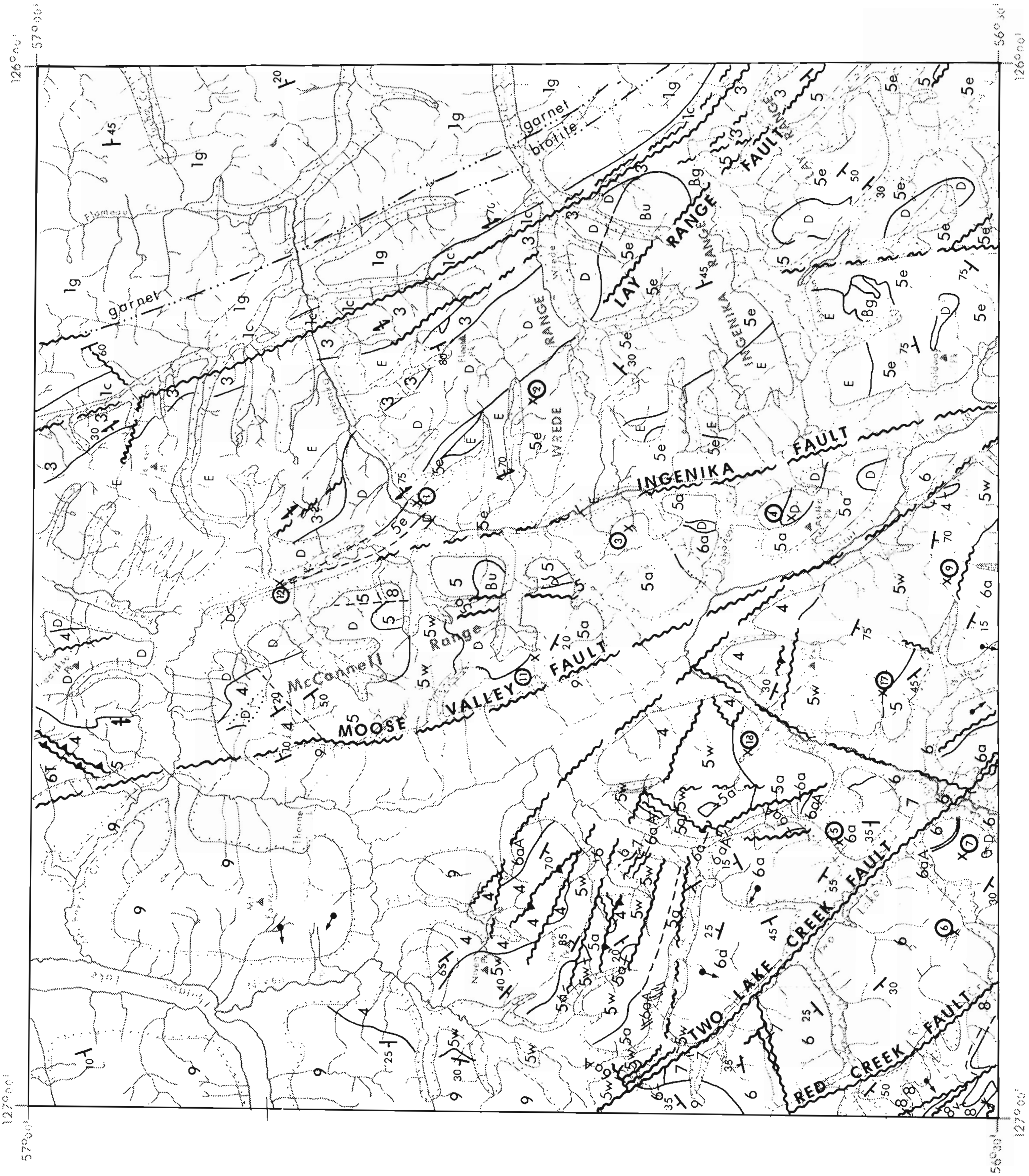
Lay Range Assemblage (map-unit 3)

The Lay Range Assemblage consists of a lower division of phyllite, phyllitic quartzite, chlorite schist, chert and foliated carbonate; and an upper division of thin bedded, green lithic tuff, volcanic breccia and minor pillow basalt. One-half mile east of the map-area, a carbonate in the lower part contains middle Pennsylvanian fossils. Rocks correlative with this assemblage in the southeastern Aiken Lake map-area contain ?Permian, possibly Lower Permian, fossils. It is probable that some of the rocks included with the Lay Range Assemblage are part of the Upper Triassic Takla Group, as the lithology of the two is very similar.

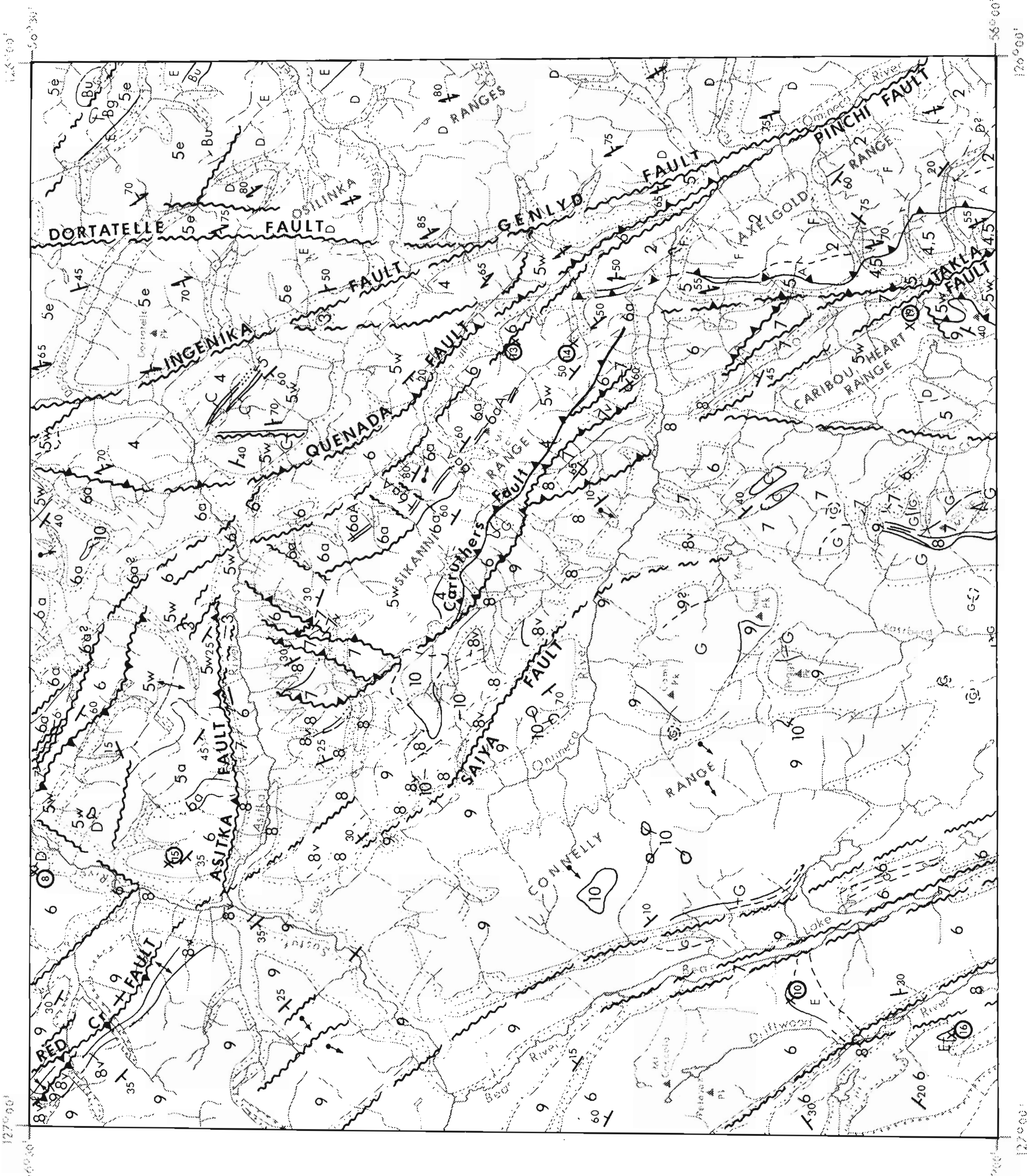
Asitka Group (map-unit 4)

The Asitka Group is best exposed near Dewar Peak area where it is divisible into three parts. A lower part comprises basalt, chert, argillite and a tuffaceous carbonate containing Lower Permian fauna. A central part consists of subaerial deposited basalt, grading upwards into rhyolite. Basalt, basalt breccia and tuff, chert and a tuffaceous carbonate with Lower Permian fauna form the upper part. Elsewhere the subdivision cannot be made as exposures are inadequate, and in the southern part of the map-area, where the rocks have been tectonically transposed into schist, phyllite and limestone pods.

The stratigraphic contacts of these Paleozoic units are either tectonic or disconformable. On the southwest side of the Sikanni Range, the Asitka Group lies in the hanging wall of an east-dipping (20-60 degrees) thrust over the Jurassic Hazelton Group. To the south, the Cache Creek Group, on a base of serpentinite,



garnet
biotite
LAY RANGE FAULT
LAY RANGE
WREDE RANGE
INGENIKA RANGE
INGENIKA FAULT
McConnell Range
MOOSE VALLEY FAULT
TWO LAKE CREEK FAULT
RED CREEK FAULT


DORTATELLE FAULT
INGENIKA
QUENADA
FAULT
GENLYD
FAULT
PINCHI FAULT
AXELGOLD
RANGE
TAKLA FAULT
CARIBOU HEART RANGE
SIKANNI
RANGE
Carruthers Fault
SAIYA FAULT
ASITKA FAULT
RED CR FAULT
CONNELLY
RANGE
OSILINKA
D RANGES
Omineca River
Lake
Driftwood
River
Bear River

STRATIGRAPHIC ROCKS

Tertiary or Quaternary

10 basalt, flows, necks, dykes, cones

Upper Cretaceous to Mid-Tertiary

9 Sustut Group; sandstone, siltstone, conglomerate, mudstone, minor coals, tuffs.

Upper Middle to Lower Upper Jurassic

8 Bowser Lake Assemblage; argillite, siltstone, sandstone, conglomerate; 8v: volcanics.

Lower Lower to Lower Middle Jurassic

Hazelton Group

7 argillite, greywacke, siltstone, sandstone, tuff, tuffaceous sediments, minor limestone.
7V: volcanics, breccia, tuff, pillow basalt.

6 red volcanics, basalt to rhyolite flow, breccia, tuff and associated intravolcanic clastics.
6T: Toodoggone volcanics.

6a: interbedded red and green volcanic conglomerate, sandstone, tuff, breccia, lahar, minor flows and pyroclastics. 6aA: polymictic conglomerate containing clasts derived from the Asitka Group mainly.

Upper Triassic

Takla Group

5 5e: augite porphyry flow, breccia, tuff and intrusion, black shale, bedded limestone, felsic tuff and breccia.

5w: augite porphyry flows, breccia, pillow basalt, tuffs and tuffaceous sediments, black shale and siltstone, minor limestone.

5a: subareally erupted flow, breccia, tuff, sandstone, mudstone and lahar conglomerate.

Permian

4 Asitka Group: basalt, chert, rhyolite, tuffaceous limestone, limestone, argillite.

Permian and Pennsylvanian

3 Lay Range Assemblage: tuff, breccia, pillow basalt, quartzite, phyllite, limestone.

2 Cache Creek Group: phyllite, metachert, greenstone, amphibolite, marble.

Proterozoic

1 Ingenika Group: lg: grit unit; chlorite-biotite-muscovite-garnet schist. lc: Cunningham limestone, marble, limy phyllite.

Selected Mineral Deposits

1	Nikos	Cu
2	Red	Cu
3	D.S.	Cu
4	Asitka	Cu
5	Group	Cu, Ag
6	Birch	Cu
7	Roy	Cu
8	Day	Cu
9	S.L.	Cu
10	Bear	Cu, Mo
11	Marmot	Cu
12	D W C	Cu
13	Pad	Cu
14	Arp	Cu
15	Pat	Cu, Zn
16	Motase	Cu, Mo
17	Willow	Cu
18	Sustut	Cu
19	Northstar	Cu

INTRUSIVE ROCKS

Mid Tertiary

G — Kastberg Intrusions: rhyolite, rhyodacite plugs, sills, dykes.

Mid Cretaceous

F — Axelgold Layered Gabbro Intrusion

Mid Cretaceous or Older

E — Satellitic granodiorite-quartz monzonite stocks to Hogem Batholith, Jensen Peak Batholith, Johanson Creek stock, Tsaytut spur stock (may be younger?)

Jurassic or Older

D — Hogem Batholith, Fleet Peak pluton; Asitka Peak, Darb Lake, Johanson Lake, McConnell Range and Fredrikson Peak stocks.

C — diabase

B — Bu: ultramafics, dunite, peridotite, pyroxenite;
Bg: gabbro, hornblendite

Upper Paleozoic

A — Serpentinite, serpentinized peridotite and dunite.

MAP SYMBOLS

bedding

overturned bedding

Foliation, schistocity

Mineral Deposit X (12)

isograd

approximate limit of exposure

Fault - Known, inferred, speculative

high angle reverse fault, inferred, speculative

Thrust Fault, inferred, speculative

geologic contact known, inferred speculative

Fold; syncline, anticline

average paleocurrent direction from crossbeds (≥ 7 measurements)

had been thrust to the west over a metamorphic assemblage that may contain elements of both the Asitka Group and lower Mesozoic strata (Paterson, 1974). West of Dewar Peak and Sustut Peak, the Asitka Group is overlain disconformably by Upper Triassic rocks, but commonly the upper contact is tectonic. All external contacts of the Lay Range Assemblage are tectonic.

Mesozoic

Upper Triassic to Jurassic rocks included by Lord (1948) in the Takla Group have been subdivided into three main units. The Takla Group (unit 5) has been restricted to Upper Triassic rocks typified by a predominance of basalt with characteristic augite phenocrysts. The Lower and early Middle Jurassic Hazelton Group (units 6 and 7) is represented by a lower unit of mainly nonmarine calc-alkaline volcanics and volcaniclastics and an upper marine sedimentary-volcanic assemblage. Stratigraphically and structurally between the Takla and Hazelton groups lies a north-westerly trending belt of volcanic clastics and epiclastic sediments (units 5a and 6a) characterized by reddish lahar, fanglomerate, conglomerate, sandstone, and minor contemporaneous volcanics. The lower part of this assemblage (map-unit 5a), of Late Triassic age, is conformable with the Takla Group and has been included with this unit, which includes Falconbridge's Sustut Copper property. Its upper part (units 6a and 6aA) is transitional lithologically with the lower unit of the Hazelton Group, but its age is unknown. Rocks of the Middle and Upper Jurassic Bowser Lake Assemblage are dominantly a clastic assemblage with deltaic affinities representing a successor basin to the earlier Hazelton Group (Jeletzky, 1976).

Takla Group (map-unit 5)

The Takla Group comprises two distinctive facies assemblages separated by the Ingenika Fault. Both assemblages are restricted to the Upper Karnian and Lower Norian, possibly Upper Norian, stages of the Upper Triassic.

East of the Ingenika Fault, the Takla Group (map-unit 5e) consists of augite porphyry tuff, breccia, flow and some high-level, subvolcanic intrusive bodies. Interbedded are black shale siltstone, tuffaceous fine-grained clastics and well bedded, laminated limestone. Dacite and rhyolite tuff and breccia are conspicuous, particularly in the Ingenika Range and west of the Dortatelle Fault, where, with the clastic rocks, they are locally more abundant. This suite of the Takla rocks is regionally metamorphosed into the lower greenschist facies.

The base of the western belt of the Takla Group (map-unit 5w) is marked by more than 500 metres of well bedded argillite, siltstone and minor sandstone. Above the sediments are a wide variety of augite porphyry volcanics and augite-bearing, volcanic-derived sediments. Between Niven Peak and Sustut Peak the volcanics attain their greatest thickness and include flow, tuff, breccia and pillow basalt. To the south, much of the section is composed of augite-bearing sediments, tuff and breccia that may be a distal facies of the northern assemblage. Most of the assemblage is of marine depositional origin, and only in the more northern exposures, between Niven Peak and Sustut Peak and between the southern McConnell Range and Asitka Peak are reddish coloured, subaerial eruptive volcanics abundant. These reddish coloured volcanics include unit 5a and contain most of the copper prospects. Commonly found within this western belt of the Takla Group is a conspicuous, bladed feldspar porphyry unit that is absent in the eastern belt. The rocks are intensely faulted and range in metamorphic grade from unmetamorphosed, through zeolite to prehnite-pumpellyite facies.

Hazelton Group (map-units 6 and 7)

The Hazelton Group has been divided into two units, a lower, nonmarine volcanic and volcaniclastic unit (map-unit 6) and an upper marine volcanic and sedimentary unit (map-unit 7).

Map-unit 6 includes two assemblages, an eastern, and possibly lower, volcaniclastic unit with subordinate volcanics (units 6a and 6aA) and a much more extensive volcanic unit (map-unit 6). Map-units 6a and 6aA form a narrow, elongate, belt of mainly reddish coloured, immature volcaniclastic and epiclastic rocks that were probably deposited in a graben-like trough whose outline is roughly similar to the present limits of outcrops. A conglomeratic marker horizon (unit 6aA), composed of Asitka Group clasts and minor granitic cobbles up to 1 metre diameter, is traceable along the entire length of the belt. In the north, the conglomerate forms the base of unit 6a, overlying unit 5a of the Takla Group unconformably, whereas to the south, in the Sikanni Range, it appears stratigraphically well up in the Hazelton section. East of Willow Creek, the basal beds of unit 6a comprise conglomerate and sandstone composed mainly of Takla and subordinate Asitka derived clasts. Most of unit 6a is composed of andesitic to dacitic feldspar porphyry clasts whose most logical source is either the calc-alkaline volcanics of unit 6 or possibly the "Toodoggone" volcanics (map-unit 6T). Volcanics, including basalt flow and breccia, andesite and rhyolite in the upper part of the unit are subordinate to the volcaniclastics. The unit attains a thickness in excess of 1000 metres east of Willow creek and is thin and of minor importance east and west of the main belt. No precise age for this unit could be determined. It appears to rest disconformably on the Takla Group, but is sedimentologically similar to map-unit 5a of the Takla Group. Its correlation and inclusion within the Hazelton Group is based upon the presence of contemporaneous calc-alkaline volcanism in its upper parts and on the composition of the clasts. The unit is unique for the Mesozoic of north-central British Columbia and forms a base to the Hazelton Group only in this local area. Alteration varies greatly with selective to wholsale replacement and cementation by epidote or laumontite. Many beds remain unaltered.

Map-unit 6 comprises more than 1000 metres of red, nonmarine, basalt to rhyolite calc-alkaline volcanics. Andesite flows and pyroclastics predominate but basalt flows are locally important. Rhyolite flows and breccias are sporadic, but may form ignimbrite-like members up to 50 metres thick. Intravolcanic sedimentary rocks, somewhat similar to parts of unit 6a, form a minor but integral part of unit 6. These volcanics form a thick monotonous assemblage west of Two Lake Creek and extend west to Bear Lake and southward for several hundreds of kilometres where they are the dominant and typical lithology of the Hazelton Group.

Conformably above the nonmarine volcanics of map-unit 6 are the marine sedimentary-volcanic facies of map-unit 7. The transition from unit 6 to unit 7 is abrupt and represents a major facies change within the Hazelton Group (Tipper, 1976). The unit can be divided into four approximate time-rock stratigraphic assemblages. Its lower part, of Late Sinemurian to Early Pliensbachian age, comprises greywacke, argillite, limestone and minor reddish tuff. A Lower Pliensbachian to Lower Toarcian assemblage consists of well bedded greywacke siltstone, argillite, acidic tuff, limy argillite, and minor basaltic breccia. Between the Upper Pliensbachian and Lower Toarcian facies is a suite of basaltic volcanics (map-unit 7v) that are mainly breccias but include common pillow basalt. Black argillite, siltstone and minor tuff are common in the Middle and Upper Toarcian. Feldspathic greywacke, siltstone, argillite, ash fall tuff and tuffaceous sediments of latest Toarcian and Bajocian age are the uppermost lithologies of map-unit 7. The thickness of this unit is not regular, but is probably in excess of 500 metres, and possibly thins and disappears in the eastern part of the sheet.

Bowser Lake Assemblage (map-unit 8)

Rocks of the Bowser Lake Assemblage were deposited between Bathonian (mid-Middle Jurassic) and Oxfordian (early Upper Jurassic) times. They represent marine deltaic clastic deposition, interrupted by two brief periods of volcanic activity. The lower part of the sequence is a fine grained clastic assemblage that comprises well bedded silty argillite, concretionary argillite, siltstone, minor sandstone and volcanic-argillite pebble conglomerate. This facies, passes upward with a gradational contact into a coarser facies containing abundant sandstone and volcanic-chert pebble conglomerate of mid-Oxfordian age. Paleocurrent data indicates sediment transport to the west-southwest. The earliest volcanic episode (map-unit 8v) is represented by up to 100 metres of marine deposited, greenish, feldspar porphyry breccia of early Oxfordian age. The second period of volcanism occurs conformably above the upper, mid-Oxfordian assemblage. These are non-marine intermediate to basaltic feldspar and augite porphyry breccias that outcrop in a linear belt west of Red Creek and east of Saiya Creek. One member of the suite is an impressive mud-flow unit containing angular clasts of augite porphyry up to 5 metres in diameter and a remarkable amount of carbonized trees, some over 1 metre in diameter.

Mesozoic and Tertiary

Sustut Group (map-unit 9)

The Sustut Group, comprising alluvial clastics rocks with minor ash fall tuff, had been investigated by Eisbacher (1974). It was divided by him into a lower, Tango Creek Formation and an upper, Brothers Peak Formation. The base of the Sustut Group rests unconformably upon the Takla Group west of Niven Peak, on the Hazelton Group at the head of Two Lake Creek and on the Bowser Lake Assemblage west of Red Creek. The basal contact is consistently at about 1600 metres elevation and represents an erosional plane developed during pre-Sustut times. Many of the clasts in both formations of the Sustut Group have been derived from the Ingenika Group, representing, within the McConnell Creek east-half map-area, the first evidence of exposure of this rock unit. Paleocurrent data from cross-beds obtained during this work from both the Tango Creek and Brothers Peak formations gave a consistent southwest sediment transport direction.

Tertiary or Quaternary

Basalt (map-unit 10)

A few areas are underlain by remnant exposures of basalt, particularly adjacent the Saiya Creek Fault and in the Connelly Range. They are mainly flows, a few cinder cones, necks and numerous small dykes. Mainly they were deposited on an old, post-Sustut Group erosion surface at an elevation of about 1500 metres, and have been subsequently deeply dissected by glacial action. One flow, east of Saiya Creek, represents a valley fill basalt.

Intrusive Rocks

Intrusive Rocks range in age from Triassic to Eocene, and in composition from dunite to rhyolite. The greatest volume of intrusive rocks lie between the Pinchi-Ingenika faults and the area underlain by the Ingenika Group (Woodsworth, 1976).

Map-unit B: Triassic or Jurassic

Oldest intrusive units are represented by small bodies of dunite, peridotite, pyroxenite, their serpentinized equivalents (unit Bu) and gabbro and hornblendite (unit Bg). All but one of these bodies intrudes the eastern belt of the Takla Group (unit 5e) with which they are thought to be congenetic (Irvine, pers. comm.).

Map-unit C: Jurassic or older

Small bodies of diabase are exposed in the southern part of the map-area.

Map-unit D: Triassic or Jurassic

Intrusive bodies of Jurassic, or possibly Triassic age, constitute the most voluminous episode of plutonism. Largest body is represented by the Hogem Batholith in the southeast corner of the map-area. Numerous stocks (Asitka Peak, Johanson Lake, Darb Lake, McConnell Range and Fredrikson Peak) north of the batholith are possibly coeval. An elongate batholithic body, the Fleet Peak Pluton, occurs east of the Lay Range Fault. Three small plugs near the Sustut River constitute the only Jurassic or older correlative intrusions within the western Takla belt (unit 5w). Both the Hogem Batholith and Fleet Peak Pluton contain a well developed cataclastic foliation, particularly along their western margins.

Map-units E and F: Cretaceous and older(?)

A series of generally north-northwest elongate stocks and batholiths of probable Cretaceous age intrude the eastern Takla belt and the Lay Range Assemblage. These are mainly unsheared, even-grained quartz monzonite and granodiorite. The Axelgold layered gabbro intrusion (unit F) is of mid-Cretaceous age, but does not appear genetically related to these eastern bodies. A small stock west of Bear Lake is included with the Cretaceous plutons, but its age is unknown and it may be correlative with the Bulkley Intrusions (Upper Cretaceous) or the Babine Intrusions (Eocene).

Map-unit G: Eocene

The Kastberg Intrusion represent a series of stocks, dykes and sills of Eocene age (K/Ag; 48 and 45 my) between Scallop Mountain and Bear Lake. They are high-level leucocratic bodies of dominantly rhyolitic composition. They intrude rocks as young as Lower Sustut.

Structure

Faults dominate the structural style and folds occur mainly in incompetent sediments in response to faulting. Faulting has occurred continuously or in many stages from late Paleozoic to post-Eocene times.

Earliest evidence of large scale faulting is present in the juxtaposition of three markedly differing Upper Paleozoic assemblages (units 2, 3 and 4) and two upper Triassic assemblages (units 5e and 5w). The locus of this faulting was probably the Takla Pinchi-Ingenika-Dortatella-Lay Range fault systems. Movement in earliest lower Jurassic is evident by the cataclastic foliation along the west margin of the older phases of the Hogem Batholith, but not found in its younger phases (Woodsworth, 1976). The earliest recognizable stage of faulting in the Hazelton Group (units 6 and 7) was associated with the development of an elongate, graben-like structure in which was deposited over 1000 metres of fanglomerate and conglomerate (map-unit 6a). Clasts in the lower members of this unit were derived from the rapidly uplifted, adjacent, Asitka and Takla groups and from nearby intrusions. Regional faulting occurred prior to the mid- to early Late Cretaceous because the basalt beds of the Sustut Group (unit 9) were deposited on an erosional surface that juxtaposes the Asitka, Takla and Hazelton groups and the Bowser Lake Assemblage. Throw on the faults is in the order of 500 metres. The maximum fault movement was probably mid-Oxfordian coinciding with the upward coarsening of the deltaic facies of the Bowser Lake Assemblage (unit 8) and the first appearance of clastic chert (off the Cache Creek Group?). Regional uplift took place in mid- to early Late Cretaceous with the exposure of the Ingenika terrane (unit 1) in the east and deposition of the Sustut Group to the west. In Eocene times a myriad of small scale, imbricate thrust faults directed to the northeast involved the Sustut Group and the Bowser Lake Assemblage (Eisbacher, 1974). These involve the Brothers Peak Formation, and are, hence, Eocene or younger. They appear dominant north of the Sustut River, and die out rapidly to the south.

References

Church, N. B.
1974: Geology of the Sustut Area; Geology, Exploration and Mining in British Columbia in 1973; B. C. Dept. Mines Pet. Resour., p. 411-455.

Eisbacher, G. H.
1974: Sedimentary History and Tectonic Evolution of Sustut and Sifton Basins, north-central British Columbia; Geol. Surv. Can., Paper 73-31.

Irvine, T. N.
1975: Axelgold Layered Gabbro Intrusion, McConnell Creek map-area, British Columbia; *in* Report of Activities, Part B, Geol. Surv. Can., Paper 75-1B, p. 81.

Jeletzky, O. L.
1976: Takla Project: Preliminary report on stratigraphy and depositional history of Middle to Upper Jurassic strata in McConnell Creek map-area (west half) British Columbia; *in* Report of Activities, Part A, Geol. Surv. Can., Paper 76-1A, rept. 13.

Lord, C. S.
1948: McConnell Creek map-area, Cassiar District, British Columbia; Geol. Surv. Can., Mem. 251.

Mansy, J. L. and Dodds, C. J.
1976: Operation Finlay and Takla Project: Stratigraphy, structure and metamorphism in northern and central Swannell Ranges; *in* Report of Activities, Part A, Geol. Surv. Can., Paper 76-1A, rept. 18.

Paterson, I. A.
1974: Geology of Cache Creek Group and Mesozoic rocks at the northern end of Stuart Lake Belt, Central British Columbia; Geol. Surv. Can., Paper 74-1, pt. B, p. 31-42.

Tipper, H. W.
1976: Biostratigraphic study of Mesozoic rocks in intermontane and insular belts of the Canadian Cordillera; *in* Report of Activities, Part A, Geol. Surv. Can., Paper 76-1A, rept. 12.

Woodsworth, G. L.
1976: Plutonic rocks of McConnell Creek and Aiken Lake map-area; *in* Report of Activities, Part A, Geol. Surv. Can., Paper 76-1A, rept. 14.

LOWER MESOZOIC ROCKS IN McCONNELL CREEK MAP-AREA (94 E), BRITISH COLUMBIA

Project 750015

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Rocks of the Upper Triassic (?) to Upper Jurassic Takla Group (as used by Lord, 1948) in McConnell Creek map-area are host to numerous copper showings, some of which appear to be stratigraphically controlled (Church, 1974; Harper, 1975). The present study was undertaken to determine the regional stratigraphy and environments of deposition of the lower Mesozoic rocks and their relationships to copper mineralization. The findings have resulted in restriction of the term Takla Group to the Upper Triassic (upper Karnian to mid-Norian), and the inclusion of the younger rocks in the Jurassic Hazelton Group, in partial agreement with Church (1974). Fossiliferous Upper Triassic and Lower Jurassic rocks are stratigraphically separated by a thick assemblage of non-fossiliferous, mainly subaerial sedimentary and volcanic rocks that contains numerous copper showings. The lower part of this assemblage, found only locally, is gradational downwards into fossiliferous Triassic rocks and is thus part of the Takla Group. The more extensive upper part, commonly separated from known Triassic rock by an abrupt lithologic change, is gradationally and conformably overlain by Sinemurian (Lower Jurassic) rocks (Tipper, 1976), and is considered to be the basal part of the Hazelton Group in McConnell Creek map-area.

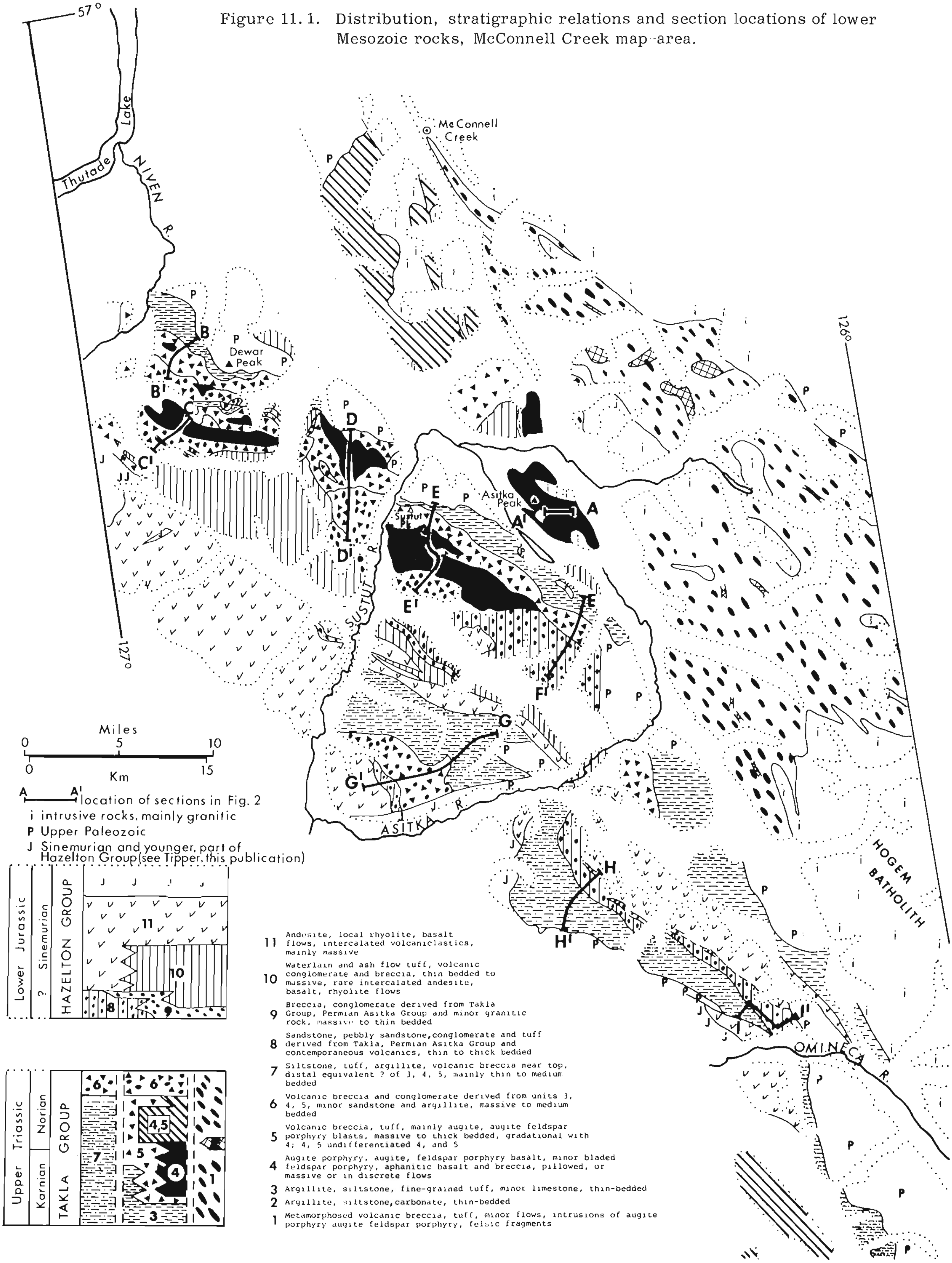
Takla Group (restricted)

The easternmost rocks of the Takla Group (units 1, 2, Fig. 11.1) lie west of a narrow belt of Permo-Pennsylvanian strata and east of the Ingenika and Pinchi faults, that run from McConnell Creek in the north to Omineca River in the south (Richards, 1976). The rocks are metamorphosed to greenschist facies and locally higher grade, and in places are highly deformed and foliated. This belt of Takla rocks has been intruded by numerous granitic bodies, of which the largest is the Hogem Batholith (Woodsworth, 1976). The rocks (unit 1) are mainly altered tuff and breccia, with augite and augite feldspar porphyry and local felsic clasts, minor flow rocks and possible high-level intrusions characterized by large (⩾1 cm in length) hornblende or pyroxene phenocrysts that may comprise up to 35 per cent of the rock. Beds of siltstone, argillite and carbonate (unit 2) within this sequence contain upper Karnian and lower Norian fossils, but neither top nor base of this section has been recognized. Mineralization occurs mainly near granitic intrusions.

Triassic volcanic and sedimentary strata (units 3 to 6) west of the Pinchi-Ingenika faults, between Niven River in the northwest and the upper part of Asitka River, range from essentially unmetamorphosed rock to prehnite-pumpellyite grade. Most deformation is by complex faulting although the lowest, incompetent unit 3 is locally highly folded. The basal part of the Takla Group (unit 3) is composed of thin bedded, graded, dark grey to brown argillite, siltstone, tuff and local thin carbonate lenses. The unit reaches a maximum thickness of about 300 metres (1000 feet) west of Dewar Peak, where it lies apparently disconformably on the Lower Permian Asitka Group (Section B-B', Fig. 11.2). Unit 3 is overlain by, and partly laterally equivalent to, a thick succession of interfingering extrusive and pyroclastic rocks (units 4, 5). The extrusive rocks (unit 4) are commonly pillowed but some are massive, and others, towards the top of the succession, form individual flows with local brecciated flow tops that are particularly well displayed on Asitka Peak (Fig. 11.2). The maximum thickness of these rocks approaches 3000 metres (9000 feet) near Sustut Peak (Section E-E', Fig. 11.2), but elsewhere the thickness varies dramatically, from 800 metres (2500+ feet) to less than 300 metres (1000 feet) in adjacent fault slices near Dewar Peak (Section B-B', C-C', Fig. 11.2). Compositions range from predominant dark grey, red-grey or locally grey-green augite porphyry and augite feldspar porphyry basalt to local aphanitic basalt or coarse-grained, 'bladed' feldspar porphyry trachy-basalt. The dark grey to green volcaniclastic sequence (unit 5), up to 1600 metres (5000 feet) in thickness, varies from massive coarse grained breccia that in places grades into the flow rock, through thick and medium bedded, graded, augite feldspar crystal lithic tuff to local thin-bedded, fine grained tuff and argillite. Breccia, conglomerate and finer grained rock of unit 6, up to 1300 metres (4000 feet) thick, gradationally overlie units 4 and 5 northwest and southwest of Sustut Peak, but differ from the older, essentially monomict breccia of unit 5 in that they contain clasts derived from a variety of Takla volcanics, show little evidence of contemporaneous volcanism, and are thus believed to be epiclastic. Near Dewar Peak (Section C-C'), unit 6 has a varicoloured, predominantly green, massively bedded lower part and an upper, red, well-bedded part. The entire sequence, inferred from enclosed fossils, is marine (Monger, 1974). Four miles northwest of Sustut Peak (Section D-D') there is a gradation upwards from a lower green and red, massive breccia, with interbedded fine grained clastics containing marine fossils, to local presumably subaerial cross-bedded conglomerate and breccia with interlayered siltstone and argillite containing mud cracks. This subaerial unit contains the mineralized member at the Sustut Copper property of Falconbridge, described by Church (1974) and Harper (1975). The transition from marine to nonmarine (fossils vs. mudcracks) lies several hundred feet below the copper-bearing horizon.

Farther southwest, Triassic rocks (units 6 and 7) extend from the Sustut River in the northwest to south of the Omineca River. Lowermost (unit 7) are up to 1600 metres (5000 feet) of grey and brown, thin- to medium-bedded, graded, argillite and fine grained

Figure 11.1. Distribution, stratigraphic relations and section locations of lower Mesozoic rocks, McConnell Creek map-area.



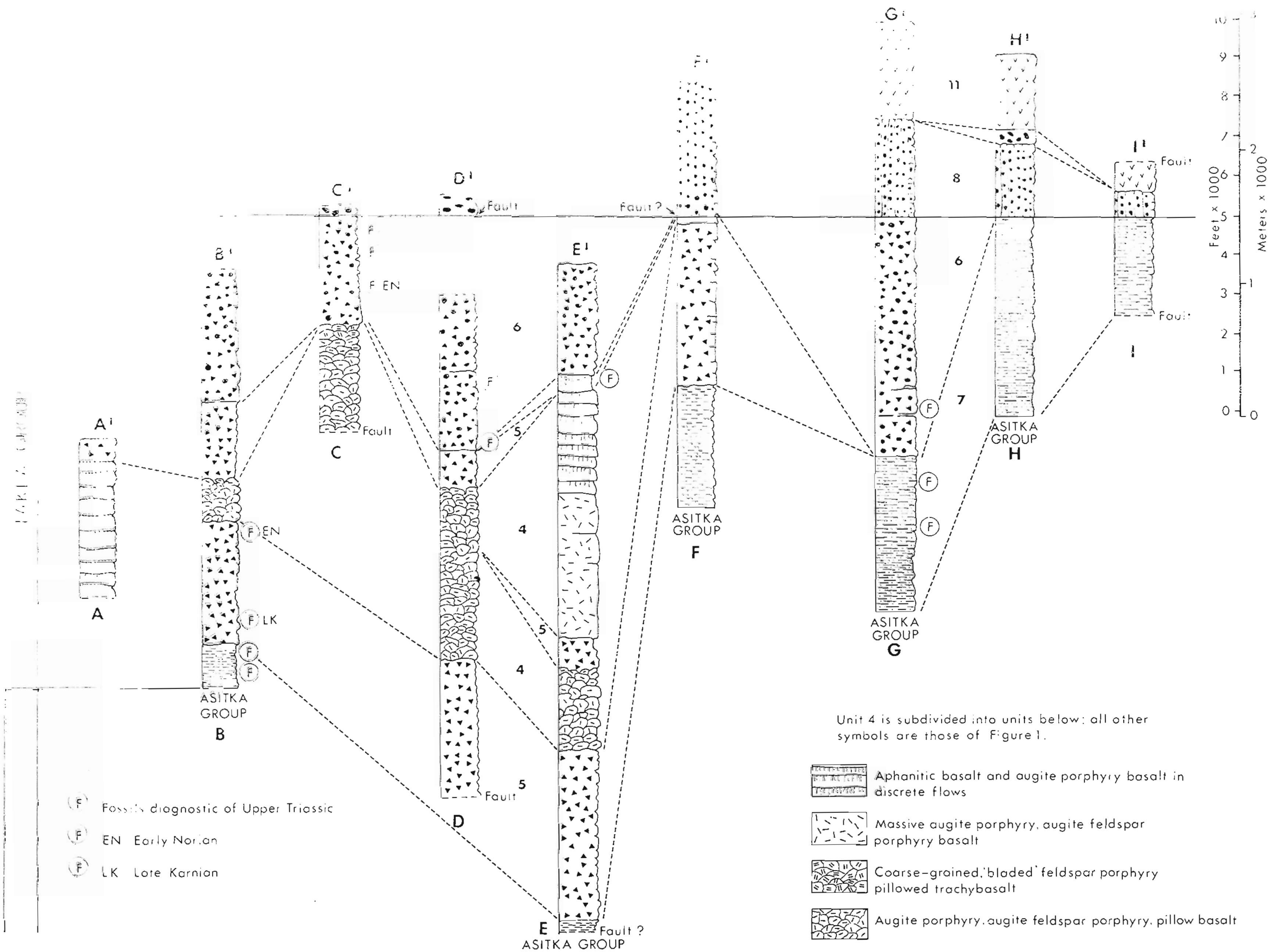


Figure 11.2. Respresentative stratigraphic sections of the Takla Group and overlying lower Hazelton Group. McConnell Creek map-area.

tuff with massive beds of volcanic breccia in the upper part. These rocks are probably the basinward, distal equivalents of units 3 and 5 to the east. North of Asitka River, this sequence is overlain gradationally by as much as 2000 metres (6000 feet) of red and green volcanic breccia and conglomerate correlated with Unit 6 to the northeast. Near Omineca River, these Triassic rocks are metamorphosed to phyllite and chlorite schist and probably form a major component of the Sitlika assemblage southeast of the map-area described by Paterson (1974).

Hazelton Group

The lower part of the Hazelton Group consists of red epiclastic breccia, conglomerate and sandstone (units 8 and 9) that grade upward into volcaniclastic rocks and intercalated flows (units 10 and 11). The predominant colour is red but, as in unit 6, many sandstone and conglomerate beds are zeolitized and grey, or epidotized and yellow-green, although the interbedded mudstones remain red. Southwest of Dewar Peak, polymict breccia (unit 8) lies in channels cut in Triassic rocks and contains not only volcanic rocks derived from the Takla Group but also clasts of chert, rhyolite and fossiliferous carbonate, derived from the Permian Asitka Group, and rare granitic clasts. Similar conglomerate, up to 500 feet thick can be traced nearly continuously south-southeastwards for at least 50 miles, and is interbedded with rocks containing probable Sinemurian fossils near Iktlaki Peak, 70 miles south-southeast, in northeastern Hazelton map-area (Richards, pers. comm.). South of Sustut Peak, the basal part of the Hazelton Group is a conglomerate consisting largely of Takla clasts that grades upwards into cross-bedded pebbly sandstone containing mixed Asitka and Takla lithologies (unit 9). In the Sikanni Range, cross-bedded, polymict sandstone, mudstone and minor tuff, locally with mudcracks, underlies the polymict conglomerate and sharply overlies graded beds of the Takla Group. An angular unconformity may separate the Hazelton from the Takla Group as the lower unit changes facies and thins markedly below the contact south of Sustut Peak and in the Sikanni Range. Above the polymict rocks the succession is dominated by volcanic rocks, mainly fine grained feldspar porphyry, andesitic pyroclastics and sedimentary rock derived from them. North of Sustut River, thin to massive bedded, cross-bedded conglomerate, sandstone with local mudcracks, breccia and tuff, some of which is welded (unit 10), pass upward and westward into andesite and local basalt and rhyolite flows and intercalated volcaniclastics (unit 11). South and southwest of Dewar Peak these rocks are overlain gradationally and unconformably by fossiliferous Sinemurian strata (Tipper, 1976). Farther south, the polymict clastics are overlain by tuff, breccia and intercalated flows of andesite and rhyolite with feldspar phenocrysts. Copper mineralization in this assemblage is generally associated with epidote and calcite in beds and fractures.

Lower Mesozoic Geological Evolution

The following applies to the low grade rocks west of the Ingenika and Pinchi faults:

Takla volcanism, represented initially by pillow lava (unit 4) and proximal breccia and tuff (unit 5), began in a marine shale basin (unit 3) in upper Karnian time. In places, the lava probably built above sea level as indicated by the discrete flow units with brecciated flow tops.

Near the end of the volcanism, in early Norian time, erosional debris from a variety of Takla volcanic rock formed local epiclastic breccia and conglomerate of unit 6. This detritus is in part the result of erosion of volcanic piles built above sea level, but there was probably tectonic uplift and erosion as well, for some clasts are derived from pillow basalts and marine sedimentary rocks. Deposition was wholly submarine in places but elsewhere marine rocks pass upward into rocks deposited subaerially.

During an interval of unknown duration, but possibly extending from middle Norian to Sinemurian time, there was regional uplift, possibly accompanied by block faulting, perhaps in response to or accompanied by intrusion of early phases of the Hogem Batholith (Woodsworth, 1976). Erosion of the Takla Group, the underlying Asitka Group and associated granitic rocks provided material that was deposited (units 8 and 9) in local channels and as sheets on the planated surface of the Takla Group. Current bedding directions (Richards, 1976) from this unit and the overlying units (10, 11) indicate that the source lay north and east of the map-area.

Following regional subsidence, sedimentation was marine (Tipper, 1976).

Copper mineralization in the subaerial rocks (units 6, 8, 9, 10) appears to have taken place in beds that were porous and permeable, as indicated by selective epidotization and zeolitization of certain members, and along fracture planes. Several clasts of Takla volcanics containing chalcopyrite and malachite/azurite in the polymict breccia (unit 8) indicate that at least some mineralization was pre-Hazelton. The source of the copper may have been Takla volcanics (units 4, 5). Preliminary analyses indicate that they have a higher copper content than equivalent rocks in either the subjacent Asitka Group or superjacent Hazelton Group.

References

Church, B.N.
1974: Geology of the Sustut area; Geology, Exploration and Mining in British Columbia in 1973; B.C. Dep. Mines Pet. Resour., p. 411-455.

Harper, G.
1975: Geology of the Sustut Copper Deposit, British Columbia; Can. Inst. Min. Bull., v. 68, no. 755, p. 52.

Lord, C.S.
1948: McConnell Creek map-area, Cassiar District, British Columbia; Geol. Surv. Can., Mem. 251.

Monger, J.W.H.
1974: The Takla Group near Dewar Peak, McConnell Creek map-area (94 D) British Columbia; *in* Report of Activities, Part B; Geol. Surv. Can., Paper 74-1B, p. 29-30.

Paterson, I.A.
1974: Geology of Cache Creek Group and Mesozoic rocks at the northern end of Stuart Lake Belt, central British Columbia; *in* Report of Activities, Part B; Geol. Surv. Can., Paper 74-1B, p. 31-42.

Richards, T.A.
1976: Takla Project: McConnell Creek map-area; *in* Report of Activities, Part A, Geol. Surv. Can., Paper 76-1A, rep. 10.

Tipper, H.W.
1976: Biostratigraphic study of Mesozoic rocks in intermontane and insular belts of the Canadian Cordillera: *in* Report of Activities, Part A, Geol. Surv. Can., Paper 76-1A, rep. 12

Woodsworth, G.L.
1976: Takla Project: Plutonic rocks of McConnell Creek and Aiken Lake map-areas, British Columbia: *in* Report of Activities, Part A, Geol. Surv. Can., Paper 76-1A, rep. 14.

12. BIOSTRATIGRAPHIC STUDY OF MESOZOIC ROCKS IN INTERMONTANE AND INSULAR BELTS OF THE CANADIAN CORDILLERA, BRITISH COLUMBIA

Project 750035

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Most of the field season was devoted to a study of Jurassic sedimentary sections in McConnell Creek map-area (94 D) in conjunction with the Takla Project (Richards, 1976; Monger, 1976). Brief investigations were also carried out in adjoining map-areas, i.e. Spatsizi (104 H), Toodoggone River (94 E), Fort Grahame (94 C), Manson River (93 N), and Hazelton (93 M). In addition, one week was spent in further study of the Maude and Yakoun formations on Queen Charlotte Islands.

Lower Jurassic Rocks of McConnell Creek (94 D) and Adjoining map-areas

The southern two-thirds of McConnell Creek map-area east-half has widespread exposures of Lower Jurassic volcanic rocks and volcanogenic sedimentary rocks. Although an Early Jurassic age has been clearly established for volcanic rocks of the Hazelton Group, only where fossiliferous sedimentary sections are in stratigraphic contact can the volcanics be related with confidence to stages or substages. The volcanics are many times more abundant than the sediments but they span only parts of one or two substages. It has not been demonstrated that Jurassic rocks older than Late Sinemurian outcrop in the map-area.

Several sections of sedimentary rocks were studied in the map-area as well as some in adjoining areas (Fig. 12.1). Most include volcanogenic sediments, thin beds of volcanic breccia, and waterlain ash-fall tuff bands. It is concluded that the sediments were deposited in a basin or basins within which (or nearby) volcanic activity was intermittent from Late Sinemurian to the end of the Early Jurassic. Late Sinemurian to Earliest Pliensbachian was the time of greatest volcanic activity – volumetrically and areally.

The sediments are dominantly siltstone with interbedded tuff and subordinate lithic sandstone. Conglomerate is rare and commonly occurs as coarse grit or as fine, well-rounded pebble conglomerate in beds a few inches thick. Limestone is minor except in Lower Pliensbachian rocks which include reef-like beds and limy siltstone replete with Weyla, other pelecypods, corals, gastropods, crinoids, and ammonites. The most complete and best exposed sections with the most abundant fauna are sections 1, 5, 6, 10 and 11 (Fig. 12.2); probably these sequences include the best fossil record of Pliensbachian and Toarcian stages in the western Canadian Cordillera.

The Late Sinemurian is documented in only two localities – sections 12 and 14 where beds with Arnioceras overlie Hazelton volcanics. In section 14 the sediments include a bed of conglomerate with well-rounded pebbles of quartzite to 4 inches in diameter; an easterly source area is suggested.

Several fossil zones of the Pliensbachian were recognized from a preliminary study of the faunas. In general the Lower Pliensbachian section thins northward and the Upper Pliensbachian thickens. Volcanics thin northward or end abruptly in the area of sections 10, 11 and 12. There is little evidence of volcanism after Sinemurian time.

Section 5 is the only well-exposed, complete Toarcian section. Early, Middle and Late Toarcian faunas abound in an essentially unfaulted, continuous sedimentary sequence. Ash-fall tuff decreases in importance upward in the sequence whereas glauconitic sands increase. Detritus is coarser than in the Pliensbachian succession and combined with generally more abundant pelecypods, suggests a shallower basin of deposition. Current structures are lacking. The rocks are planar-laminated, well-sorted, and lack conspicuous graded bedding.

Section 13 is mainly siltstone of Early Toarcian age without noticeable volcanic detritus. In several beds detrital muscovite is abundant. Conglomerate and sandstone with coal (Armstrong, 1949, p. 55), apparently beneath the siltstone includes pebbles of quartzite for which an easterly source is suggested. Section 13 apparently was deposited in an eastern extremity of the Toarcian basin, somewhat removed from active volcanism.

In contrast to other sections, sections 13 and 14 lie east of the Hogem Batholith and east of several major faults. None of the sections to the west clearly suggest an area of provenance or proximity to a shoreline whereas those to the east suggest an easterly source and proximity to a landmass. It seems reasonable to suggest that during some stages or substages in Early Jurassic time the Hazelton sea extended across the terrane now underlain by the Hogem Batholith although no Jurassic rocks are known in that belt. Presumably the belt was uplifted in post-Hazelton time with complete removal of Jurassic rocks.

Bajocian (early Middle Jurassic) Rocks of McConnell Creek (94 D) and Adjoining map-areas

Bajocian sediments overlie Toarcian beds conformably in section 5 (Fig. 12.2) and elsewhere are found in thin sequences in fault contact with older or younger strata. Similarity of Toarcian and Bajocian sediments and a gradational contact in one section suggests the rocks are conformable. Upper Bajocian faunas were not found. Middle Bajocian Humphresianum and Sauzei

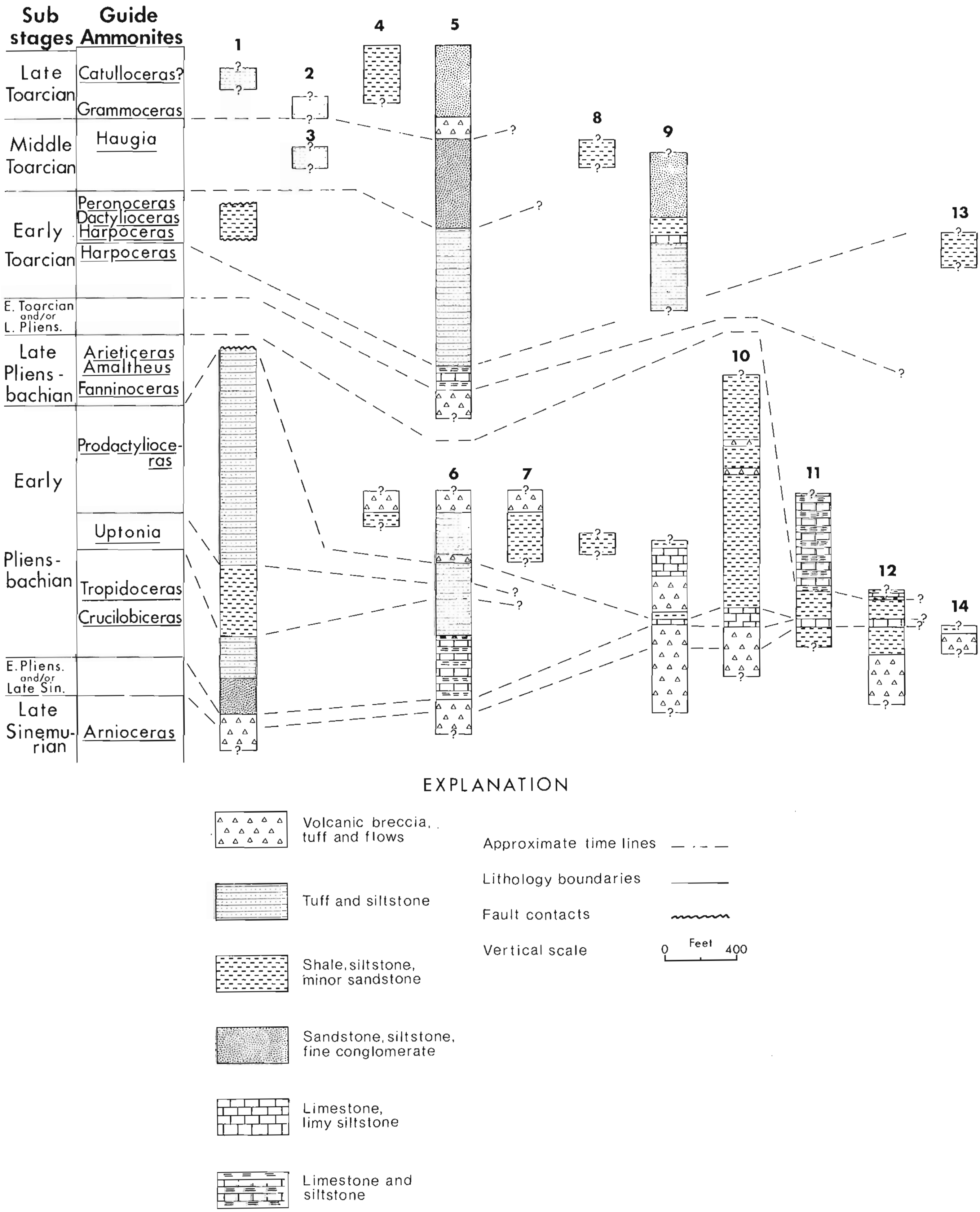

Figure 12.1. Lower Jurassic columnar section in McConnell Creek and adjoining map-areas.

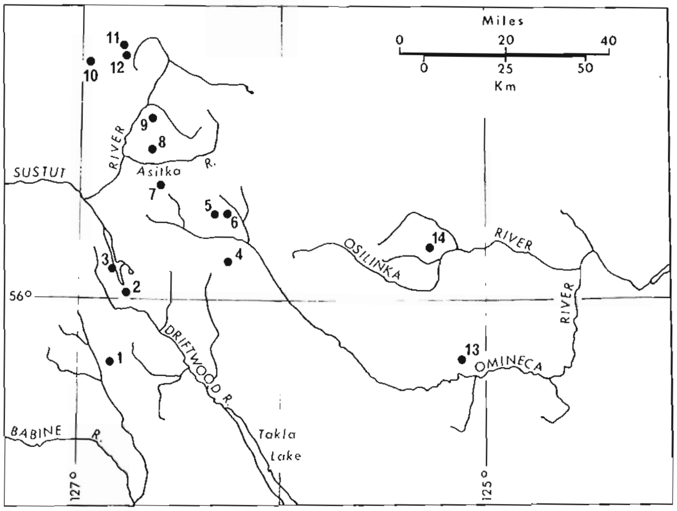

Figure 12.2. Index map of Lower Jurassic Columnar Sections.

ammonoids are commonly in the same section but no Sowerbyi zone fauna were identified. The Lower Bajocian ammonite, Tmetoceras, was obtained at locality 15 (Fig. 12.3). Although the Bajocian fauna is abundant and varied in the Smithers and Hazelton map-areas to the southwest, in the McConnell Creek area the fauna consists mainly of ammonites and is rarely abundant. Many seemingly unfossiliferous parts of the sequences may represent apparently missing zones and the Bajocian section may be in reality more complete than indicated.

In Figure 12.3, localities 15 to 20 have yielded Middle Bajocian faunas. At all of these localities sonniniids, presumably of the Sauzei zone are present and at localities 15, 16, 18, 19 and 20 stephanoceratids of the Humphresianum zone are stratigraphically higher. The enclosing rocks are typically hard siltstone and fine sandstone in beds 6 to 12 inches (15 to 30 cm) thick with thin shaly partings 1 to 2 inches (2.5-5 cm) thick. A rectangular fracture is common. Cliff exposures display a characteristic banded appearance, alternating black, grey, to greenish grey in colour. The beds of the Sauzei zone at localities 15 to 19 are characterized by white to buff weathering bands that range from ¼ inch to 10 inches thick of fine ash-fall tuffs that become finer grained and thinner to the northeast. The beds of the overlying Humphresianum zone have few, if any, beds of ash-fall tuffs. At locality 20 near Diagonal Mountain (Jeletzky, 1976) reddish-brown weathering bands of siliceous ash-fall tuffs characterize beds containing Sauzei zone sonniniids whereas the overlying Hymphresianum zone includes no tuff bands. At locality 21, sonniniids are associated with siliceous reddish-coloured "Toodoggone volcanics" (Gabrielse *et al.*, 1976) and these may well be the source of the ash-fall tuffs at locality 20. At locality 22 a sequence of red, grey, and green siliceous flows and fragmental volcanic rocks are stratigraphically below Lower Callovian and/or Bathonian shale. These may be correlative with the "Toodoggone volcanics" as well as other volcanic rocks mapped by Eisbacher (1974). If this is so, a broad area of early Middle Bajocian volcanics was deposited on the northern and northeastern margin of the Middle Jurassic marine basin.

Triassic Rocks near Manson Creek (93 N)

Samples of black, carbonaceous limestone were collected from a sequence of black, carbonaceous, phyllitic shale, siltstone, and minor black limestone near Manson Creek (Fig. 12.2). In the past (Armstrong, 1949, p. 40) these beds have been included with the Manson Creek belt of the late Paleozoic Cache Creek Group. The samples yielded conodonts that B.E.B. Cameron (pers. comm.) has dated Late Karnian (Late Triassic). These rocks correlate with and are lithologically similar to Upper Triassic rocks near Prince George and Quesnel, in Quesnel Lake and Bonaparte Lake areas, and in areas farther southeast (Campbell, 1963; Campbell and Tipper, 1971). Throughout this belt these phyllitic rocks lie east of the volcanic Upper Triassic rocks or interfinger with them.

Lower Jurassic Rocks of Queen Charlotte Islands

In 1974 the type section of Maude Formation was examined in detail on the southeast side of Maude Island, Skidegate Inlet in Queen Charlotte Islands. Part of the Kunga Formation conformably below the Maude Formation and part of the Yakoun Formation in fault contact with the upper part of it were also investigated. In 1975 the exposures of Maude Formation at Whiteaves Bay, southeast of Maude Island on the north shore of Moresby

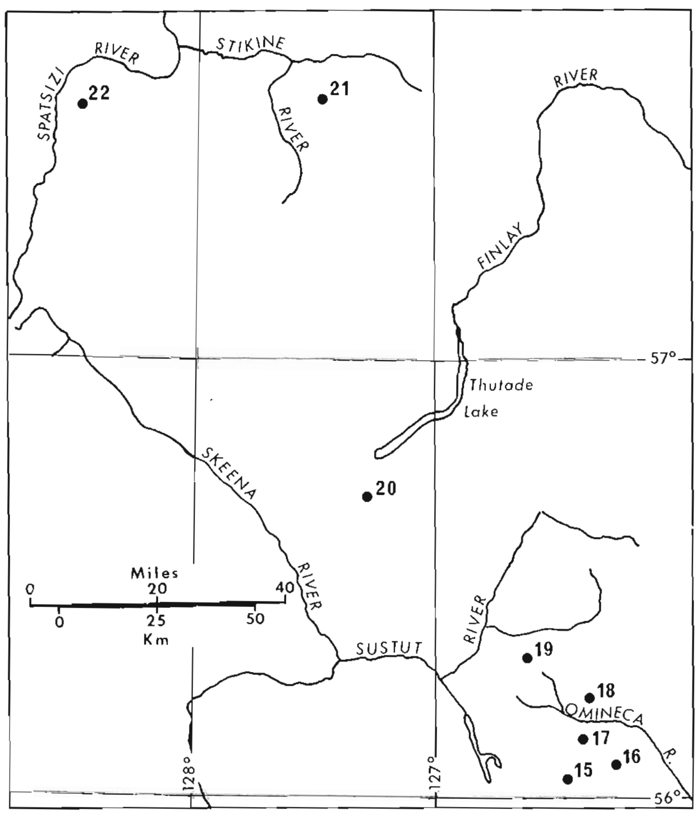

Figure 12.3. Index map of Bajocian localities.

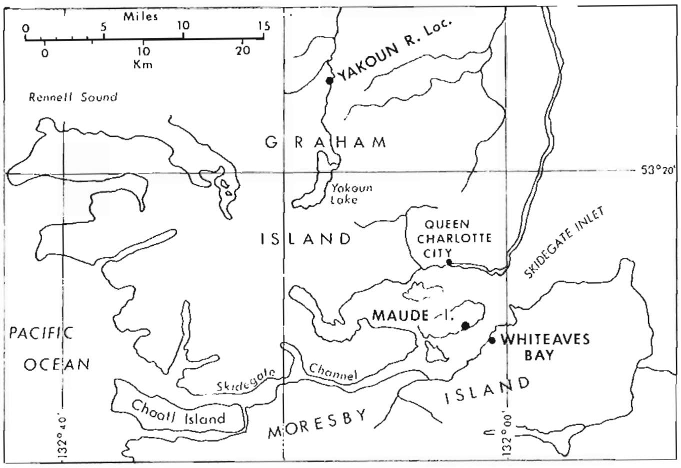

Figure 12. 4. Index map of Skidegate Inlet and vicinity, Queen Charlotte Islands.

Island (Fig. 12. 3) were studied. Although not as thick or as fossiliferous as the type section, the succession served to clarify and confirm stratigraphic details of the Maude Formation.

In these two sections, the Maude Formation is entirely sedimentary. It is a minimum of 450 feet (137 m) thick comprising predominantly black to dark grey shale and siltstone in the lower 200 feet (61 m) overlain by 100 feet (30. 5 m) of interbedded thin grey limestone beds, shale, and siltstone with large lenses or nodules of limestone, overlain in turn by about 100 feet (30. 5 m) of well-bedded, fine to medium grained sandstone and, finally, overlain by 50 to 60 feet (15-18 m) of limy siltstone with large limestone lenses or nodules. Except for the sandstone, the formation displays 6- to 12-inch (15-20 cm) thick uniform beds. Shale and siltstone predominate. The sandstone is calcareous, well-sorted in 6- to 36-inch (15-90 cm) beds that are distinctly laminated but not cross-bedded. The fauna is almost entirely ammonites and belemnites except in the bed with Fanninoceras and Arieticeras where pelecypods and brachiopods abound. Well-preserved radiolaria have also been obtained from several beds well-dated by ammonites. The top of the section is in fault contact with the Yakoun Formation of Bajocian and Callovian age and the base grades abruptly into finely banded argillites of the Kunga Formation.

The Maude Formation has yielded a rich and varied ammonoid fauna in current and earlier investigations (McLearn, 1949; Sutherland Brown, 1968). The more abundant and significant genera are indicated in Figure 12. 5. Early Pliensbachian time is represented

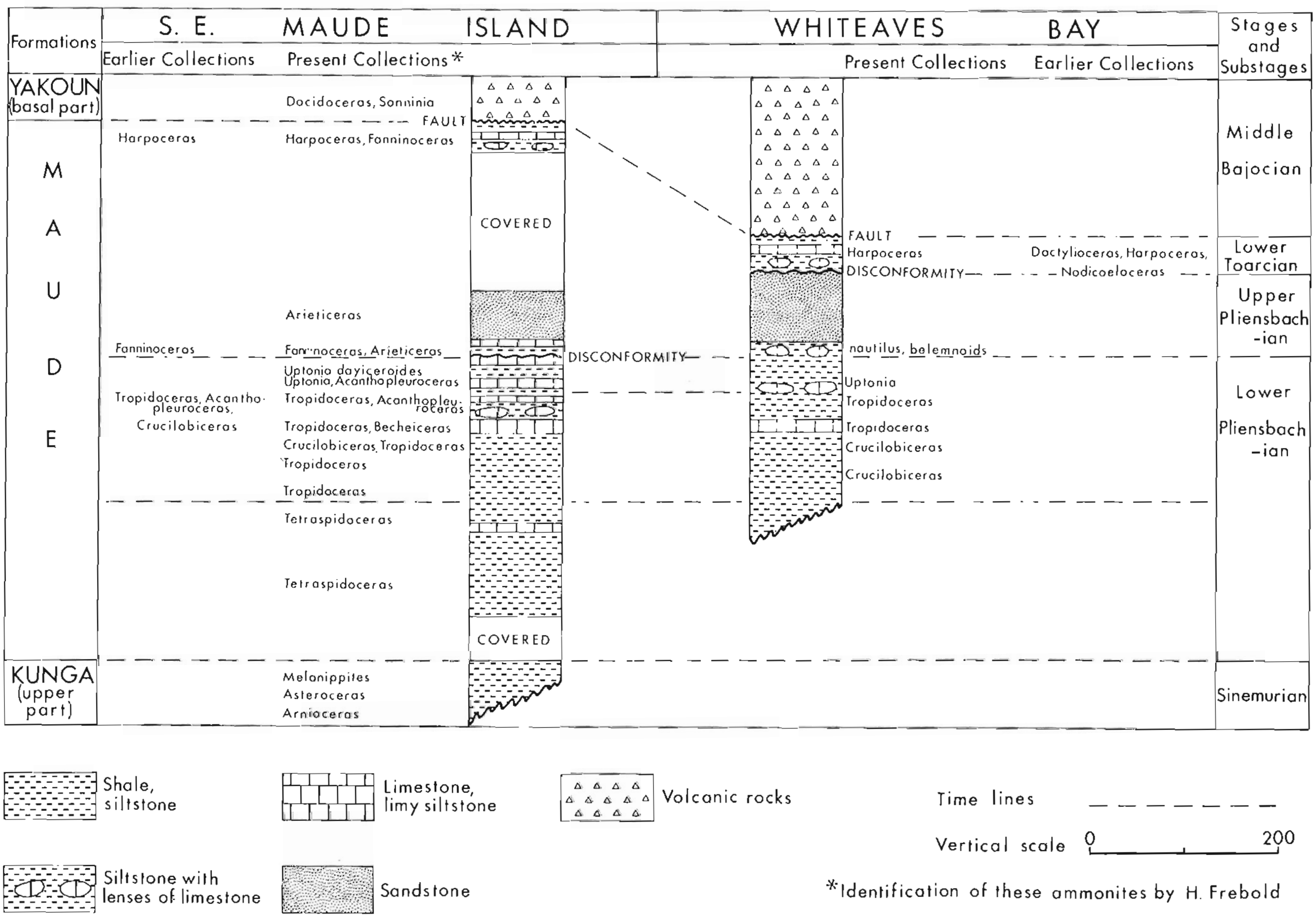

Figure 12. 5. Columnar sections of Maude Formation.

by a fauna comparable to that of the Jamesoni and the Ibex zones of the European standard and these are succeeded by a Late Pliensbachian fauna with Fanninoceras and Arieticeras. A hiatus is suggested both above and below the Late Pliensbachian sandstone beds (Fig. 12.5) but no angular discordances were seen. The overlying Lower Toarcian beds apparently have only the Falcifer zone represented; the section is cut off by a fault immediately above. The similarity of the Maude Formation to the earliest known beds of the Yakoun Formation and the finding of a Middle Toarcian ammonite in drift suggest that the entire Toarcian section may be more complete but has not as yet been encountered or recognized.

One sequence on Maude Island and another along Yakoun River that were referred by Sutherland Brown to the Maude Formation have yielded an early Middle Bajocian fauna and the rocks, although essentially sedimentary are coarser and clearly volcanogenic in whole or in part. Tentatively it seems wise to refer these sections to the volcanic and sedimentary Middle Jurassic Yakoun Formation with which they are in unfaulted, conformable stratigraphic contact.

References

Armstrong, J.E.
1949: Fort St. James map-area, Cassiar and Coast Districts, British Columbia; Geol. Surv. Can., Mem. 252.

Campbell, R.C.
1963: Quesnel Lake, east half, British Columbia; Geol. Surv. Can., Map 1-1963.

Campbell, R.B. and Tipper, H.W.
1971: Geology of Bonaparte Lake map-area, British Columbia; Geol. Surv. Can., Mem. 363.

Eisbacher, G.H.
1974: Sedimentary history and tectonic evolution of the Sustut and Sifton basins, north-central British Columbia; Geol. Surv. Can., Paper 73-31.

Gabrielse, H., Dodds, C.J., and Mansy, J.L.
1976: Operation Findlay; *in* Report of Activities, Part A, Geol. Surv. Can., Paper 76-1A, rept. 17.

Jeletzky, O.L.
1976: *In* Report of Activities, Part A, Geol. Surv. Can., Paper 76-1A, rept. 13.

Jeletzky, O.L.
1976: Takla Project: Preliminary report on stratigraphy and depositional history of Middle to Upper Jurassic strata in McConnell Creek map-area (west half), British Columbia; *in* Report of Activities, Part A, Geol. Surv. Can., Paper 76-1A, rept. 13.

McLearn, F.H.
1949: Jurassic Formations of Maude Island and Alliford Bay, Skidegate Inlet, Queen Charlotte Islands, British Columbia; Geol. Surv. Can., Bull. 12.

Monger, J.W.H.
1976: Takla Project: Lower Mesozoic rocks in McConnell Creek map-area; *in* Report of Activities, Part A, Geol. Surv. Can., Paper 76-1A, rept. 11.

Richards, T.A.
1976: Takla Project: McConnell Creek map-area; *in* Report of Activities, Part A, Geol. Surv. Can., Paper 76-1A, rept. 10.

Sutherland Brown, A.
1968: Geology of the Queen Charlotte Islands, British Columbia; B.C. Dept. Mines Pet. Resour., Bull. 54.

13. PRELIMINARY REPORT ON STRATIGRAPHY AND DEPOSITIONAL HISTORY OF MIDDLE TO UPPER JURASSIC STRATA IN McCONNELL CREEK MAP-AREA (94D WEST HALF), BRITISH COLUMBIA

Project 750016

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During the 1975 field season, previously unmapped rocks of the Jurassic "Bowser Assemblage" were examined in the west half of the McConnell Creek map-area. In addition, brief examinations were made of outcrops in the east half of the map-area, as well as in adjoining map-areas. The object of the study was to establish the stratigraphy and sedimentary history of the "Bowser Assemblage" and to relate it to the tectonic evolution of the Bowser Basin during the (?) Middle to Late Jurassic.

Field work was carried out on foot from a series of small fly camps. Helicopter support for camp moves and for several reconnaissance flights was provided by the "Operation Takla" base camp at Johanson Lake. Fossils were tentatively identified in the field by H. W. Tipper, T. A. Richards and T. P. Poulton.

Geology

It was found that the vast area formerly considered to be underlain almost exclusively by rocks of Upper Jurassic "Bowser Assemblage" is geologically far more complex than anticipated. In at least two widely separated areas, lower Middle Jurassic strata were mapped. At one of these, the sedimentary succession appears to be conformable with Upper Jurassic rocks.

Upper Jurassic, as well as the underlying Middle Jurassic strata, are penetratively deformed into a series of tight asymmetrical folds overturned to the northeast and east. A pervasive cleavage, which dips steeply to the west dies out to the east near the north-south trending Red Creek Fault (Richards, 1976), and to the south. It was not observed in central parts of the Hazelton west half map-area (Richards and Jeletzky, 1974). Fold axes trend between 130° and 190°, the majority clustering around 150°. Thrust faults are common on the limbs of the folds but displacements are generally minor.

Southern Tatlatui Range

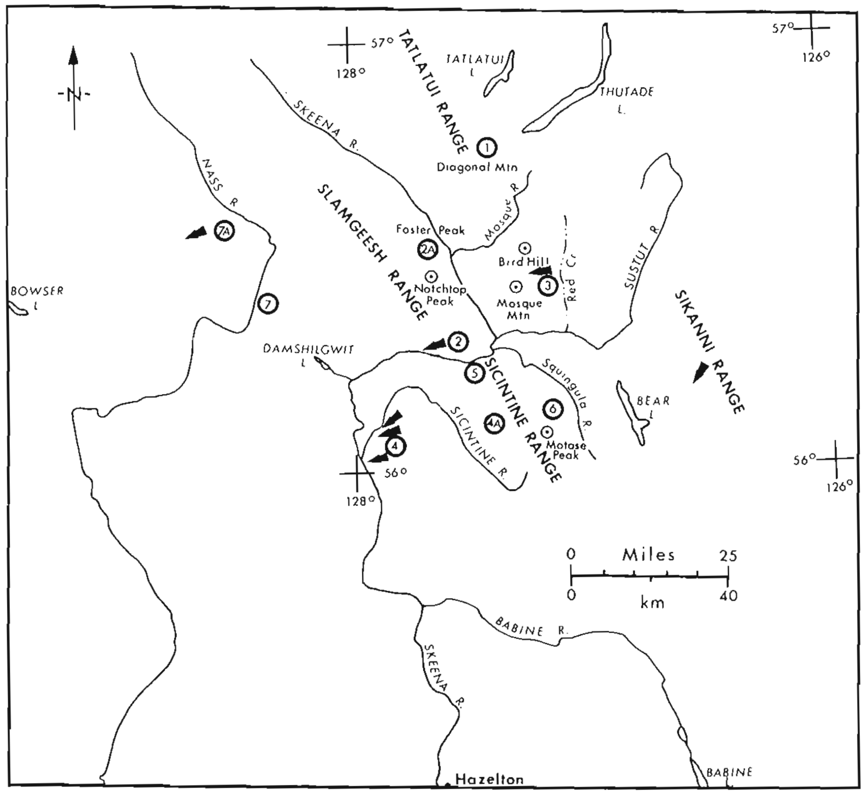
The southern part of Tatlatui Range (Fig. 13. 1), was characterized by delta front to delta slope sedimentation in Callovian to Oxfordian time. On Diagonal Mountain, rocks as old as Middle Bajocian are involved in folding and thrusting of the "Bowser Assemblage". Although the sections are far from complete, it appears that the Middle Bajocian through to Upper Jurassic rocks form a conformable sequence.

Middle Jurassic rocks consist of a minimum thickness of 800 feet (250 m) of thinly laminated dark grey siltstone with numerous thin tuff bands. Red-weathering highly acid ash-fall tuffs and the siltstones with thin black shale interbeds form composite beds 4 to 8 inches (10-20 cm) thick. Local pods to lenticular beds of light grey arenaceous limestone which are laminated, rippled and cross-laminated occur. They become more common towards the top of the section, whereas tuff bands become scarcer. In addition, limestones are more rippled and cross-laminated in the upper part. Rare, very fine grained sandstone beds also appear in the uppermost part of the section and indicate an overall shallowing of the sequence upwards. Ammonites found in siltstones and limestones have tentatively been identified as *Stephanoceras* sp. and *Sonninia* sp. by H. W. Tipper. Rare belemnites occur in the lowermost part of the succession.

Bathonian rocks containing a prolific ammonite fauna of (?) Early Bathonian age are poorly exposed marine siltstones to very fine grained sandstones with rare lenticular beds of grey arenaceous limestone.

The Callovian succession represents an interfingering of sediments of prodelta facies with those of a delta front, including some very shallow water perhaps intertidal facies. Intensely rippled fine sandstones to siltstones (in part flaser bedded) predominate at the base and are gradationally overlain by finely laminated siltstones to silty mudstones of the prodelta. These rocks grade upwards into more rippled siltstones and fine sandstones with a few beds of medium-grained sandstone. The sandstones show small scale to medium scale cross-bedding, convolute lamination, and, being moderately well sorted, exhibit virtually no grading. Rare grey, sandy limestone beds occur as pods or elongated lenses. The fauna is sparse, consisting of ammonites, rare belemnites and extremely rare pelecypods. *Kepplerites* sp., *Lilloettia* sp. and *Cadoceras* sp. have been identified tentatively.

Callovian rocks appear to be conformable with overlying Lower Oxfordian strata on Diagonal Mountain, and at several other localities in the McConnell Creek map-area (Red Creek and Northern Sikanni Range). The major change appears to be the sudden influx of coarser clastics slightly higher stratigraphically than the first appearance of the Lower Oxfordian ammonite, *Cardioceras* sp. Another possible "marker" in this area may be the first appearance of clasts of dark grey to black (?) Cache Creek chert. The lowest conglomeratic units in the Oxfordian, such as those on Diagonal Mountain, are composed almost entirely of clasts of acid to intermediate volcanics derived from the Takla-Hazelton assemblage. The environment of deposition of this facies was probably a series of distributary channels and interdistributary bays into which a few ammonites drifted from the open sea. The sandstone bodies are thick and trough cross-bedded.



(1) Southern part of the Tatlatui Range
(2) Southern part of the Slamgeesh Range; (2A) Notchtop Peak and Foster Peak areas
(3) Northern part of the Hogem Range, including the Red Creek area and Mosque Mountain
(4) Mount Tommy Jack. (4A), Western part of the Sicintine Range
(5) Northern part of the Sicintine Range
(6) Northeast spurs of Motase Peak
(7) Western Bowser Lake Map-area. (7A) Westernmost observed locality of oyster banks

Arrows indicate paleocurrent directions

Figure 13. 1.

Index map of McConnell Creek and adjoining Nass River, Bowser Lake and Hazelton map-areas.

Figure 13. 2 (opposite page)

A. Bedding plane of whole and fragmental oysters from an upper unit of a sand bar; Mount Tommy Jack, McConnell Creek map-area (west half).

B. Section through an oyster bed in an environment similar to 13. 2a. Shells are mostly unbroken and articulated; west of the Nass River in the Bowser Lake map-area (*see* loc. 7A, Fig. 13. 1).

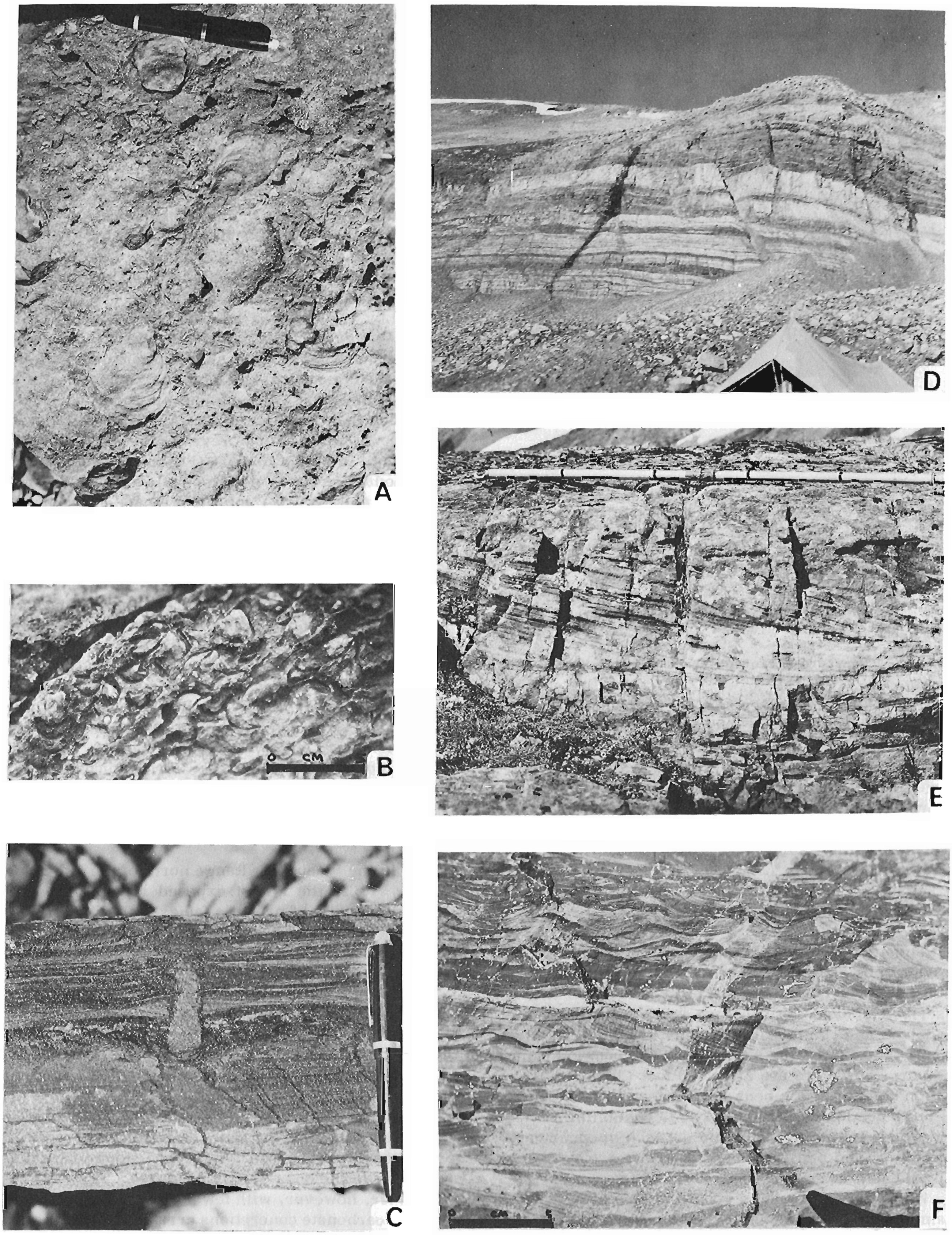
C. Burrows typical of the rippled silty sandstone of shallow subtidal facies — Mount Tommy Jack, McConnell Creek map-area.

D. Typical outcrop of shallow marine to possibly subaerial delta topsets. Note the lenticularity of units and individual beds. The dark units are predominantly fine silt showing fine plane-lamination, burrows, plant fragments and tiny isolated fine

D. sand lenticles; the light units are sandstone. Sandstone at the base is flaser bedded to mega-rippled; the more massive upper beds are extensively trough to low angle cross-bedded; Mount Tommy Jack, McConnell Creek map-area. Thickness section from talus to the top of the dark unit is about 280 feet (90 m).

E. Large scale cross-bedding in a distributary channel system of the Upper Oxfordian. The tent pole is approximately 5 feet (1. 6 m) long. Southern part of Slamgeesh Range, McConnell Creek map-area.

F. Fine sandstone showing ripple drift at base, appearance of mud flasers in the middle and fining to silt with isolated fine sand ripples at the top — Mount Tommy Jack, McConnell Creek map-area.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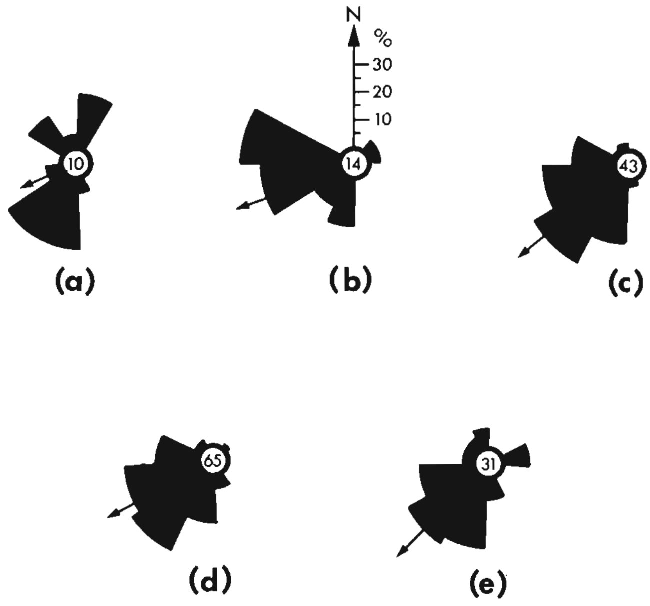

3a–3c Mount Tommy Jack
3a lower part
3b middle part
3c upper part

3d Northern Sicintine Range immediately south
3e of Skeena River

Western Sicintine Range west of Motase Peak

Figure 13. 3. Paleocurrents in the Oxfordian part of the "Bowser Assemblage" McConnell Creek map-area, west half.

Large brown-weathering carbonate concretions are invariably present, as are carbonized pieces of wood and plant fragments. Unfortunately the overlying strata are highly faulted and the sequence is poorly understood.

Northern Part of the Hogem Range

The area including Mosque Mountain, Bird Hill, and outcrops along Red Creek (Fig. 13. 1), exposes a more proximal facies of the Lower Oxfordian than described above. As in more westerly areas thick sandstone and conglomeratic sandstone units are interbedded with dark grey siltstones. The coarse grained members, however, are up to 120 feet (36 m) thick and contain conglomeratic beds up to 4 feet (1. 2 m) thick. All units are lenticular. Shell beds containing varied pelecypods including *Vaugonia doroschini* occur rarely but bioturbated beds and many burrows are evidence of biotic activity.

Where the exposure is good it can be seen that the channel sands are extensively cross-bedded (medium-sized troughs and medium to large scale tabular). Paleocurrent data indicate a sedimentary transport to the south-southwest.

Mount Tommy Jack and the Northern Sicintine Range

The lower part of the succession on Mount Tommy Jack consists of a succession of sand bars colonized by a heavy oyster population (Fig. 13. 2A) periodically destroyed by channel shifts, then built up again and re-colonized. Finely laminated siltstones between the sand bodies grades into more massive siltstone with *Buchia concentrica*, or into siltstone with isolated fine grained sand ripples (lenticular bedding), or, flaser bedded fine grained sandstone (Fig. 13. 2F). In the coarser units narrow vertical to sub-vertical burrows occur; in the siltstones, wider, less regular burrows made by (?) molluscs predominate (Fig. 13. 2C).

Above the cyclic part of the section, the sediments become coarser grained although laminated siltstones with *Buchia concentrica* (Upper Oxfordian) still occur. Oyster beds can also be found but these are much less common than in lower units. A change is apparent in paleocurrent directions revealed by polymodal paleocurrents in the lower part (Fig. 13. 3A) and a more unimodal direction, perhaps reflecting shallowing conditions and even (?) subaerial topset environments in the upper part (Fig. 13. 3B and 13. 3C). Lenticularity is high both in the individual beds and in the sandstone bodies (Fig. 13. 2D). Bedding planes in the sandstone are commonly covered by carbonized branches and logs, whereas the finer interbeds are, in many places, full of moderately well preserved zeolitized plant fragments. Fossils were not found in the upper part of the section.

The uppermost part of the succession contains sands which are better sorted and more texturally mature than in the lower part. Bases of beds are locally conglomeratic. A similar sequence is exposed in the western part of the Sicintine Range west of Mount Motase. Paleocurrents, as at Mount Tommy Jack, are fairly unimodal and directed to the southwest (Fig. 13. 3e).

Southern Part of the Slamgeesh Range

In the Slamgeesh Range north of the Skeena River (Fig. 13. 1) the lowest exposed rocks represent an interfingering of prodelta facies with distal delta-front sediments. The rocks are mostly siltstones to silty mudstones with a few interbedded sandstone bodies with abundant rip-up clasts concentrated in vague layers. These sandstones probably represent sand swept off the unstable edge of a delta front rather than deep water turbidites which they superficially resemble. The sandstones are moderately well sorted, are not graded, and are trough cross-bedded and lenticular where exposures are good.

Higher in the succession flaser- to lenticular-bedded silts and fine sands, probably representing a shallower environment, are interbedded with sandstones similar to the ones described above. They are slightly cleaner, however, with fewer rip-up clasts. Grapefruit-sized carbonate concretions arranged in distinct rows are abundant. Large scale tabular cross-beds (Fig. 13. 2E) are common.

At this locality no fossils were found because of intense cleavage. The facies was traced about 50 miles (80 km) northward along Skeena River and at various places fossils indicating a mid-Oxfordian age were collected. *Cardioceras* sp. and *Buchia concentrica* as well as other as yet unidentified ammonites occur. The facies was also traced to the western part of Mount Motase and at least 50 miles (80 km) westward into Bowser Lake map-area. *Buchia concentrica* was found in the rocks, establishing their contemporaneity with the Slamgeesh and Sicintine Range localities. This facies is also a time equivalent of the shallow marine to (?) subaerial topsets of Mount Tommy Jack and the northern Sicintine Range.

Other reconnaissance stops in the Bowser Lake map-area also showed shallow water facies instead of the expected shale-turbidite basin. Northwest of Damshilgwit Lake (Fig. 13.1), a succession containing silts with well preserved large ferns and other plants, thin coal seams and mats of zeolitized plant remains is interbedded with a few fairly well sorted cross-bedded sandstones indicating topset swamp to lagoonal environment.

About 5 miles (8 km) west of the Nass River (Fig. 13.1), an oyster bank facies probably represents sand bars. The lithology, fauna and cyclicity are all nearly identical to those of Mount Tommy Jack. The sandstones are similarly intensely cross-bedded, with paleocurrent data showing transport to the southwest.

Northern Sicintine Range

Distributary channel facies predominate in this area as indicated by the dominant cross-bedded sandstones with interbeds of sandy silt to fine siltstone. The latter are intensely current-rippled. Interbeds include carbonaceous mudstone and beds of solid broken plant material including macerated ferns. No fossils were found, thus suggesting a subaerial delta topset origin. Paleocurrents are unimodal to the southwest (Fig. 13.3d). The rocks are believed coeval with the uppermost part of the Mount Tommy Jack section.

Work in the Hazelton map-area (Richards and Jeletzky, 1974), indicates that there were two periods of delta progradation in the Upper Jurassic, the first in the Lower Oxfordian and the second in the Upper Oxfordian. Deltaic sedimentation in the McConnell Creek map-area probably followed a similar pattern. The first progradation in the Hazelton map-area, known as the Trout Creek facies are thought to be the sequences in the northern part of the Hogem Range (i.e. Red Creek, Bird Hill and Mosque Mountain). The second episode, occurring above *Buchia concentrica* near Hazelton, are perhaps equivalent of the oyster beds of Mount Tommy Jack, and possibly those in the Nass River. Significantly, the shell beds of this episode are everywhere characterized by an oyster fauna indicating lowered salinities.

The work has shown that facies in the "Bowser Assemblage" are definitely mappable units on a regional scale which can be given formation status. It is already apparent that the facies typical of the Slamgeesh Range is mappable within the McConnell Creek map-area. A very similar (lithologically) succession of rocks also outcrops in the Hazelton map-area and has informally been called the Thomlinson facies after its main area of outcrop (Richards and Jeletzky, 1974). Its age has also been established as being Late Oxfordian by the presence of *Buchia concentrica.* Similarly the "coquina" facies (both the Upper and Lower Oxfordian) are mappable units within three map-areas. However, due to the lithological subtleties involved, mapping at the 1:250 000 scale is inadequate to completely delineate rocks of the "Bowser Assemblage".

References

Richards, T.A.
1976: Takla Project: McConnell Creek map-area (94D east half); *in* Report of Activities, Part A, Geol. Surv. Can., Paper 76-1A, rept. 10.

Richards, T.A. and Jeletzky, O.L.
1974: A preliminary study of the Upper Jurassic Bowser Assemblage in the Hazelton map-area, west half, British Columbia (93M W½); *in* Report of Activities, Part A, Geol. Surv. Can., Paper 75-1A, p. 31-36.

14. PLUTONIC ROCKS OF McCONNELL CREEK (94 D WEST HALF) AND AIKEN LAKE (94 C EAST HALF) MAP-AREAS, BRITISH COLUMBIA

Project 750016

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Two months were spent mapping the northern Hogem Batholith (north of 56°N) and examining granitic rocks of McConnell Creek (94 D E½) map-area.

Hogem Batholith

The northern Hogem Batholith is a structurally and lithologically complex body of plutonic rocks emplaced into weakly metamorphosed strata of the Takla Group. The granitoid rocks of the batholith were separated into four major tectonic units, each having distinctive lithologic[1] and structural characteristics (Fig. 14. 1):

1. Mafic-rich, quartz-poor plutons
2. Biotite and quartz-rich foliated plutons
3. Syenite and related rocks of the Duckling Creek Syenite Complex
4. Leucocratic, massive granodiorite plutons

Mafic-rich quartz-poor plutons

Three bodies of dark dioritic to monzonitic rock comprise nearly half the northern Hogem Batholith. Both extend south into the area mapped by Garnett (1969). The Thane Creek pluton forms most of the eastern half of the map-area and is dominantly a dark green-grey diorite to monzodiorite. Mafics comprise between 15 and 30 per cent of most rocks; hornblende is generally more abundant than biotite. The quartz content is almost everywhere less than 5 per cent, except in the extreme southeastern part of the pluton where it may be as high as 15 per cent. Potassium feldspar is interstitial and forms up to 70 per cent of the rock, but amounts greater than 20 per cent are rare. The Detni Creek diorite body is similar to the Thane Creek pluton except for its uniformly low K-feldspar and quartz contents.

All rocks in the Thane Creek and Detni Creek bodies show some degree of alteration. In the least altered rocks hornblende is the dominant mafic mineral, but in most samples the hornblende is largely altered to fine grained, randomly-oriented biotite and chlorite. Plagioclase is commonly greenish due to epidotization, and many plagioclase crystals are recrystallized to a sugary mosaic of finer grains. Alteration is most intense along epidotized fractures; the rock outwards from the fractures for several centimetres is commonly salmon-pink in colour. In some areas these fractures are closely spaced, and rocks over the entire outcrop are pink, with green mafics. Staining indicates that in some samples the pink coloration is due to K-feldspar introduced along fractures, but in many specimens the pink mineral is plagioclase. The overall impression is that a fluid-filled fracture system led to the low-temperature alteration and hydration of these plutons to a greenschist facies assemblage. The random orientation of the biotite indicates that little or no directed stress accompanied the alteration. The irregular distribution of K-feldspar on the scale of both hand specimen and pluton and the obvious fracture-control of much of the K-feldspar, may indicate that most, if not all, of this mineral was introduced metasomatically during alteration.

Biotite-rich quartzose plutons

The large body of foliated plutonic rock (Mesilinka pluton) along the western side of the northern Hogem Batholith shows a wide variation in composition, but is predominantly medium grained quartz monzodiorite and granodiorite. Quartz rarely forms less than 5 per cent and commonly as much as 20 per cent of the rock. Mafic minerals, in contrast to the Thane Creek and Detni Creek bodies, are dominantly biotite and rarely constitute less than 15 per cent of the rock. Potash feldspar forms both small interstitial grains and porphyroblasts up to several centimetres across. These megacrysts vary from euhedral crystals with few inclusions to poorly defined patches, conspicuous only on stained surfaces, which contain abundant inclusions of plagioclase, mafic minerals and quartz. One area of the pluton (unit P2A) contains little K-feldspar and is mapped as quartz diorite, but is otherwise indistinguishable from the rest of the pluton.

The Mesilinka pluton is the most heterogeneous unit in the northern Hogem Batholith. A single outcrop may show large variations in composition, texture, and degree of foliation. Many outcrops have streaks and schlieren of mafic material, and narrow screens or large irregular inclusions of amphibolite. Some areas of the pluton are agmatitic and are cut by several generations of aplite and pegmatite dykes. A striking feature of the area south of Ferriston Creek is a horizontal "sheeting": layers of megacrystic quartz monzodiorite about 70 metres thick alternate with sheets up to 30 metres thick of dark, fine grained, amphibolitic material. This alternating sequence is laterally continuous across some cirque walls and may repeat over vertical distances of up to 600 metres. The origin of the sheeting is not known. Possibly the amphibolitic material represents the metamorphosed roof of the pluton that has been forcibly injected by the quartz monzodiorite.

[1] The lithologic classification used is that given by IUGS Subcommission on the Systematics of Igneous Rocks, Geotimes, October 1973, p. 26-30.

The Mesilinka pluton in contrast to the Thane Creek body, shows relatively little alteration. In most areas of the pluton biotite and feldspars are fresh, but in some zones of intense deformation biotite is completely chloritized.

The roughly elliptical quartz monzodiorite pluton (unit P2B) south of Abraham Creek is tentatively included with this unit, but is much more homogeneous than the Mesilinka pluton. The pluton is moderately foliated, coarse grained, and has biotite as the only mafic mineral. The body intrudes the Thane Creek pluton and the small body of monzodiorite to the northwest, and in turn is cut by massive granodiorite of unit P5. Its relationship with the Mesilinka pluton is not known.

Duckling Creek Syenite Complex

Only the extreme northern end of the Duckling Creek Syenite Complex was examined. As in the area to the south mapped by Garnett (1969), the complex shows wide variations in grain size, texture and composition and subdivision was not attempted. Most rocks are syenitic to monzonitic in composition, but screens of schist and irregular blocks of biotite pyroxenite are common. Schist and pyroxenite have been strongly

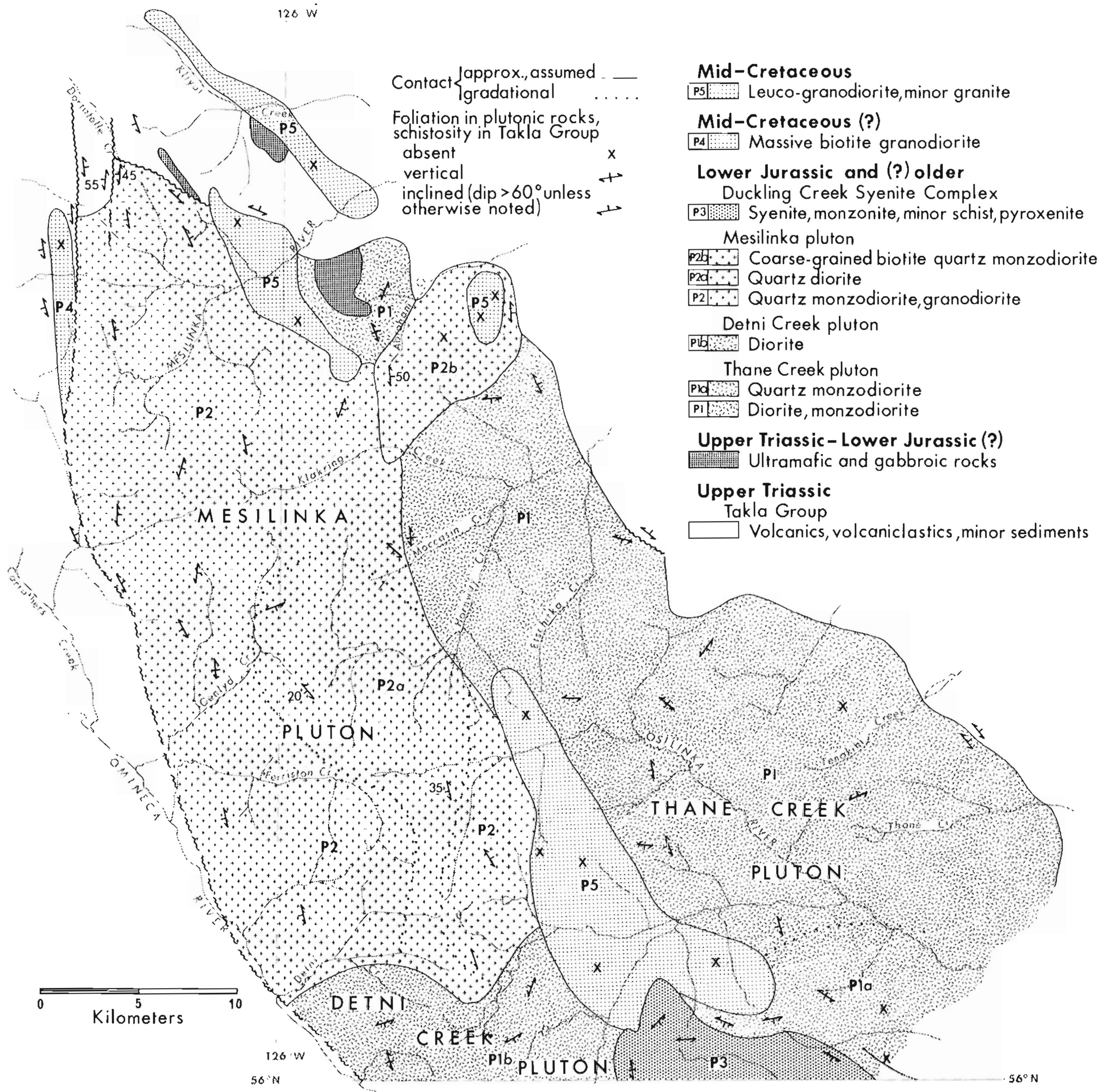

Figure 14. 1. Geological sketch map of the Northern Hogem Batholith.

affected by K-feldspar metasomatism, and their original nature is uncertain. The syenite complex, though, shows little field evidence of the pervasive hydrothermal alteration and recrystallization that has affected the Thane Creek pluton.

Massive Granodiorite Plutons

Several bodies of massive homogeneous granodiorite to granite are clearly intrusive into all other units in the northern Hogem Batholith. These bodies are elongated northwesterly, parallel with the dominant structural grain of the batholith. The granodiorite is characterized by a very low mafic content (usually less than 3 per cent); biotite is the sole mafic mineral. Conspicuous quartz eyes are common, and miarolitic cavities are present here and there.

The small elongate pluton (unit P4) along the extreme northwest margin of the Hogem Batholith differs from the other plutons of this group only in having a higher mafic content.

Structure and Age-relations

With the exception of the massive granodiorite plutons, all units in the northern Hogem Batholith are foliated to some extent. Foliation in the Thane Creek and Detni Creek plutons is well-developed, but has been partly obliterated by alteration and recrystallization as described above. Foliation is generally parallel with the pluton margins, but the overall patterns appear complex. Foliation in the Mesilinka pluton, on the other hand, consistently strikes north to northwest, with only local deviations from this trend, and at neither the north nor south ends of the pluton does the foliation show any east-west trend.

The foliation in the Mesilinka pluton is formed by superposed cataclasis. Aplites, pegmatites, and quartz veins cutting the pluton are foliated parallel with the plutonic rock, rather than with the vein margins. The horizontally-sheeted area south of Ferriston Creek has a steep northwest-trending foliation that has been imposed on both plutonic rock and the horizontal amphibolitic layers. Cataclasis within the pluton is greatest on the west side, especially at the north end where the marginal parts of the pluton are commonly augen gneisses and mylonites. Adjacent Takla strata are also sheared, and schistosity is parallel with foliation in the adjacent pluton.

The Duckling Creek Syenite Complex has a well-developed foliation defined by parallelism of pyroxene and K-feldspar. In the schistose screens the foliation is parallel with that in the enclosing syenite and with the long dimension of the screens. Many outcrops of syenite show a well-developed layering defined by the relative proportions and grain sizes of constituent minerals; this layering is everywhere parallel with the foliation. The foliation pattern parallels the margins of the syenite complex and, at the extreme northwest end, dips south and southeast, inwards towards the syenite.

Foliation is absent from the leucocratic granodiorite plutons and from the small granodiorite pluton along the northwest margin of the batholith.

Contacts between the Thane Creek pluton and the Takla Group are commonly marked by zones of slightly schistose hornfels and by leucocratic dykes cutting the Takla strata. North of Matetlo Creek the contact is marked by a series of faults, parallel with the contact, that offset dioritic dykes cutting the Takla Group near the batholith. Contacts between the massive granodiorite plutons and other rocks are marked by a zone in which the intruded rocks are cut by swarms of leucogranodiorite dykes up to 1 km from the contact.

The contacts between the Mesilinka pluton and Takla strata are, wherever examined, faults marked by mylonite zones. In most places few dykes are found in the Takla strata adjacent to the contact, and the pegmatite, aplite, and quartz veins so abundant in the pluton do not persist into the country rock. At the head of Dortatelle Creek the northern contact of the Mesilinka pluton dips about 45° north, under the Takla Group strata. Rolled K-feldspar megacrysts in the plutonic rock indicate that the movement of Takla rocks during mylonitization was to the southwest, up and over the pluton. The contact between the Mesilinka and Thane Creek plutons is sharp, but their relative ages are not known.

Available data indicate a complex history for the northern part of the Hogem Batholith. The mafic, quartz-poor Thane Creek and Detni Creek bodies and the more quartzose Mesilinka pluton are the oldest plutonic rocks in the area. The intense deformation of the Mesilinka pluton (perhaps related in part of the Pinchi Fault system?) has not affected the Thane Creek and Detni Creek plutons, which have been altered and recrystallized under more static conditions. The Mesilinka pluton thus appears to be the oldest unit in this part of the Hogem Batholith. The Duckling Creek Syenite Complex is also unaffected by regional deformation; to the south the complex clearly intrudes rocks correlative with the Thane Creek pluton (Garnett, 1969). The massive leuco-granodiorite plutons and the small granodiorite body along the west margin of the batholith are the youngest plutons in the map-area.

Only one isotopic age determination is available from the area. The leuco-granodiorite stock south of Abraham Creek gave a K-Ar age of 122 ± 6 m.y. on biotite (Irvine, *in* Wanless *et al.*, 1972, p. 11). Similar plutons south of the map-area give similar ages (Garnett, pers. comm.). The Duckling Creek Syenite Complex gives K-Ar ages of about 175 m.y. (Garnett, 1974). All that can be said of the ages of the Detni Creek, Thane Creek, and Mesilinka plutons is that they are Late Jurassic or older. The intense deformation of the west margin of the Hogem Batholith and enclosing strata thus appears to have been restricted to the relatively limited interval between the deposition of the Takla Group (Late Triassic) and the emplacement of the Duckling Creek Syenite Complex (Early Jurassic).

Other Plutonic Rocks, McConnell Creek Map-Area

Except for the Eocene Kastberg Intrusion and several high-level middle Cretaceous (?) stocks in the southwest corner of the map-area, plutonic rocks in McConnell Creek map-area are almost entirely confined to a 20-km-wide belt extending north-northwest from the Hogem Batholith. Brief comments on some individual plutons are given below. Map-units and geographic names refer to Figure 10.1 (Richards, this publ., rept. 10).

The Fleet Peak pluton (unit D, in part) is a long, narrow body that extends northwest from the Ingenika Range ultramafic body along McConnell Creek and into the Toodoggone map-area (Gabrielse *et al.*, 1976) to the north. The total length of the pluton is about 55 km; its average width is about 5 km. The body is lithologically heterogeneous, but is dominantly moderately foliated diorite and monzodiorite. Quartz is usually less than 3 per cent; mafics (hornblende > biotite) are seldom less than 20 per cent. Along the Ingenika River and on Fleet Peak the pluton is distinctly layered: schlieren, irregular bands and agmatitic zones of biotite hornblende and leuco-diorite alternate with and grade into the normal diorite to monzodiorite. Lithologically, the rock is similar to the least altered parts of the Thane Creek and Detni Creek plutons in the northern Hogem Batholith.

The east margin of the pluton is in sharp contact throughout much of its length with a belt of altered volcanic rocks that may belong to the Lay Range assemblage (unit 3). Foliations in both volcanics and pluton are parallel with the pluton margins, but it is not clear if the contact is intrusive or a fault. The western margin of the pluton along McConnell Creek is intensely sheared, but the south end of the body in the Ingenika Range appears to intrude Takla Group rocks.

The Jensen Peak pluton is the large homogeneous body of unit E that underlies much of the area east of the Fleet Peak pluton. The rock is dominantly a white, massive to faintly-foliated quartz monzodiorite. Most specimens are medium-grained and equigranular, with hornblende dominant over biotite. Small miacrolitic cavities are common. The east contact of the pluton is sharp, with few dykes occurring in the adjacent strata. Northeast of Fleet Peak the east contact is a fault, but it is not known whether or not the entire east margin is faulted. The west margin of the pluton, where examined east of McConnell Creek and east of Fleet Peak, is faulted.

The stock on Mount Fredrickson is dominantly a massive pinkish quartz monzodiorite. The rock is generally coarse grained, and biotite is subordinate to hornblende. Miarolitic cavities are absent. On the west the pluton is in lit-par-lit contact with the Asitka Group strata; the eastern margin of the stock is faulted against Asitka strata. The pluton extends a short distance north into the Toodoggone map-area (Gabrielse *et al.*, 1976), where it cuts rocks of the Fleet Peak pluton. Similar stocks form much of the McConnell Range north of Jensen Creek. The roof of one of these stocks is exposed just north of Jensen Creek, and it appears that the plutons are just beginning to be exposed at the surface.

The stock south of Jensen Creek is finer grained and more mafic than those to the north. It is a massive quartz diorite to quartz monzodiorite, with hornblende being more abundant than biotite. The pluton is high-level and generally intrusive, although the east contact of the body is faulted. Very similar to this pluton is the slightly porphyritic quartz diorite stock on Asitka Peak, for which a K-Ar age of 178 ± 7 m.y. on biotite has been obtained (N. Church, pers. comm.). Other somewhat similar high-level stocks are situated east and southeast of Johanson Lake.

The large pluton at the west end of the Ingenika Range is a biotite > hornblende, medium-grained, equigranular quartz diorite. The rock has a good foliation that generally strikes west-northwest at an angle to the intrusive contact between the stock and the Takla Group. The pluton is somewhat similar to the small stock 6 km west of Fleet Peak, but is otherwise unlike any other pluton seen in McConnell Creek map-area or the northern Hogem Batholith.

East of the Ingenika Fault and west of the Lay Fault the strata of the Takla Group are deformed and metamorphosed to a slightly higher degree than strata west of the Ingenika Fault, yet roughly the same stratigraphic level is exposed in both areas (Monger, 1976). As the belt of more metamorphosed strata largely coincides with a belt containing many plutons it is tempting to ascribe the metamorphism to widespread contact metamorphism resulting from the emplacement of numerous plutons. However, the available data do not support this hypothesis. The northern end of the Hogem Batholith is in fault contact with the Takla, and both pluton and country rock have been deformed simultaneously. Further, the metamorphic grade of the Takla strata shows no obvious increase towards the Hogem Batholith. Most of the clearly intrusive stocks in the map-area, such as those near Johanson Lake, have very narrow contact aureoles that appear to be later than the more regional metamorphism. Rather than the regional metamorphism in this belt being a direct result of plutonism, it may be that metamorphism, plutonism, and regional deformation are all related consequences of other, more obscure, events.

References

Gabrielse, H., Dodds, C.J., and Mansy, J.L.
1976: Operation Finlay; *in* Report of Activities, Part A, Geol. Surv. Can., Paper 76-1A, rept. 17.

Garnett, J.A.
1969: Preliminary geological map of part of the Hogem batholith, Duckling Creek area; B.C. Dept. Mines Pet. Resour., Prelim. Map no. 9.

1974: Geology and copper-molybdenum mineralization in the southern Hogem batholith, north-central British Columbia; Can. Inst. Min. Met., Bull., v. 67, no. 749, p. 101-106.

Monger, J.W.H.
1976: Takla Project: Lower Mesozoic rocks in McConnell Creek map-area; *in* Report of Activities, Part A, Geol. Surv. Can., Paper 76-1A, rept. 11.

Wanless, R.K., Stevens, R.D., Lachance, G.R., and Delabio, R.N.
1972: Age determinations and geological studies; K-Ar isotopic ages, Report 10; Geol. Surv. Can., Paper 71-2.

PART 1: AXELGOLD GABBRO INTRUSION, McCONNELL CREEK MAP-AREA

T.N. Irvine

Supplementary to field work done in 1973 (Irvine, 1975), a week was spent mapping two satellites of the Axelgold layered gabbro intrusion and doing a reconnaissance of the northern tip of the Stuart Lake Belt of Cache Creek Group strata and associated alpine type ultramafic rocks into which the gabbro intrusion is emplaced. A principal finding was the discovery of a high angle thrust or reverse fault along the west side of the Stuart Lake Belt in the McConnell Creek area.

The satellitic gabbro bodies occur just north of the main intrusion in vaguely en echelon arrangement with it (Fig. 15.1). Both had been mapped previously by Lord (1948). The larger body is about 2.7 km (1.7 mi) long and 1.3 km (0.8 mi) wide; the other is exposed on only one small hill just north of the Omineca River and is probably no more than a kilometre in diameter. Both consist of the same distinctive, fresh, medium to coarse grained, slightly leucocratic, brown weathering olivine gabbro that is common to the main intrusion, and although they show only traces of layering, there is no doubt that they are cogenetic intrusions. Several dykes of similar gabbro were seen in the Cache Creek rocks between the larger satellite and the main intrusion and the larger satellite is cut by countless dykes of darker, fine to medium grained basalt or gabbro identical to dykes in the main intrusion (Irvine, 1975). Also like the main intrusion, the satellites do not have visible fine grained border facies or chilled margins, and both are fringed by contact metamorphic aureoles in which Cache Creek metasediments have been converted to fine grained biotitic hornfels. A small ultramafic body near the west side of the larger satellite has apparently been transformed by contact metamorphism and metasomatism from serpentinite to coarse, "regenerated" olivine (now partially reserpentinized) and interstitial phlogopite.

The dykes of dark gabbro in the larger satellite appear to form a stockwork, but because of very marked granular disintegration of the host gabbro and spheroidal weathering of the dykes, it was difficult to trace the dykes for any distance or even to define their trends. Consequently it was not possible in the time available for the study to identify any systematic pattern that may exist in their disposition. It is clear, however, that they are virtually exclusive to the Axelgold intrusion and its satellites; none were seen in the country rocks. In a previous report Irvine (1975, p. 85) suggested that the dark gabbro dykes might represent intercumulus liquid from the host gabbro that was segregated into fractures at a late stage of solidification. In view of their extensive development in the larger satellite, however, and their consistently relatively fine grain

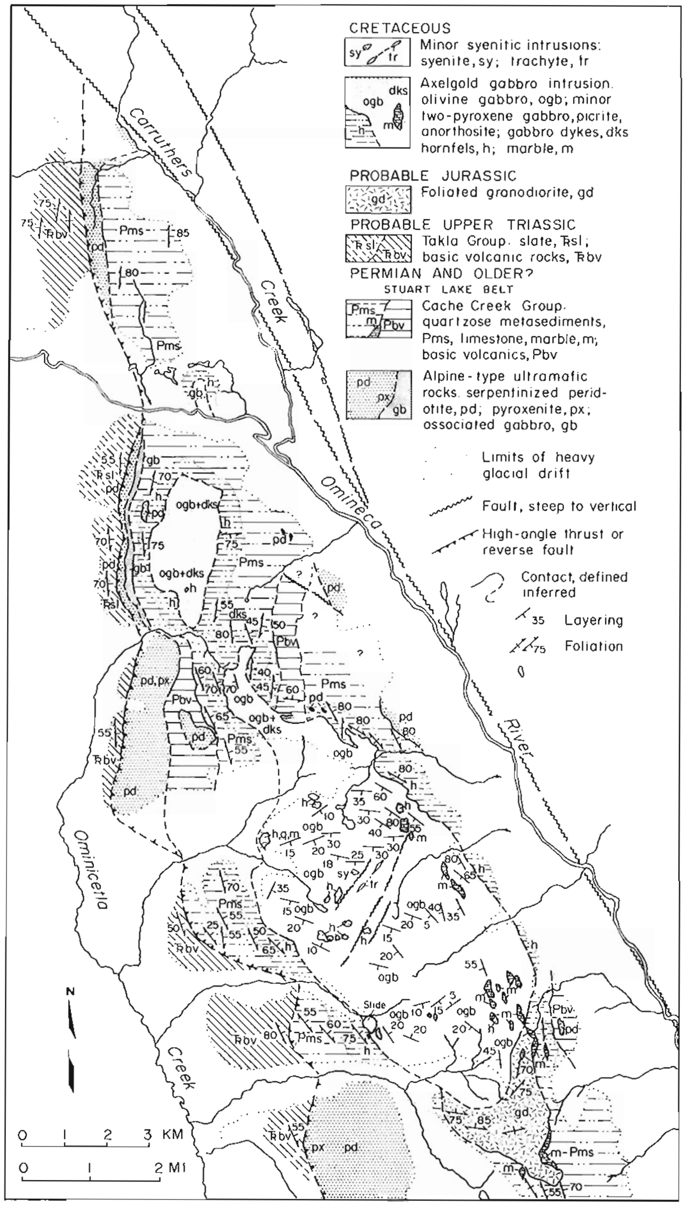

Figure 15.1. Geological map of the Axelgold Range in the McConnell Creek area, showing the Axelgold layered gabbro intrusion and its satellites, and the high-angle thrust or reverse fault (the Ominicetla thrust) discovered along the west side of the Stuart Lake Belt of Cache Creek Group rocks. The framed area extends from 56°N to 56°15'N and from 126°W to 126°15'W.

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size, it now seems more likely that they represent a distinctly later (although no doubt related) episode of primary magmatism that was closely controlled spatially by the structure of the host intrusion and its feeder system.

The trace of the thrust or reverse fault along the west side of the Stuart Lake Belt extends along the edge of the Axelgold Range just west of the Axelgold intrusion (Fig. 15.1). The fault was distinguished on the basis that regionally metamorphosed schists of the Permian Cache Creek Group are in juxtaposition with relatively less metamorphosed slate and sheared greenish volcanic rocks that are believed to belong to the Upper Triassic Takla Group (*see* this publication reports 10-13). Shearing along the junction and the foliation in both rock suites dip consistently eastward at 55°-75°, and this is inferred also to be the dip of the fault. The fault has not been mapped in detail over the southern half of its length in Figure 15.1 and undoubtedly is more complicated than indicated, but there seemed no doubt as to its existence at each of the several spots where it was examined in this interval, and it can be correlated with similarly disposed faults mapped by Patterson (1974) along the west side of the Stuart Lake Belt on Mount Ogden and in the Vital Range on strike to the southeast. Where the fault is best defined in the McConnell Creek area, the base of the uplifted block is delineated by a line of highly deformed peridotite and serpentinite bodies, a feature that was a useful guide in following its trace. Another, less well defined string of ultramafic bodies along the east side of the Axelgold intrusion may delineate a second fault of the same type.

The age of the thrust fault relative to the mid-Cretaceous Axelgold intrusion is unknown inasmuch as they have not been found in contact. On the basis that the fault was evidently a fundamental structure in the emplacement of the Stuart Lake Belt, and in the absence of any known faults of similar displacement in the gabbro intrusion, it would appear that the fault is older. A younger similarly east-dipping high-angle thrust fault, however, has been identified by Takla Project Group (reports 10 to 16) on Kaza Peak just west of the Axelgold intrusion, along which Upper Triassic Takla volcanics have been uplifted over Eocene strata. Thus there is a possibility that the fault mapped in the present work also is Eocene or younger, and that the Axelgold intrusion was uplifted as part of the thrust block.

References

Irvine, T.N.
1975: Axelgold layered gabbro intrusion, McConnell Creek map-area, British Columbia; *in* Report of Activities, Part B, Geol. Surv. Can., Paper 75-1B, p. 81-88.

Lord, C.S.
1948: McConnell Creek map-area, British Columbia; Geol. Surv. Can., Mem. 251, 72 p.

Paterson, I.A.
1974: Geology of Cache Creek Group and Mesozoic rocks at the northern end of the Stuart Lake Belt, central British Columbia; *in* Report of Activities, Part B, Geol. Surv. Can., Paper 74-1B, p. 31-42.

PART 2: ALASKAN-TYPE ULTRAMAFIC-GABBROIC BODIES IN THE AIKEN LAKE, McCONNELL CREEK, AND TOODOGGONE MAP-AREAS

T.N. Irvine[1]

In conjunction with the Takla Project, about 5 weeks were spent on further mapping and sampling of the cluster of Alaskan-type ultramafic-gabbroic bodies previously identified in the Aiken Lake and McConnell Creek areas (Irvine, 1974a), and one week was spent on another body of the same type near Lunar Creek in the Toodoggone area, discovered in 1973 by Gabrielse and Dodds (1974). The regional distribution of the rocks is shown in Figure 15.2; local maps of 9 of the bodies appear in Figure 15.3. A map of the largest body, the Polaris Complex, was presented previously (Irvine, 1974b, Fig. 2).

Alaskan-type ultramafic-gabbroic intrusions are exemplified by such occurrences as the Duke Island and Union Bay complexes in southeastern Alaska, the Tulameen complex in southern British Columbia, and the Nizhni Tagil complex in the Soviet Ural Mountains (*see* Taylor, 1967 for a general summary). They are essentially characterized by the following ultramafic rock series, although the complete series is not represented in every example: dunite; wehrlitic peridotite; olivine clinopyroxenite (locally hornblende- and (or) magnetite-bearing); hornblende clinopyroxenite (locally magnetite-rich); and hornblendite. These rocks typically are free of plagioclase and orthopyroxene, but orthopyroxene-bearing gabbro and even norite may be associated. Hornblende gabbro or diorite, and pegmatitic segregations of hornblende or of hornblende and very calcic plagioclase, are commonly present. Several of the major bodies show a rudely concentric structure, featuring a core of dunite or peridotite surrounded by successive zones of the pyroxene- and

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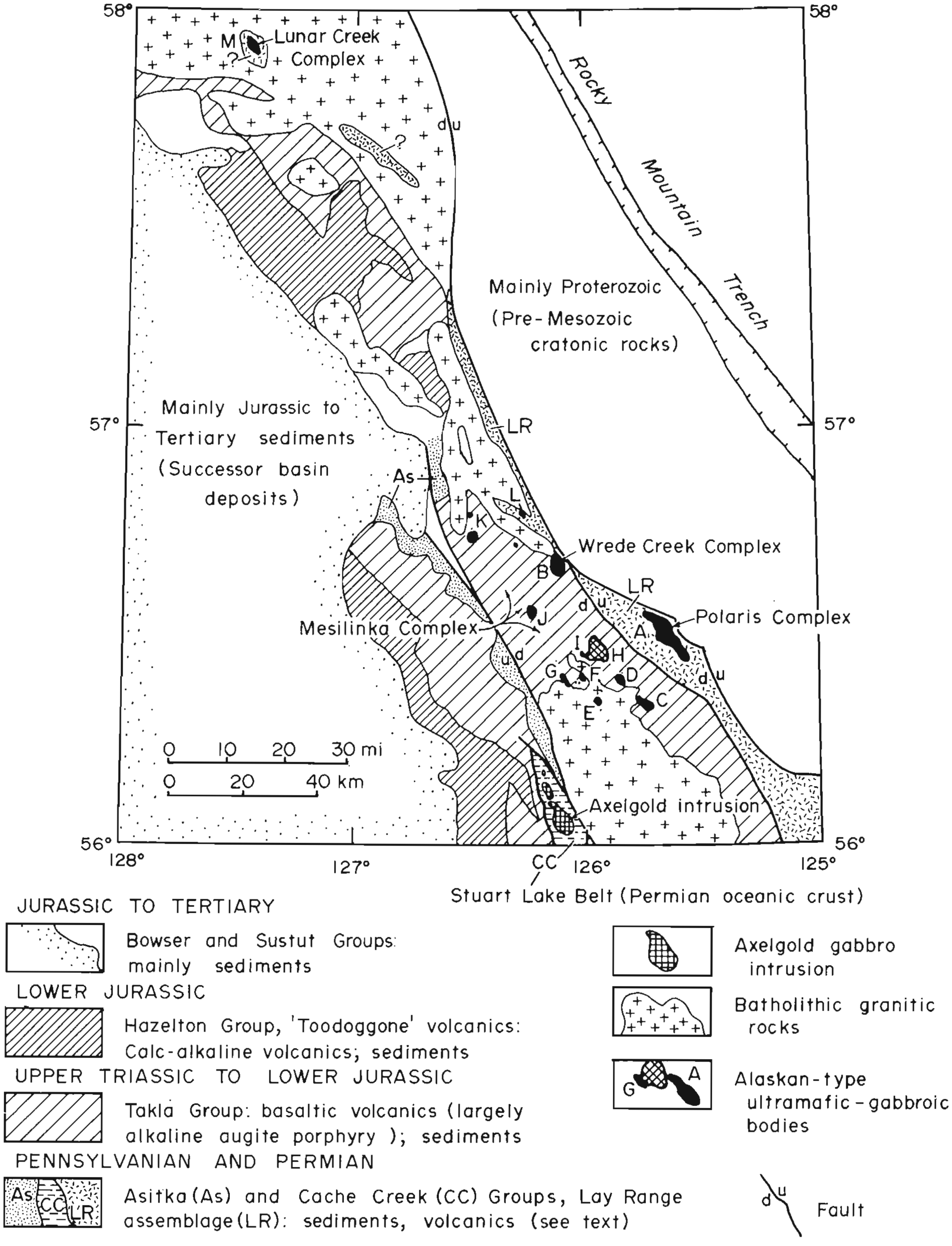

Figure 15. 2. Generalized geological map showing the regional distribution and tectonic setting of the Alaskan-type ultramafic-gabbroic bodies examined in this study. Bodies C to K are collectively termed the Mesilinka complex.

Figure 15. 3.


LUNAR CREEK
M
Lunar Cr.
5000
4000
hpxm
opx
hgb
gd
pd
du
opx,pd
tf
MENARD CREEK
K
Bladed feldspar porphyry dyke swarm
Ingenika River
Menard Cr
5500
5000
opx
gb
pd
DORTATELLE CREEK
KLIYUL CREEK
G
F
Glacier
hpxm
Gl.
pd
opx
hgb
hb
hpx
JOHANSON LAKE
J
Johanson Cr
Johanson Lake
Spawning Lake
Solo Creek
Darb Lake
hgb,peg
hgb
gb
opx
gd
5000
6000
5500
0 1 2 3 MI
0 1 2 3 4 5 KM
N
WREDE CREEK
B
Proterozoic quartzose sediments
Pls
Psl
Ctf
Cls
Ptf
du
opx
pd
hpx
hgb
6000
5000
ABRAHAM CREEK
MT. ELSIE
TUTIZZI LAKE
E
D
C
Mesilinka River
Abraham Creek
MT. ELSIE
hgt,peg
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sy
4000
5000
6000

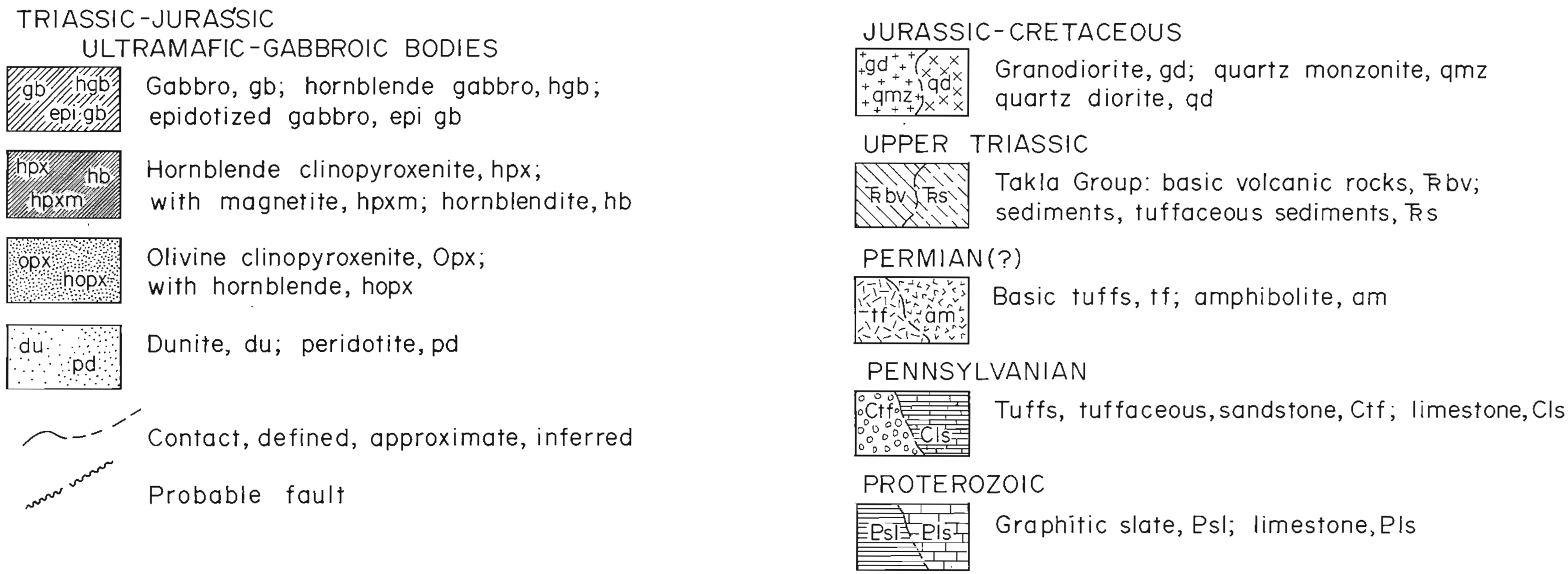

Figure 15. 3. Geological maps of 9 of the ultramafic-gabbroic bodies in Figure 15. 2. The geology around the bodies is in part from Lord (1948), Roots (1954) and from the publications, reports 10 to 16. Only limited observations were made on the Dortatelle Creek body (G) and relations in its northern part are based entirely on a description by Lord (1948).

hornblende-rich rocks and gabbro, and the bodies therefore have sometimes been called "concentrically zoned", "concentric", or "zoned" complexes, although this structure is not a general feature. The Duke Island complex has provided strong evidence that the main ultramafic rock types are cumulates and that the ultramafic rock series is essentially developed by crystallization differentiation (Irvine, 1974b). On this basis it was suggested that the concentric structure, where developed, is generally the result of diapiric re-emplacement of stratiform piles of the cumulates before they were completely solidified and cooled, the early olivine-rich units having been squeezed upward by tectonic compression into overlying pyroxene- and hornblende-rich units.

All of the bodies examined in the present study exhibit at least part of the rock series listed above but, as also is characteristic of Alaskan-type complexes, the proportions of the various rock types differ widely from body to body (Fig. 15.3). Thus the two largest occurrences, the Polaris and Wrede Creek complexes, and the bodies labelled F and G in Figures 15.2 and 15.3 are composed mainly of dunite and peridotite at the present erosion surface, whereas the Tutizzi Lake body (C) consists mainly of olivine clinopyroxenite, and the Abraham Creek (E), Johanson Lake (J) and Menard Creek (K) bodies, mostly of gabbro with only subordinate ultramafic rocks. The rocks do not show any definite primary layering, but it is apparent from other petrographic features that they are largely cumulates.

Perhaps the most interesting lithological feature concerns the relationship of the ultramafic and gabbroic rocks. In other areas, this relationship is enigmatic; in some cases the gabbroic rocks are older than the ultramafic units, in others they appear younger; at Duke Island and Union Bay they do not even seem to be derived from the same kind of magma. The bodies examined in the present work also show contradictory age relations; for example, in the Menard and Lunar Creek bodies, the gabbroic rocks contain what appear to be large inclusions of various ultramafic rocks, whereas the gabbro in the Abraham Creek body is cut by dykes or veins of hornblendite. However, the association of ultramafic and gabbroic rocks is common to so many intrusions that there can be little doubt of a direct genetic relationship. Contacts between gabbro and the dunite and peridotite typically are sharp (within a few metres), but this relationship is to be expected from crystallization differentiation processes and is also characteristic of the classic differentiated stratiform intrusions. On the other hand, several bodies (most notably Johanson Lake and Abraham Creek) show a transition from pyroxene-bearing gabbro to hornblende gabbro, to hornblende-plagioclase pegmatite and hornblendite such as occur in the ultramafic rock series.

Concentric zoning is not clearly exposed in any of the bodies but is suggested in several cases. The Wrede Creek complex comprises a roughly circular mass of dunite and peridotite, about 4 km (2.5 mi) in diameter, fringed around more than half its perimeter by small outcrops of olivine clinopyroxenite, hornblende clinopyroxenite, and hornblende gabbro. The Polaris complex shows a similar, but more complicated arrangement (Irvine, 1974a), and in the Lunar Creek and Johanson Lake intrusions, the ultramafic rocks are largely enveloped by gabbro. In the other bodies, the different rock types occur in various irregular patterns, with perhaps the only systematic feature being that they tend to occur spatially in the sequence listed above, presumably because that was the order in which they were differentiated. In all cases, however, the distribution can be readily interpreted as the result of deformation of cumulates during diapiric re-emplacement, and this interpretation is supported by detailed structural and petrofabric relations. In the Polaris, Wrede Creek and Lunar Creek complexes, peridotitic rocks with cumulus characteristics both intrude adjoining olivine clinopyroxenite as dykelike bodies and include irregular to angular blocks of it. Microscopic penetrative deformation is widespread in the olivine-bearing rocks, and the Wrede Creek peridotite locally is visibly cleaved on the outcrop scale. In the Lunar Creek complex, the dunite and peridotite show pronounced flow banding and deformation; the gabbro and diorite are strongly foliated; and large slab-like units of amphibolite, lithologically distinct from the schistose mafic tuffs that host the complex, appear to have been carried in with the intrusive rocks when they were re-emplaced.

On a regional basis, the ultramafic-gabbroic bodies occur in two types of host rocks. The Polaris complex, body L, and the Lunar Creek complex are emplaced in dark grey to epidote green, fine grained, locally well-bedded, siliceous mafic tuffs, while the other bodies intrude Takla Group basaltic augite porphyry fragmental rocks and metasediments. The Polaris host rocks stratigraphically overlie Pennsylvanian strata and undoubtedly predate the Takla Group, which is regionally dated as Upper Triassic. Similar tuffs on strike about 20 km to the southeast of Polaris have yielded conodonts tentatively dated as Permian (J. W. H. Monger, pers. comm., 1975). The host rocks of the Lunar Creek complex are not dissimilar and, on the basis that they lie in the same general tectonic zone, might be correlative. Granitic rocks of the region, which are generally Jurassic to Cretaceous in age, are younger than the ultramafic-gabbroic bodies at all places where definitive relations have been seen.

The exact age of the ultramafic-gabbroic bodies, however, is still uncertain. Hornblende and biotite from a peridotitic rock in the Polaris complex gave Jurassic K-Ar ages of 155 and 164 m.y., respectively, but the significance of these values is somewhat doubtful in view of the extensive granitic plutonism in the region at that time. The close spatial association of the bodies with the Takla volcanics suggests a genetic relationship, and on the basis of petrography at least, the pyroxenite and gabbro units could be derived from magma like the Takla augite porphyry. However, no volcanics have been found in the region sufficiently rich in olivine to be representative of the parental magma of the dunite and peridotite, in either the Triassic or Jurassic.

Tectonically, the ultramafic-gabbroic bodies lie in a zone between, on the east, a terrane of metamorphosed Proterozoic sediments representative of the Pre-Mesozoic craton, and on the west, the projected strike line of the Stuart Lake Belt of Cache Creek Group strata and alpine-type peridotite, rocks that evidently constitute Permian oceanic crust (Fig. 15.2). Thus some relationship to a suture zone is suggested (Irvine, 1975). The bodies emplaced in Takla rocks are also contained in a fault-bounded, graben-like structure that is flanked on opposite sides by contrasting Upper Paleozoic strata of similar age that generally dip away from it. This structure is complicated by other faults not shown in Figure 15.2 (Takla Project, reports 10-16) and by granitic plutonism, but on a broad scale it appears to be continuous with the Quesnel Trough in south-central British Columbia (described by Campbell and Tipper, 1970), which similarly is largely filled with Upper Triassic basic volcanics and which extends southward to the area of the Tulameen ultramafic-gabbroic complex.

No mineral deposits of economic interest have been found in association with the ultramafic and gabbroic rocks. The little asbestos form serpentine that has been seen in the bodies is coarse and harsh; chromite concentrations are few and very small, and the chromite characteristically has high Fe/Cr ratios; sulphides are rare; and the visible zones of magnetite-bearing pyroxenite are too small and low grade to warrant consideration as sources of iron. The only apparent possibility of interest concerns the platinum group metals. At Tulameen and in the Ural Mountains, Alaskan-type ultramafic bodies have been the source of significant placer deposits of platinum, and platinum has been reported in association with placer gold in the McConnell Creek area (Lord, 1948). Exploration of the stream gravels around the ultramafic bodies, therefore, might be worthwhile.

References

Campbell, R.B. and Tipper, H.W.
1970: Geology and mineral exploration potential of the Quesnel Trough, B.C.; Can. Min. Metall., Bull., v. 63, p. 174-179.

Gabrielse, H. and Dodds, C.J.
1974: Operation Finlay; *in* Report of Activities, Part A, Geol. Surv. Can., Paper 74-1A, p. 13-15.

Irvine, T.N.
1974a: Ultramafic and gabbroic rocks in the Aiken Lake and McConnell Creek map-areas, British Columbia; *in* Report of Activities, Part A, Geol. Surv. Can., Paper 74-1A, p. 149-152.

1974b: Petrology of the Duke Island ultramafic complex, southeastern Alaska; Geol. Soc. Am., Mem. 138.

1975: Polaris complex and Axelgold intrusion, contrasting ultramafic and gabbroic bodies in the Omineca Mountains, British Columbia; Geol. Assoc. Can., Cordilleran Sect., Programme and Abstracts, p. 18-19.

Lord, C.S.
1948: McConnell Creek map-are, British Columbia; Geol. Surv. Can., Mem. 251, 72 p.

Roots, E.F.
1954: Geology and mineral deposits of the Aiken Lake map-area, British Columbia; Geol. Surv. Can., Mem. 274, 246 p.

Taylor, H.P., Jr.
1967: The zoned ultramafic complexes of southeastern Alaska; *in* Ultramafic and Related Rocks, P.J. Wyllie, ed., J. Wiley and Sons, New York, p. 97-121.

16. STRUCTURE AND METAMORPHISM IN SOUTHERN SWANNELL RANGES, BRITISH COLUMBIA

Project 750016

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Introduction

At the northern end of the Wolverine complex in the Omineca Mountains metamorphic isograds and gross anticlinal structure appear to plunge northerly into a terrane of less metamorphosed rocks of the Ingenika and Tenakihi groups. The goal of this project was a detailed study of structure, metamorphism, and stratigraphy in this transition zone.

Field work was carried out for a M. Sc. degree at the University of British Columbia; logistical support was provided by the Geological Survey of Canada as part of the Takla project.

During reconnaissance studies of the Aiken Lake map-area Roots (1954) recognized large-scale structure in the Swannell and Finlay ranges as delineated by thick mappable limestone, quartzite, and phyllite units which are part of the Tenakihi and Ingenika groups of late Precambrian and, in places, lower Cambrian age. Mansy (1972, 1974) has continued studies in the Ingenika Group in these ranges, describing large-scale southwest-verging folds and thrusts which appear to be continuations to the northwest of structures mapped by Roots.

Higher grade rocks near Chase Mountain (characterized by the presence of kyanite and sillimanite) display at least two phases of earlier intense tight to isoclinal folding which are in turn refolded by northwest trending structures correlated with those described by Roots (1954) and Mansy (1972, 1974). At least one generation of brittle structures occurred after this late large-scale warping. Within the area studied earlier high-temperature assemblages have undergone retrograde metamorphism.

Stratigraphy

Near Chase Mountain the rocks of the Tenakihi Group consist of quartz-muscovite-biotite±feldspar ±garnet schist, impure muscovite-biotite quartzite, feldspathic quartzite, and quartz-feldspar-mica gneiss with very rare, thin, and discontinuous layers of marble, conglomerate and amphibolite. The major rock types are mostly thin- to medium-layered and of uneven thickness. Due to boudinage, repeated folding, and possibly discontinuous inherited depositional characteristics, it is impossible to trace the thicker individual layers for more than several hundred metres. Aluminous schists are rare to absent, and provide limited control of metamorphic grade. The rocks could not be subdivided stratigraphically and consequently there is little control on large-scale effects of the earlier deformations. Repeated folding and obvious local repetition of strata render any attempt to construct stratigraphic sections of dubious value.

Phase 1

Figure 16. 1 contains sketches of phase 1 folds (F1); attitudes of compositional layering, axial planes and fold axes of F1 folds, several zones of detachment, and major antiformal trace modified from Roots (1954). F1 folds have axial plane schistosity which is perpendicular to layering in fold hinges; they are mostly similar in style and isoclinal in form, and are commonly rootless due to shear along parallel limbs. These folds have a very flattened appearance resulting from renewed closure by phase 2 (F2) deformation. The axes of these folds trend and plunge north to north-northwest on the northeast limb of the major antiform, and south-southeast to southwest on the southwest limb. F1 folding has resulted in the transposition of bedding into the axial plane, so that the two are subparallel. Thus F1 axial planes define the antiform in the same sense as the compositional layering. Most outcrops display nearly planar layering with subparallel schistosity. The layering is discontinuous, commonly spectacularly boudinaged, and generally appears extremely flattened. A few F1 hinges are found within the secondary foliation.

Phase 2

Figure 16. 2 contains sketches of F2 structures, F2 axial plane and fold axis symbols, and representative attitudes of compositional layering. Zones of more intense structural dislocation (tectonic slides) and recrystallized mylonitic rocks are also plotted. F2 folds are distinguished from F1 structures by the presence of a folded and crenulated schistosity in the cores of folds, and by their tight (as opposed to strictly isoclinal) character with rounded hinges. F2 folds are dominantly flexural-flow folds that fold the transposed layering. Transposition by F2 of both layering and schistosity in places forms a second schistosity but in general is not complete and, as a consequence, a northeastward sense of vergence is consistently recognizable across the area. Many exposures (Fig. 16. 2a, 16. 2c) illustrate these folds 'cascading' to the east where the compositional layering is clearly distinguishable. F2 folds are coaxial with F1 folds but their axial planes are usually distinct. This later deformation results in the tightening of F1 folds and the production of tectonic slides on various scales parallel with both F1 and F2 axial planes. The section a-a' (Fig. 16. 2) near Chase Mountain illustrates the progressive transposition of the layering into the F2 axial plane along zones of intense

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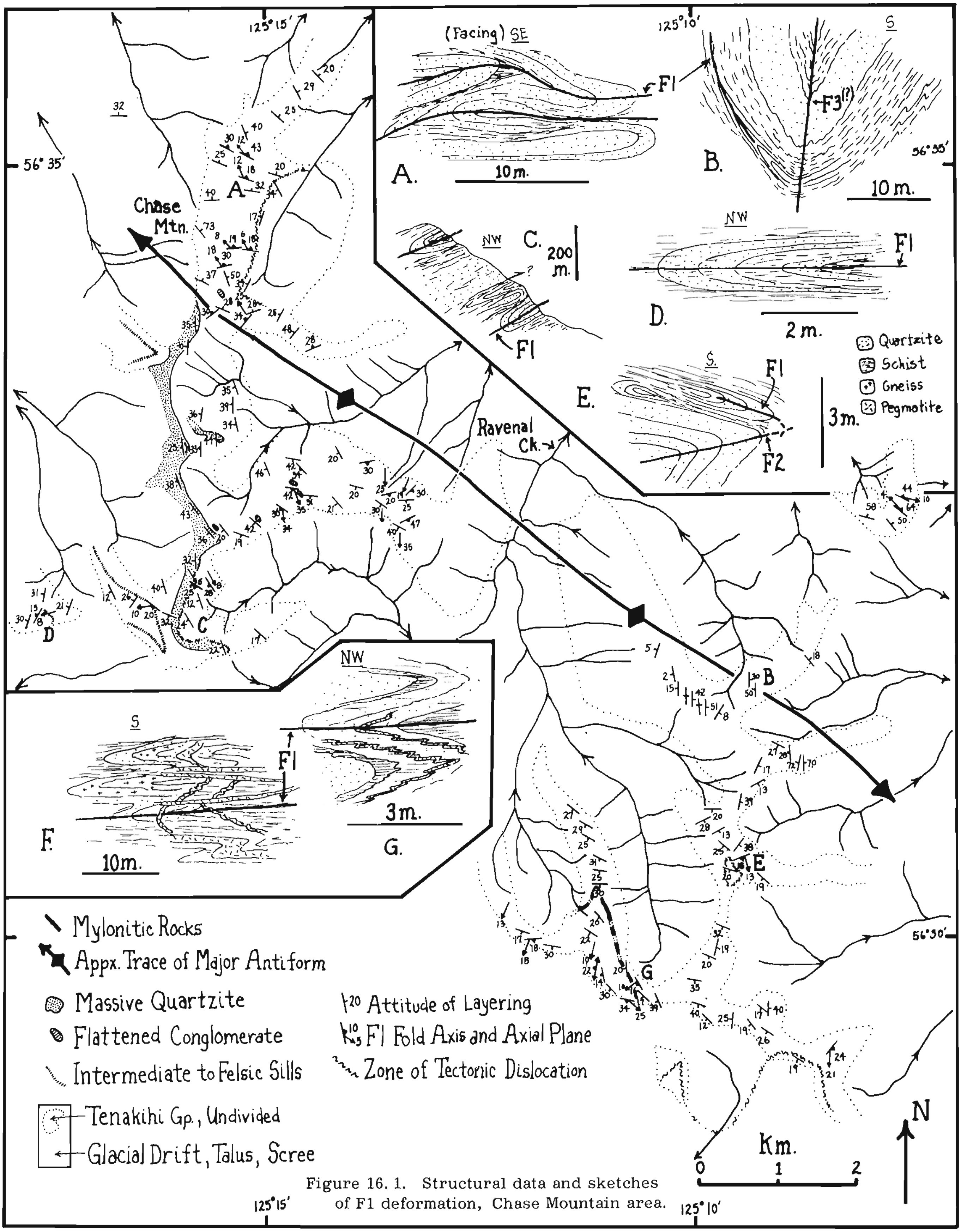

Figure 16. 1. Structural data and sketches of F1 deformation, Chase Mountain area.

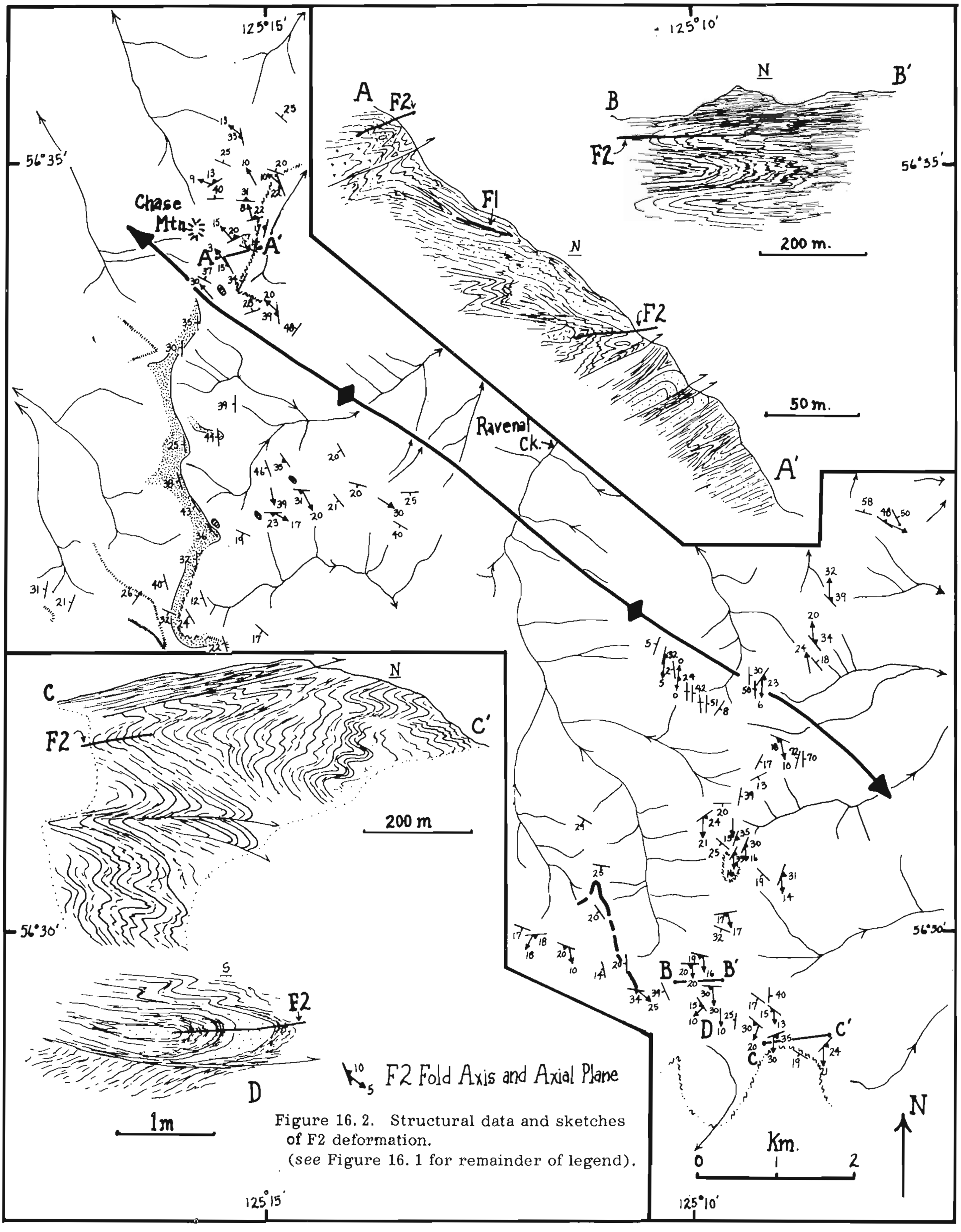

Figure 16.2. Structural data and sketches of F2 deformation.
(see Figure 16.1 for remainder of legend).

strain and flattening, producing mylonitic fabrics in rocks near the base of the cliff. The detachment zones mapped are thought to be F2 structures. Their displacements are not known but are not thought to be great.

Post-Phase 2 Regional Antiform

The major antiformal trace (Figs. 16.1, 16.2) was recognized by Roots (1954). However, rather than a fold defined by attitudes of bedding, its position is defined by the attitudes of compositional (transposed) layering and F1 axial planes. Judging from respective dips of layering on opposite limbs, this antiformal trace is northwest-trending, upright, and open in style plunging both northwest and southeast producing an elongate dome or culmination. Minor structures reflecting the geometry of this larger structure are rare or absent. Fhis structure is probably correlative to those present to the northwest in Swannell Ranges as described by Mansy (1972, 1974).

Other Post-Phase 2 Structures

North trending, upright to steeply eastward inclined, normal to tight folds are present near B in Figure 16.1. Several of these large folds are *en echelon* near the axial trace of the major antiform. The wavelength and amplitude of these folds can be up to several hundred metres. This north-trending fold system with axial planes varying from vertical to steeply inclined to the east could be interpreted to have interfered with the major antiform to produce the elongate dome or culmination in the area. These are post-F2, but their age with respect to the major antiform is indeterminate. The folds might conceivably have been produced by a room problem in the core of the major antiform, reactivating older structural fabric.

Other evidence of later deformation is expressed in mesoscopic folds and warps, kink and crenulation structures, and brittle fractures; most of these are steeply dipping, but have variable strike. Few convincingly consistent attitudes in these later structures were found.

Metamorphism

Phases 1 and 2 appear to have been accompanied by medium to high grade metamorphism. Sillimanite is observed locally but migmatites are lacking so that the reaction:

muscovite+quartz+plagioclase ⇄ K-feldspar +plagioclase+quartz components of anatectic melt + an-richer plagioclase+sillimanite+H_2O

probably did not occur. There is no difference in metamorphic grade or structural style between rocks mapped by Roots (1954) as Wolverine complex and adjacent rocks mapped as 'normal' Tenakihi Group. Quartz-K feldspar-plagioclase-muscovite pegmatites are both synkinematic and postkinematic with (respect to F2. Retrograde metamorphism has been extensive, altering nearly all garnet to chlorite, but leaving biotite largely intact. The role of metamorphism with respect to post-F2 events is not yet clear.

Metamorphic and granitic rocks immediately northeast of Blackpine Lake (21-25 km south-southeast of Chase Mtn.: 125°20', 56°22') were examined to determine when these igneous rocks were emplaced in the structural sequence. The structural style and orientation there is similar to that described above, and it was found that several generations of quartz-muscovite-feldspar+garnet pegmatites and quartz-feldspar-muscovite±garnet±biotite granitic sills, dykes, and small stocks are present. Essentially all of the smaller igneous bodies were emplaced pre- or syn-kinematically with respect to F1 and F2; nearly all bear some deformational fabric of F2, and some are intricately folded (Fig. 16.1F, 16.1G). Some of the metamorphic host rocks have conspicuous sillimanite and rarely sillimanite pseudomorphs after kyanite. Retrograde metamorphism is much less important in these rocks. The larger granitic body (quartz-feldspar-biotite-granodiorite) and several small cross-cutting dikes can be interpreted to have been emplaced post-tectonically, but a faint foliation is sometimes detected. The nature of its contact with adjacent metamorphic rocks is difficult to determine in detail due to abundant sills and dikes of synkinematic phases which render the contact gradational.

The absolute ages of these structures is not known. K/Ar dating on rocks of the Wolverine complex typically yields 45-50 m.y. (Muller, *in* Wanless *et al.*, 1972). Rb/Sr and K/Ar dating will be carried out at the University of British Columbia in hopes of providing ages of synkinematic granitic phases and thus maximum ages for F2. The K/Ar dates probably reflect uplift and erosion or a thermal event during the early Tertiary, and this may have obscured the record of older events.

References

Mansy, J.L.
1972: Stratigraphy and structure of the Ingenika Group in Finlay and Swannell Ranges (94 E, F); *in* Report of Activities, Part A, Geol. Surv. Can., Paper 72-1A, p. 29-32.

1974: Operation Finlay; *in* Report of Activities, Part A, Geol. Surv. Can., Paper 74-1A, p. 17-18.

Roots, E.F.
1954: Geology and mineral deposits of Aiken Lake map-area, British Columbia; Geol. Surv. Can., Mem. 274.

Wanless, R.K., Stevens, R.D., Lachance, G.R., and Delabio, R.N.
1972: Age determinations and geological studies, K-Ar isotopic ages, Report 10; Geol. Surv. Can., Paper 71-2, p. 11-12, 20-27.

OPERATION FINLAY (Reports 17-19)

Project 700047

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During the 1975 field season, geological reconnaissance was nearly completed in Toodoggone River (94 E) map-area. Figure 17.1 shows the distribution of the main rock units and some of the major faults. The most important results arising from the summer's field activities are as follows:

1. A large area in northwestern Toodoggone River map-area bounded by Lunar Creek on the east and Stikine River on the south is underlain by a sequence, possibly more than 1000 m thick, of low grade metamorphic rocks comprising chloritic phyllite, tan-brown sericitic phyllite, sheared dark green volcanics, sheared tuff and crystalline limestone. The rocks are cut by granitic bodies of probable Early Jurassic age and apparently are overlain by Lower Jurassic volcanic conglomerate. Their age is tentatively considered to be late Paleozoic. Near the headwaters of Lunar Creek they are the host rock to an ultramafic complex (Irvine, 1976).
2. A batholith of megacrystic biotite hornblende granodiorite or quartz monzonite in the northwest part of the map-area is similar in composition to the batholith northeast of Sturdee and Firesteel rivers.
3. Foliated, medium to coarse grained hornblende-quartz diorite forms a batholith underlying much of the Peak Range and is separated from the larger, lithologically similar Pitman Batholith to the north by a southeast trending screen of metasedimentary and metavolcanic rocks northeast of Chukachida Lake. The Pitman Batholith in the north-central part of the map-area is bounded to the east by a batholith of foliated megacrystic quartz monzonite underlying most of the Thudaka Range. The quartz monzonite is mylonitized along its borders and is sheared for distances of more than a mile inward from the contacts.
4. Leucocratic, pink weathering, granitic rocks around Mount Albert Dease, west of Chukachida Lake and elsewhere in the central part of the map-area are cut by basaltic dyke swarms. The granitic rocks appear to be high level intrusions and are closely related to numerous large feldspar porphyry dykes. The latter are commonly spatially related to gossan zones.
5. Highly sheared ultrabasic bodies were noted in hornblende quartz diorite west of Geese Creek and in amphibolite west of Thudaka Creek.
6. Lower Cambrian and Ordovician strata of the Atan and Kechika groups are exposed in Ruby Range. North of Ridgeway Lake, these strata form the limbs of a major southerly plunging anticlinorium with late Proterozoic rocks in the core. Limestone and pure quartzite in Sifton Range are believed to be Lower Cambrian in age. If so, the grade of metamorphism affecting Lower Cambrian strata increases progressively from essentially unmetamorphosed rocks in the Finlay Ranges, to chlorite grade in Ruby Range and northwesternmost Kechika Ranges to kyanite grade in Sifton Range.
7. Numerous major faults are well documented and are commonly marked by intense zones of dislocation and topographic depressions. Of fundamental importance is the fault that borders the Thudaka quartz monzonite on the west and continues southerly along Finlay River, where it delimits the western extent of Proterozoic rocks on the southwest side of the Swannell Ranges. This fault can be traced northerly in Kechika map-area (Gabrielse, 1962), where it separates Thudaka quartz monzonite from the foliated hornblende quartz diorite of the Pitman Batholith and, farther north, upper Proterozoic rocks to the east from upper Paleozoic strata to the west. To the south, the fault can be traced through northeastern McConnell Creek map-area (Richards, 1976) and southwestern Aiken Lake map-area (Roots, 1954). A steep fault, east side down, runs along the eastern side of Thudaka Range and Swannell Ranges. Three northwest trending faults in the central part of the map-area, all east side down, cause repetition of sequences that commonly dip southwesterly. A conspicuous fault detected by a series of depressions, a zone of shearing in bedrock and a sharp contrast in vegetation in drift-covered areas trends easterly from the western boundary of the map-area at latitude 57°48'N to Lunar Creek.
8. Spectacular, essentially westward verging, refolded isoclinal folds involving a sequence of semi-pelitic schist, pure quartzitic marble, and amphibolite are exposed in Sifton Range. At the southeast end of the range, just northwest of Fox Pass, an equigranular medium grained hornblende biotite granitic body intrudes augen gneiss and amphibolite of the above sequence.
9. Copper minerals are present in many localities but are particularly abundant in or near coarse-bladed feldspar porphyries of the Upper Triassic volcanic sequence in a northwest trending block including Mt. McNamara, Claw Mountain, Oxide Peak and Contact Peak. Chalcopyrite occurs in a vein on the south side of the rugged peak 1.6 km north of Harmon Peak. Augite porphyry and granitic rocks in the fault zone west of Swannell Ranges contain copper minerals in several localities from Giegrich Peak north to near Reef Canyon. Fault zones with quartz veins in strata of the Kechika Group in Ruby Range and in northwestern Kechika Ranges contain copper minerals locally. If, as suggested above, Sifton Range is in part underlain by Lower Cambrian strata, the possibility of lead-zinc occurrences should not be overlooked.

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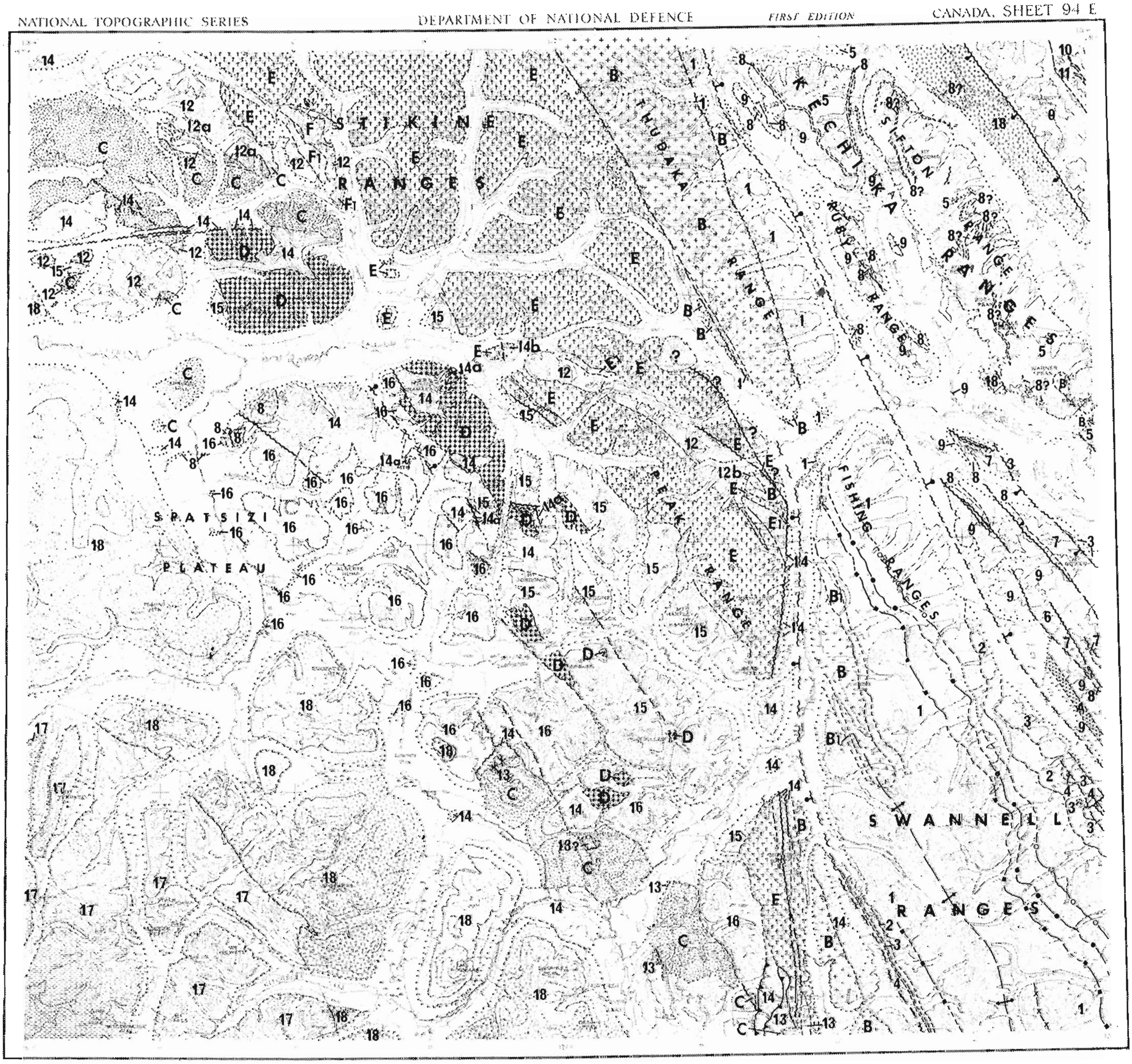
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SPATSIZI PLATEAU
SWANNELL RANGES

Figure 17.1

LEGEND FOR FIGURE 17.1

TERTIARY AND UPPER CRETACEOUS

18	SUSTUT GROUP: SIFTON FORMATION	Nonmarine conglomerate, shale, siltstone, tuff, minor fetid limestone. Gabbroic dykes and sills.

MIDDLE AND UPPER JURASSIC

17	'BOWSER ASSEMBLAGE'	Shale, siltstone, pebble conglomerate.

LOWER AND/OR MIDDLE JURASSIC

16	'TOODOGGONE' volcanic rocks	Dacite, latite, rhyolite, tuff, breccia, flows; local maroon weathering conglomerate of uncertain age; includes local intrusive equivalents.

LOWER JURASSIC ?

15	HAZELTON GROUP	Volcanic conglomerate, breccia, lahar; abundant pink feldspar porphyry dykes and sills; may include some 14 and 16.

UPPER TRIASSIC

14	TAKLA GROUP	Coarse-bladed plagioclase porphyry, augite porphyry, tuff, agglomerate; 14a, limestone; may locally include 15.

UPPER PALEOZOIC

13	ASITKA GROUP	Chert, argillite, limestone, greenstone.
12		Sericite and chlorite phyllite, foliated, chloritic greenstone, grit; acidic tuff (?), minor red chert; 12a, limestone; 12b, chlorite schist, grit, amphibolite.

DEVONIAN AND MISSISSIPPIAN (?)

11		Conglomerate, siltstone, shale.

ORDOVICIAN, SILURIAN AND DEVONIAN

10		Siltstone, shale, calcareous shale.

CAMBRIAN AND ORDOVICIAN

9	KECHIKA GROUP	Phyllitic limestone, calcareous shale, limestone.

LOWER CAMBRIAN

8	ATAN GROUP	Limestone, siltstone, dolomite.
7		Impure quartzite, shale; local sandstone conglomerate.
6		Orthoquartzite.

PROTEROZOIC AND LOWER CAMBRIAN (UNDIVIDED)

5		Mica schist and phyllite, quartzite, amphibolite, augen-gneiss, crystalline limestone.

PROTEROZOIC

4		Siltstone, sandstone, shale, limestone.
3		Limestone (locally oolitic and pisolitic), minor dolostone.
2		Sericitic phyllite.
1		Quartzo-feldspathic, gritty sandstone, siltstone, shale and conglomerate; minor limestone; metamorphic equivalents from chlorite to kyanite grade.

GRANITIC ROCKS

TERTIARY

A	Dacite (?) dyke

MID-CRETACEOUS

B	Quartz monzonite, mainly foliated, mylonitized along contacts; B_1, gneiss and migmatite.

LOWER JURASSIC

C	Quartz monzonite and granodiorite, locally megacrystic.
D	Granodiorite, leucocratic, pink; fine- to medium-grained.
E	Hornblende-quartz diorite, commonly contains biotite; foliated; E_1, migmatite.

ULTRABASIC ROCKS

F	Dunite, clinopyroxenite; peridotite; F_1, hornblende, gabbro.

- - - - - - - -	geological boundary	—o—o—	chlorite isograd
. . . .	limit of geological mapping	—●—●—	biotite isograd
~~ ~~ ~~	fault	—■—■—	garnet isograd
—▲—▲—▲—	thrust fault	—x—x—	kyanite isograd

References

Gabrielse, H.
1962: Kechika, B.C.; Geol. Surv. Can., Map 42-1962.

Irvine, T.N.
1976: Studies of Cordilleran gabbroic and ultramafic intrusions, British Columbia; *in* Report of Activities, Pt. A, Geol. Surv. Can., Paper 76-1, rept. 15, pt. 2.

Richards, T.A.
1976: Takla Project: McConnell Creek map-area (94 D east half); *in* Report of Activities, Pt. A, Geol. Surv. Can., Paper 76-1A, rept. 10.

Roots, E.F.
1954: Geology and mineral deposits of Aiken Lake map-area, British Columbia; Geol. Surv. Can., Mem. 274.

18. STRATIGRAPHY, STRUCTURE AND METAMORPHISM IN NORTHERN AND CENTRAL SWANNELL RANGES

Project 700047

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Swannell Ranges, enclosing an area outlined by Pelly Creek in the east, Swannell River in the south, Fredrikson and LaForce creeks in the west and the bend of Finlay River in the north, consist entirely of Proterozoic rocks disposed in a broad asymmetrical anticlinorium (Fig. 18.1). A monotonous formation of clastic rocks (unit 1)[2] comprising quartzo-feldspathic gritty sandstone, siltstone, shale and minor conglomerate and limestone (with metamorphic equivalents up to at least kyanite grade) is by far the most widespread lithology. Overlying this unit and outcropping on the extreme flanks of the ranges is a sequence of sericitic and calcareous phyllite, commonly pyritiferous (unit 2), overlain by a limestone (locally oolitic and pisolitic) with minor dolostone (unit 3). Locally, especially within Fort Grahame map-area, west half (Espee Range and Forres Mt. area), unit 3 is overlain by a succession of siltstone, sandstone, shale and limestone (unit 4) and this by a resistant, generally pure white orthoquartzite (unit 5).

The twofold subdivision of the lower part of the Proterozoic assemblage (Lower Ingenika Group and Tenakihi Group) as proposed by Roots (1954) for the Aiken Lake map-area, has been found difficult to apply. The formations are very similar lithologically; although the Tenakihi Group is perhaps slightly more pelitic. The contact between them, however, appears to correspond to a significant change in structural style (from relatively undeformed strata to the onset of flow folding) and occurs at or just above the garnet isograd.

Structurally the north and central Swannell Ranges are occupied by a broad asymmetrical anticlinorium, the vergence of folds being consistently towards the southwest. Culminations and depressions in the anticlinorium are suggested by mesascopic elements (plunges of lineations and minor fold axes). To the north the axis of the major structure is truncated by faulting in the Fishing Lakes and Reef Canyon region of Finlay River. To the south near Mt. Lay, the axis plunges southeast exposing carbonates of the upper part of the succession.

On the east flanks, rocks of the upper part of the sequence (chlorite to subchlorite grade) are moderately to strongly folded and display a fold style comparable to that in Russel Range described by Mansy (1972). Folds are southwest verging, upright to slightly overturned and generally similar in style. The fold pattern is further complicated by an array of normal, high angle reverse, and some southwest directed thrust faults. Farther west (within biotite to garnet grade), folding is generally restricted to very minor flexures and warps within a generally northeast dipping homocline. Within the axial region of the anticlinorium, however, in rocks of garnet through staurolite/kyanite grade, the fold style changes dramatically. Folds are tightly isoclinal and fold hinges appear dislocated by bedding-plane slip and thrusting. Tracing of individual beds within this zone and hence the deciphering of megascopic structure is exceedingly difficult because of the monotony of stratigraphy and lack of any distinct marker beds. Good examples of tight isoclinal packages occur in the Mt. Lay area and in the general area bordering Flameau Creek. The west flank of the anticlinorium appears to be more complex and is less understood. Tight isoclinal folds, the earliest fold episode recognized, are domed by the major antiformal structure and are reoriented with axial planes dipping steeply to the southwest. Tight folding persists within the higher grade part. Folds, however, become more open on decrease in metamorphic grade. The pattern and style of these folds are much complicated by faulting on the western flank of the major structure.

Regional metamorphism of Barrovian type has occurred throughout the north and central Swannell Ranges, reaching at least kyanite grade. Chlorite, biotite and garnet isograds can be readily delineated. The isogradic distribution appears regular and fairly widely spaced for the entire length of the eastern flank of the ranges, wheras that on the western flank is more tightly spaced and is complicated by faulting. A widespread retrograde metamorphism, however, has erased much of the evidence of the earliest regional metamorphism being particularly apparent in the axial region where higher grade assemblages are only sporadically preserved. A second phase of metamorphism has produced a new generation of chlorite, biotite, garnet and albite, which cross-cut earlier fabrics. Deep wine-red garnet of the first generation is xenoblastic to subidioblastic, up to 2 cm in diameter, and exhibits all degrees of retrograde alteration from chloritic selvages to almost complete pseudomorphic replacement by chlorite and white mica. Garnet of the second generation is much smaller, generally pale pink, commonly idioblastic and generally fresh. Biotite of the first phase is commonly fine grained, whereas that of the second is coarse grained (up to 1 cm). Locally, spotted coarse chalk-white porphyroblasts of albite occur within the axial region of the major structure. In places, albite appears to have formed in contact metamorphic aureoles near granitic sills or dykes.

There is a complete lack of "Wolverine type" muscovite tourmaline pegmatites throughout the northern part of the Swannell Ranges. Pegmatites are apparently restricted to the somewhat deep zones of the belt such as that to the southeast in the southern Swannell Ranges

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[2]Map-units refer to Figure 17.1 of Gabrielse *et al.* (1976).

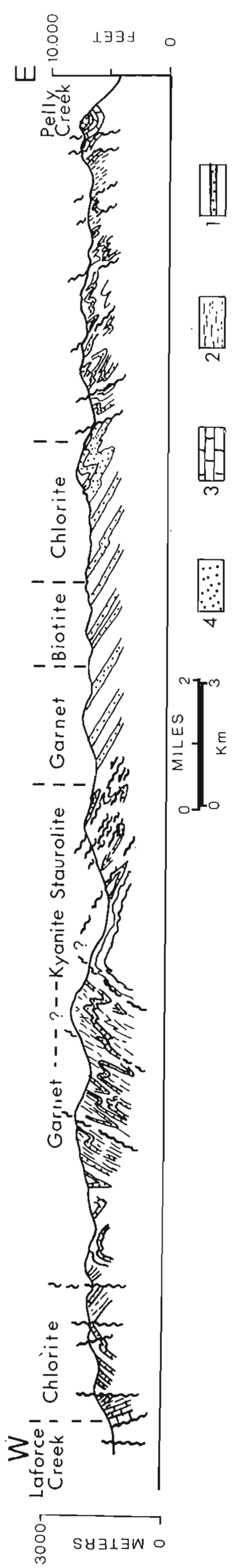

Figure 18.1. Cross-section, central Swannell Ranges (*see* Figure 17.1, Gabrielse *et al.* (1976) for legend).

(Parrish, 1976). Near Whudzi Mountain and occupying the axial part of the anticlinorium is an elongate megacrystic muscovite-biotite-quartz monzonite body which has a lit-par-lit injection border zone. Much of this body is strongly foliated, but it is cut by a haphazard network of veins of undeformed equigranular quartz monzonite and aplite, which are commonly garnet bearing. The Whudzi body is compositionally and texturally identical to that in the Thudaka Range to the north. Radiometric K-Ar dates indicate a mid-Cretaceous emplacement of the Thudaka and Whudzi bodies, as does a single determination on biotite from axial zone of the anticlinorium.

A much simplified structural and metamorphic history of the region is as follows:

(1) regional metamorphism to at least kyanite grade with development of first phase isoclinal folds.
(2) extensive retrograde metamorphism either pre- or during the early stages of updoming, giving rise to the major anticlinorium.
(3) intrusion of Whudzi body conceivably sometime during development of the major anticlinorium, with renewed heat flow to give rise to the second phase of metamorphism.

References

Mansy, J. L.
1972: Stratigraphy and structure of the Ingenika Group in Finlay and Swannell Ranges; *in* Report of Activities, Part A, Geol. Surv. Can., Paper 72-1A, p. 29-32.

Parrish, R.
1976: Structure and metamorphism in southern Swannell Ranges; *in* Report of Activities, Part A, Geol. Surv. Can., Paper 76-1A, rept. 16.

Roots, E. F.
1954: Geology and mineral deposits of Aiken Lake map-area, British Columbia; Geol. Surv. Can., Mem. 274.

Project 700047

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Two sections (A and B) were measured in Cambrian strata in north-central Ware map-area, west half (lat. $57^{\circ}57\frac{1}{2}'$, long. $125^{\circ}31'$ and lat. $57^{\circ}56\frac{1}{2}'$, long. $125^{\circ}30'$; Fig. 19.1). Lateral facies variations are evident in the two localities and are particularly marked in comparison with strata described by Fritz (1972) in the Mount Lloyd George area about 26 km (16 mi.) to the east of sections A and B. In general the abundance of clean sandstone decreases, and the abundance of shale increases westerly.

Between South Gataga River and Weissener Creek the lowermost beds which are more than 200 m (650 ft.) thick are mainly dark green and grey shales and siltstones interbedded with lenticular bodies of limestone containing numerous archeocyathids. The largest lenses of limestone are as much as 30 to 40 m (98-130 ft.) long and about 10 m (33 ft.) thick. This unit is in sharp contact with overlying white medium to coarse grained quartzitic sandstone forming a resistant member 50 m (160 ft.) thick. The sandstone contains fragments of trilobites, and locally, well developed tubes perpendicular to bedding. Graded beds and tabular cross-beds are conspicuous and the uppermost bed displays well developed ripple marks.

Minor brown shale and quartzite marks the transition between the quartzite member and thick bedded to massive carbonate. The basal carbonate sequence comprises dolostone and minor limestone with red, argillaceous laminae. Pisolites and oncolites are common in places. In places thin beds of white dolostone alternated with blue-grey weathering limestone. Interbedded with the carbonate are beds of quartzitic sandstone and fine grained conglomerate. In Section A they form two prominent units near the centre of the carbonate member. The uppermost beds of the carbonate member are cherty limestones and these underlie a thick sequence of calcareous shale and wavy banded, silty limestone of presumed post-Early Cambrian age. Total thickness of the carbonate member is about 350 m (975 ft.).

Sedimentary structures suggest that strata in the measured sections are mainly of shallow water origin. Thicknesses and facies of equivalent formations west of Northern Rocky Mountain Trench are considerably different (Mansy, 1972, in press). There the member containing archeocyathids is almost entirely carbonate and the overlying beds are mainly clastic.

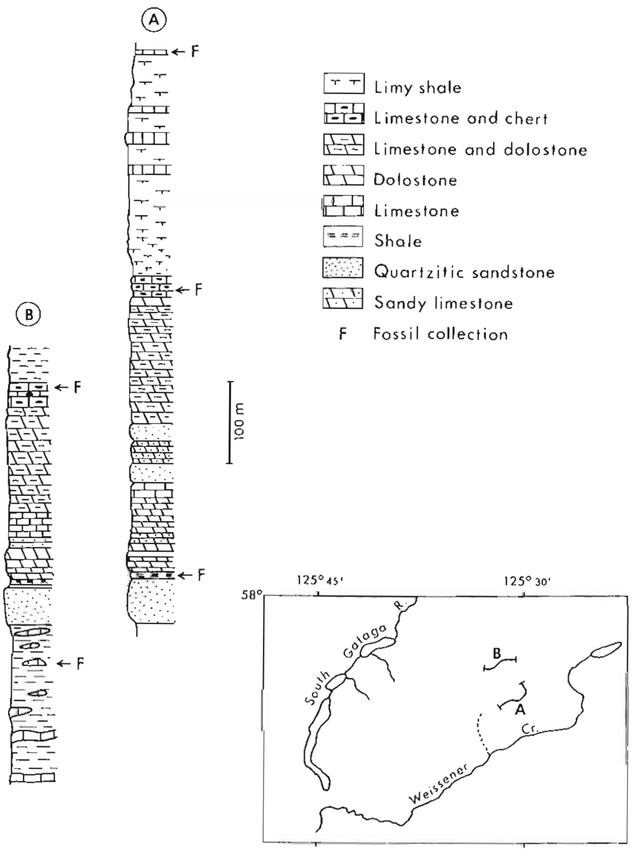

Figure 19.1. Cambrian stratigraphic sections, Ware W½ map-area.

References

Fritz, W.H.
1972: Cambrian biostratigraphy Western Rocky Mountains, British Columbia (83 E, 94 C, 94 F); *in* Report of Activities, Part A, Geol. Surv. Can., Paper 72-1A, p. 208-211.

Mansy, J.L.
1972: Stratigraphy and structure of the Ingenika Group in Finlay and Swannell Ranges (94 E, F); *in* Report of Activities, Part A, Geol. Surv. Can., Paper 72-1A, p. 29-32.

Etude sedimentologique et essai de reconstitution paleogeographique des formations Precambriennes et Cambriennes des Monts omineca, cassiar et cariboo situes sur le bord occident du fosse des Montagnes Rocheuses (Colombie Britannique, Canada); Ann. Societe Geol. du Nord (Lille). (in press)

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LARDEAU MAP-AREA (82 K WEST HALF), BRITISH COLUMBIA

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Four weeks of field work in 1975 were devoted to: (1) relocation of boundaries between Lardeau, Milford and Kaslo groups in the area west of Trout Lake to the Northeast Arm of Upper Arrow Lake; (2) continued investigation of the structure and stratigraphy of the Lardeau Group northwest of Incomappleux River; and (3) resolution of correlation problems arising from detailed mapping of Fyles and Eastwood (1962) and regional mapping of Read and Wheeler (1975).

West of Trout Lake City, logging roadcuts expose a basal conglomerate of the Milford Group which contains clasts of the underlying Lardeau Group with the earliest foliation varying in orientation from clast to clast (20.2). Southwest of this conglomerate, grey phyllite and other rocks of the Lardeau Group form the core of a second phase antiform (Fig. 20.1). The area underlain by the Lardeau Group is larger than shown by Read and Wheeler (1975), but its southwestern limit has not

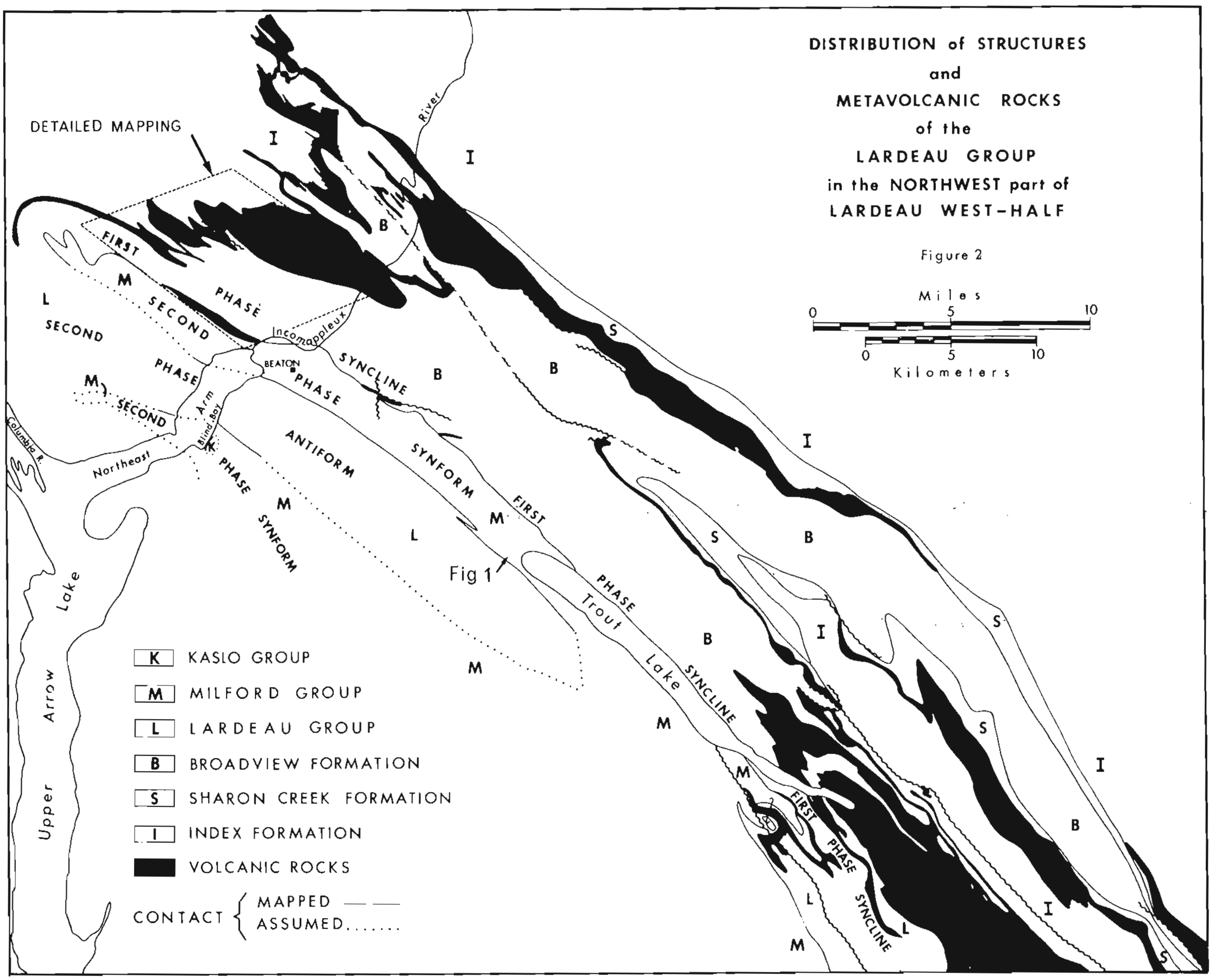

Figure 20.1. Distribution of structures and metavolcanic rocks of the Lardeau Group in the northwest part of Lardeau West-Half.

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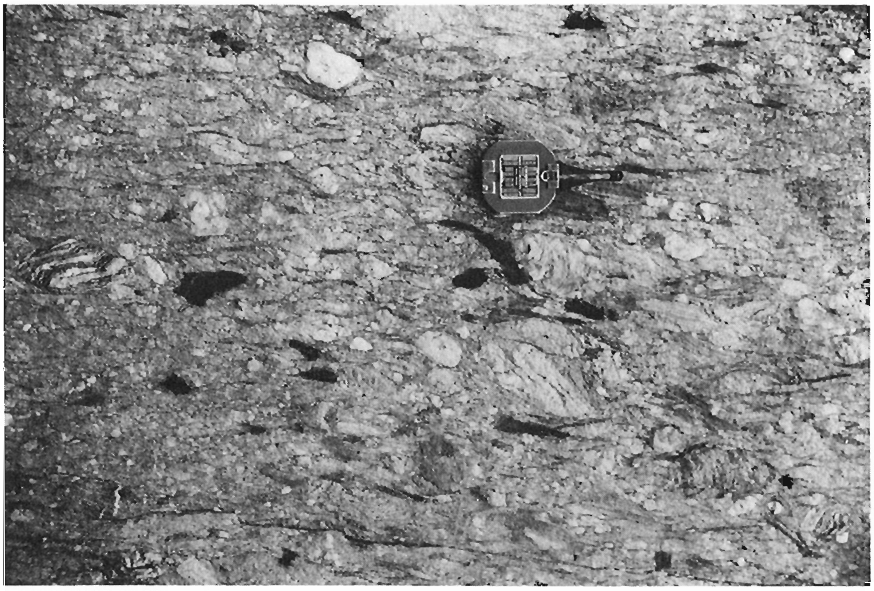

Figure 20. 2

Conglomerate at the base of the Milford Group near the north end of Trout Lake (Fig. 13. 2). Clasts with a randomly oriented first phase foliation lie in a foliated (second phase) matrix oriented parallel to the compass.

been relocated. In the core of this antiform many of the metavolcanic and limestone units previously assigned to the Milford Group more likely belong to the Lardeau Group. At Blind Bay metavolcanic rocks belong to the Kaslo Group.

Mapping of metavolcanic and adjacent rocks northwest of the Northeast Arm of Upper Arrow Lake yields a distribution of metavolcanic rocks and mesoscopic structures implying that a first phase syncline lies just northeast of the Milford Group and extends throughout Lardeau West-Half map-area. Metavolcanic rocks, locally preserved under the angular unconformity at the base of the Milford Group, define the southwest limb of the first phase syncline. Southwesterly towards Columbia River Valley, the southwest limb of the syncline may expose the Index and Badshot Formations and Hamill Group.

This summer's work has been incorporated into Open File 288 which also includes the Mineral Deposit map for all of Lardeau map-area (west half).

References

Fyles, J. T. and Eastwood, G. E. P.
1962: Geology of the Ferguson area, Lardeau District, British Columbia; B. C. Dept. Mines Pet. Resour., Bull. 45.

Read, P. B. and Wheeler, J. O.
1975: Lardeau west-half geology; Geol. Surv. Can., Open File 288.

21. STRATIGRAPHIC AND STRUCTURAL STUDIES IN THE PELLY MOUNTAINS, YUKON TERRITORY

Project 730037

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Introduction

Quiet Lake (105F) and Finlayson Lake (105G) map-areas are of particular interest because they straddle Tintina Trench and lie close to the northwestern extremity of the Omineca Crystalline Belt, where it merges with the Yukon Crystalline Terrane. Field work in the 1975 field season continued stratigraphic and structural studies of the northern extremity of the Omineca Crystalline Belt begun in 1973. Tempelman-Kluit completed regional mapping of Quiet Lake map-area and part of Finlayson Lake map-area and this report is a summary of the main findings of that work. Read completed field work in a study of Lower Cambrian strata near Ketza River which will form the basis of a master's thesis at the University of Calgary; principal findings of this work were published earlier (Tempelman-Kluit *et al.*, 1975). Read also began a detailed study of the Silurian and Devonian platform-like succession of the Pelly Mountains and a summary of that work is contained herein. Gordey studied the structural style south of Tintina Trench in Finlayson Lake map-area, this will form the basis for his doctoral dissertation at Queen's University.

General Setting

Two contrasting sequences of stratified rocks are exposed in the Pelly Mountains. On the northeast near Tintina Trench a succession of shale, sandstone and carbonate rocks was deposited during shallow marine conditions near the margin of the stable continental platform. This miogeoclinal succession ranges in age from late Proterozoic Triassic. On the southwest a late Paleozoic metamorphosed eugeoclinal assemblage which includes argillaceous chert, siliceous tuff, basalt and serpentinite occur. The eugeoclinal assemblage lies in an allochthonous sheet above the autochthonous miogeoclinal succession. The autochthonous succession is imbricated by large northeast directed thrust faults and the allochthonous assemblage is broken by steep faults. All the rocks were arched over two large northwest trending anticlines which were the locus of relatively long-lived heat flow accompanied by intrusion of elongate mid-Cretaceous granitic batholiths. Transcurrent movement along Tintina Fault occurred shortly after the intrusive event.

Stratigraphy and Facies Relations of the Miogeoclinal Rocks

Figure 21.1 is a schematic diagram of the facies relations of the main stratigraphic units across the Pelly Mountains. It portrays two important breaks in sedimentation, one about Middle Cambrian time and a second during Late Silurian and Early Devonian time. The diagram also demonstrates that the sequence on the northeast near Tintina Trench is nearly complete, but that much of the early Paleozoic sequence on the southwest is missing, presumably through erosion.

The Windermere to Lower Cambrian sequence (Fig. 21.2) is relatively uniform in facies across the Pelly Mountains (*see* also Tempelman-Kluit *et al.*, 1975). Generally it resembles a succession of the same age to the south in the Omineca Crystalline Belt (Campbell *et al.*, 1973; Gabrielse, 1963; Gabrielse and Dodds, 1974). The continuity of facies of this assemblage across the Pelly Mountains and along much of the Omineca Crystalline Belt implies long-lived stable depositional conditions.

The Upper Cambrian-Ordovician-Silurian succession can be divided into four lithologically distinct units. Near Tintina Trench Upper Cambrian calcareous slate and phyllite and argillaceous limestone overlie the Lower Cambrian rocks unconformably. The calcareous slate, about 1000 m thick, is overlain conformably by about 1000 m of recessive weathering fissile black graptolitic slate ranging from Early to Late Ordovician. Locally the black slate contains thin beds of medium grained orthoquartzite, thin flows of intermediate volcanic rocks, thin discontinuous beds of dolomitized mudstone and lenses of crinoidal wackestone. Southwestward these two lithologic units lose their identity and as a result of a change in facies merge to be replaced by a third unit which is 1500 to 2000 m of noncalcareous, medium-grey phyllite, with abundant intermediate to basic volcanic flows, and sills in its upper part. The fourth unit, Lower and Middle Silurian laminated to thin-bedded dolomitic siltstone overlies the black graptolitic slate or the equivalent noncalcareous phyllite and is intertongued with these rocks at the contact. The upper part contains lenses of dolomitized crinoidal packstone and wackestone. Rusty weathering purple and green tuff of intermediate composition are locally important with the dolomite. The entire siltstone unit represents a gradual shoaling from the relatively deeper water, restricted environment under which the black graptolite unit was laid down. The volcanic rocks are submarine flows.

The Upper Cambrian calcareous phyllite and Ordovician black graptolitic slate are equivalent to, and of the same facies as the Kechika Group

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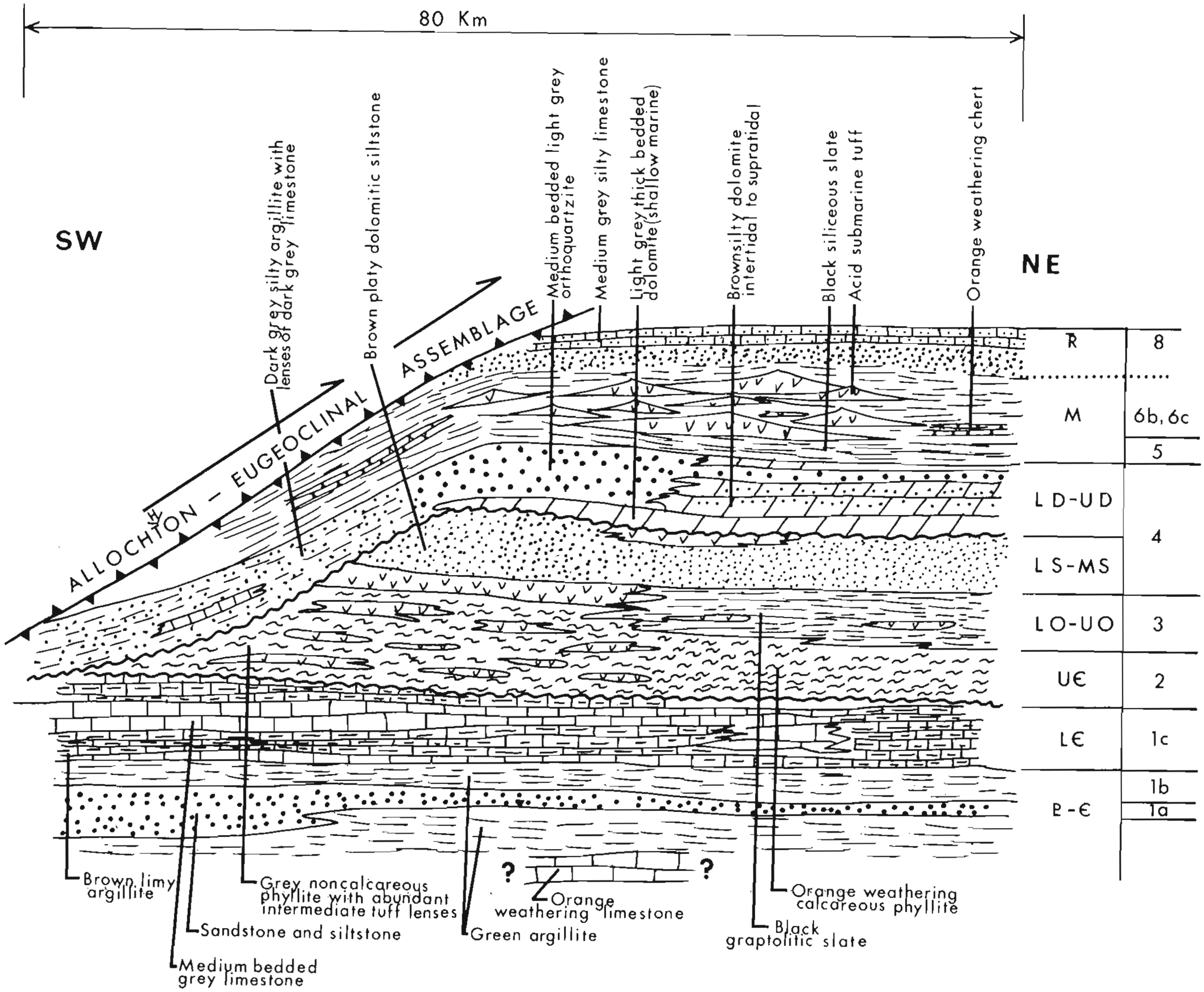

Figure 21. 1. Schematic diagram of facies relations in miogeoclinal strata of the Pelly Mountains. The ages of rock units and their correlation with the map-units of Wheeler *et al.* (1960a, b) are indicated on the right.

(Gabrielse, 1963). Their facies equivalent, the noncalcareous phyllite with abundant volcanics, is identical to the "phyllite unit" in the Anvil district (Map-unit 3, Tempelman-Kluit, 1972).

A disconformity and elsewhere an angular unconformity separates the Middle (and Lower) Devonian dolomite and orthoquartzite from the underlying strata (Fig. 21. 3). On the northeast the Middle Devonian sequence includes, from the base up, about 600 m of thick-bedded light grey dolomite, 700 m of thin-bedded, brownish weathered silty dolomite, 300 m dolomitic mudstone, 800 m of thin-bedded brownish weathered silty dolomite and 300 m of light grey medium grained orthoquartzite. This sequence was deposited in an intertidal environment, but parts represent supratidal deposition (Fig. 21. 4 and 21. 5). To the southwest the upper part of the dolomite sequence is replaced by 800 m of medium-bedded, medium to light grey, mature orthoquartzite and sandy dolomite evidently deposited in a beach environment.

Mississippian black siliceous slate with minor greywacke, roughly 500 m thick, overlies the dolomite and sandstone sequence in many places without a significant break. The black slate is a transgressive sequence that represents a return to deeper water, quiet sedimentation following subsidence of the stable shelf on which the underlying carbonate rocks formed. Lenses of thinly laminated barite, locally as thick as 50 m, occur in the upper part of the black slate unit. Conodonts and brachiopods, tentatively identified as mid-Mississippian (B. E. B. Cameron, pers. comm. ; E. W. Bamber, pers. comm.), were collected from the upper part of the black slate. The slate is overlain by 100 m of orange weathered, thin-bedded, pale green tuffaceous chert which is in turn covered by laminated strongly bioturbated shale and siltstone (Fig. 21. 6)

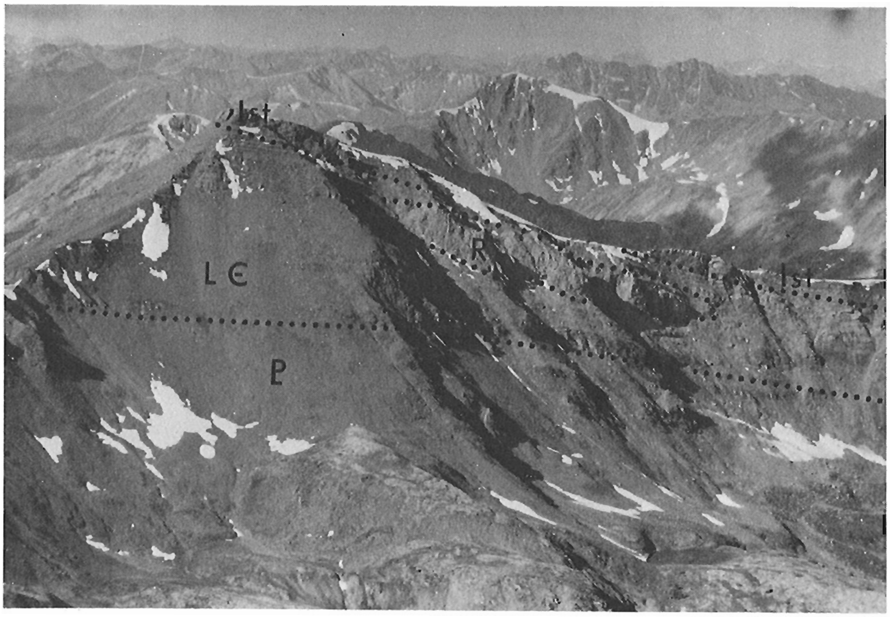

Figure 21. 2.

Late Precambrian and Lower Cambrian stratigraphy near Ketza River. Green argillite with interbedded siltstone and quartzite (Ᵽ) forms the base of the hill. Brown limy argillite and argillaceous limestone (LЄ) with archeocyathid "reefs" (R) comprise the lowest fossiliferous beds and is capped by medium-bedded limestone (1st) found extensively in the Lower Cambrian both in the Pelly Mountains and in other parts of the Omineca Crystalline Belt.

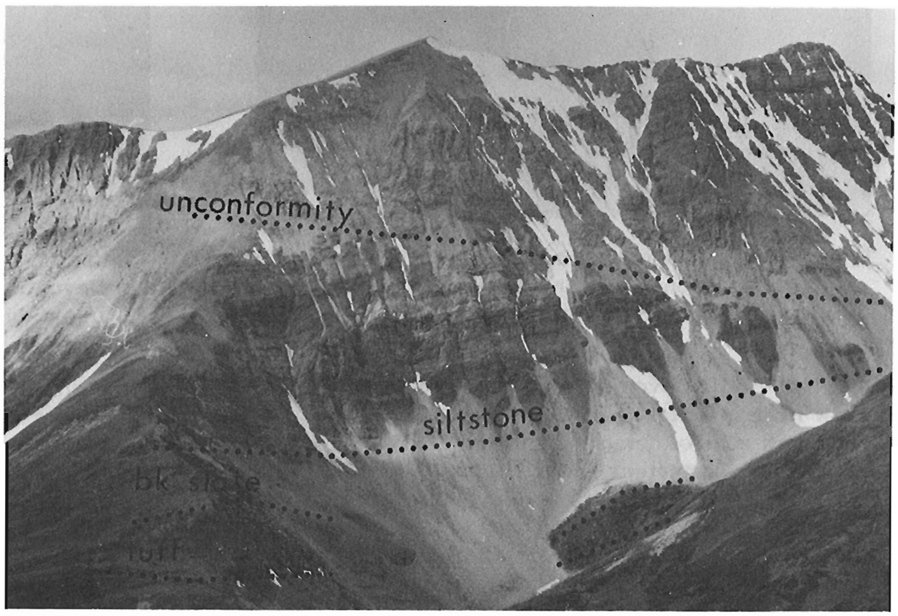

Figure 21. 3.

Angular unconformity between Silurian brown thin-bedded dolomitic siltstone with lenses of intermediate tuff and Lower and Middle Devonian light grey, thick-bedded dolomite in west-central Finlayson Lake map-area. Ordovician black slate, with tuff lenses, underlies the siltstone conformably.

several hundred metres thick. Silty and sandy, medium grey, thin-bedded limestone (Fig. 21. 7) as much as 500 m thick overlies the siltstone and contains Middle or Upper Triassic conodonts (B. E. B. Cameron, pers. comm.). Southwestward the orange weathered chert grades into acid to intermediate, submarine explosive volcanic rocks that are, with intercalated black slate, 900 m thick, Syenite, presumably a subvolcanic relative of the extrusive rocks, occurs as several small plugs within the volcanics. The acid volcanic pile is made up of a number of coalescing sheets of ejecta extruded from several centres. Still farther southwest the acid volcanic rocks grade once again into orange weathered chert.

Structure of the Miogeoclinal Sequence

The miogeoclinal succession is imbricated by northeasterly directed thrust faults (Fig. 21. 8) which involve strata as young as Middle Triassic. One of these, the Porcupine thrust (Fig. 21. 9), is folded over an open syncline-anticline pair about 15 km across. The internal response of the different rocks to the imbrication by thrusting varies with the lithology and with the degree to which the rocks are metamorphosed. Argillaceous strata below the Middle Devonian are folded and bedding is transposed to various degrees (Fig. 21. 10), whereas the competent Silurian siltstone

Figure 21. 4. Birdseye structure in dolomitized mudstone (Middle Devonian) near Porcupine Creek indicating a supratidal environment of deposition.

Figure 21. 5. Oncolites in dolomitic siltstone (Middle Devonian), near Porcupine Creek indicating an intertidal environment of deposition.

and the Middle Devonian carbonate are faulted, but show little internal deformation. Consequently an important zone of detachment (Fig. 21. 8) separates the relatively incompetent Ordovician and older strata from the overlying Silurian siltstone. Southwestward where the miogeoclinal sequence is weakly metamorphosed the detachment zone has migrated up section and is found between the thick-bedded Middle Devonian carbonate and the siltstone beneath it.

Allochthonous Eugeoclinal Assemblage

The eugeoclinal assemblage in the southern part of the Pelly Mountains includes quartz mica schist, the metamorphic product of argillaceous chert and siliceous slate, amphibolite and greenstone (formerly basalt) and serpentinized peridotite. The metamorphism in the lower greenschist facies was accompanied by extensive shearing and the development of foliation. The age and internal stratigraphy of this assemblage is unknown. Parts resemble the Anvil Range Group (Tempelman-Kluit, 1972) while other sections are like the low grade metamorphic rocks found in the contiguous Englishman's Complex (Mulligan, 1963) and the Sylvester Group (Gabrielse, 1963). On the basis of these broad similarities and the few fossil collections made in equivalent strata the eugeoclinal rocks of the southern Pelly Mountains may be largely or in part late Paleozoic.

The eugeoclinal rocks lie on the miogeoclinal sequence and a single sharp, gently dipping contact, not modified by later events separates the two assemblages (Fig. 21. 11). In most places the youngest beds of the underlying sequence are the Mississippian black slate and orange weathered chert. South of McNeil Lake the Middle Triassic beds are exposed beneath the eugeoclinal rocks. In most places the eugeoclinal strata are distinctly more metamorphosed than the miogeoclinal sequence on which they lie (Wheeler *et al.*, 1960a; Gabrielse and Wheeler, 1961, p. 21). These relations indicate that the eugoeclinal rocks are allochthonous and are thrust over the miogeoclinal strata. Emplacement of the allochthon postdates the Middle Triassic, but predates the latest Triassic and therefore occurred during the Late Triassic (? Karnian?).

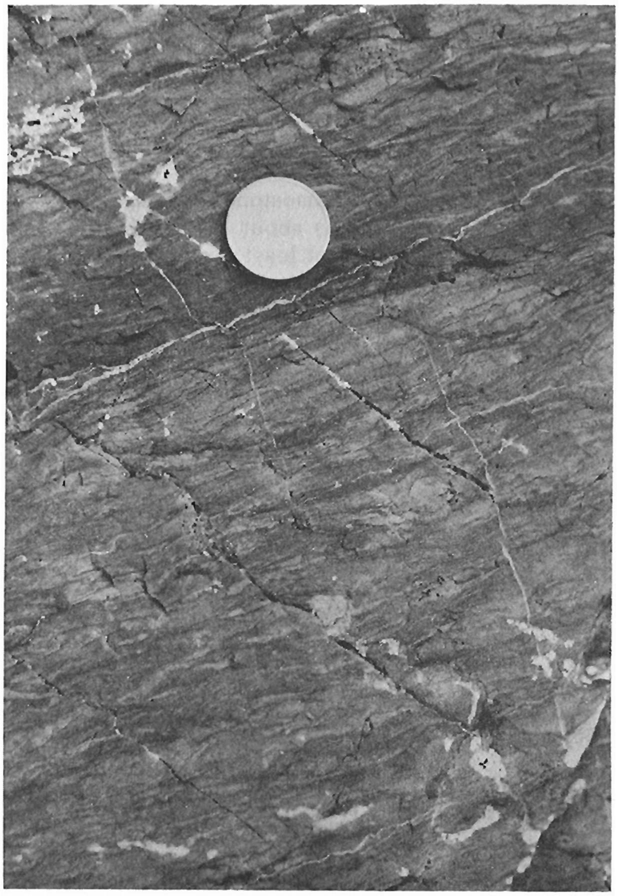

Figure 21. 6.

Brown bioturbated argillaceous siltstone (Carboniferous or Permian) found in the starved shallow marine sequence beneath Triassic beds in the Pelly Mountains.

Figure 21. 7.

Grey thin-bedded silty and sandy limestone with abundant fossil fragments and rounded limestone clasts deposited under shallow marine conditions and relatively high energy (Triassic). These are the youngest rocks of the miogeoclinal sequence in the Pelly Mountains.

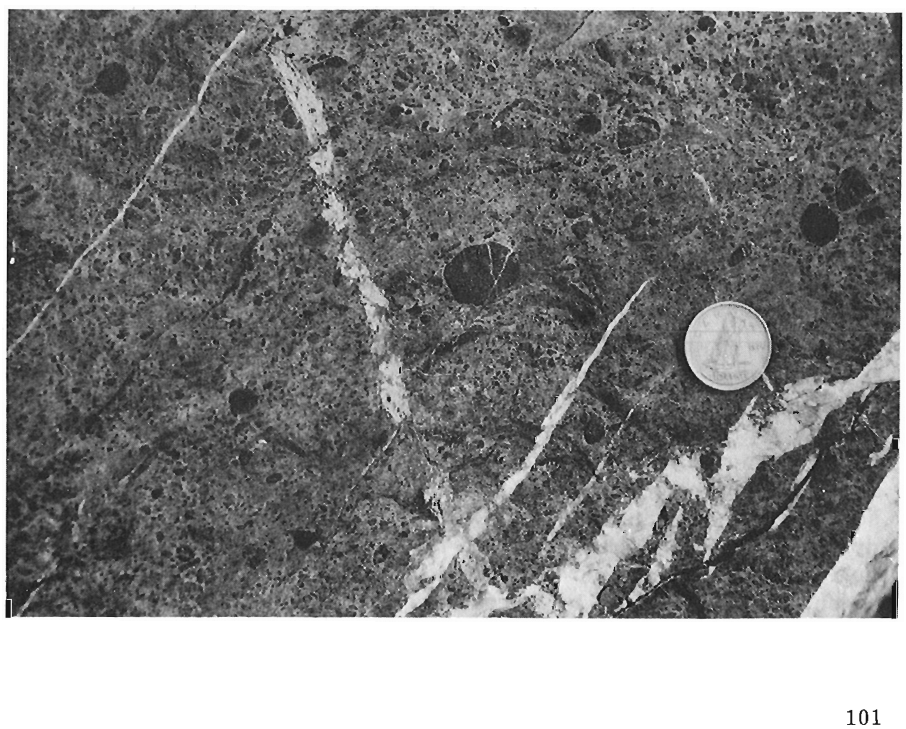

Figure 21. 8.

A single limestone unit (lower Cambrian) about 150 m thick is repeated at least ten times by northeast dipping, northeast directed thrust faults northwest of Big Salmon Lake. Only a few of the repetitions are see in this view in which the imbricate sheets are numbered arbitrarily.

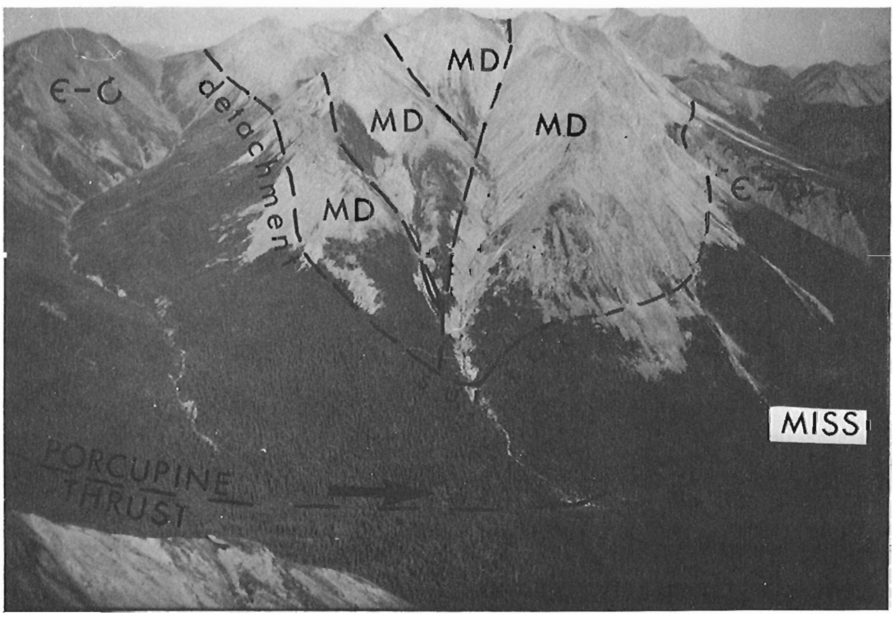

Figure 21. 9.

Thick-bedded competent Middle Devonian dolomite (MD) lies in the core of a syncline on Cambro-Ordovician calcareous phyllite with tuff lenses. The incompetent phyllite is deformed internally by small-scale transposition of bedding (*see* Fig. 21. 10), but the dolomite is broken only by large faults and lacks internal small-scale deformational structures. A surface of detachment separates the two sequences. The dolomite in the core of the syncline lies in four discrete fault-bounded slabs. The Cambro-Ordovician phyllite is separated from the underlying Mississippian black slate (MISS) by the Porcupine thrust, a large, synclinally folded, northeast directed thrust fault.

Mid-Cretaceous Plutonic Rocks

Mid-Cretaceous granitic rocks occupy the cores of two broad, northwest-trending arches that are flanked by mantling zones of injection migmatite up to 10 km wide. The granitic rocks are medium grained biotite quartz monzonite with distincitve large phenocrysts of subhedral pink potash feldspar. The injection migmatite is a chaotically mixed zone of relatively dark coloured mica rich gneissic and schistose screens and septa invaded by a variety of sills and dykes of lighter coloured, medium to fine grained quartz monzonite. Upper Proterozoic rocks of the green argillite unit can be traced laterally into the migmatite zones and the migmatite clearly includes the metamophosed and partly mobilized Windermere sequence. The rocks contain biotite and garnet and metamorphism was in the lower amphibolite facies. The migmatite zones, with small quartz monzonite bodies, represent the upper parts of a sedimentary succession metamorphosed in place at relatively low temperature and high water pressure along two roughly parallel, long-lived zones of relatively high heat flow. The rocks experienced little internal deformation during metamorphism. Broad arching of the imbricated miogeoclinal rocks and the overlying allochthonous strata accompanied the slow rise of material in the zones of high heat flow.

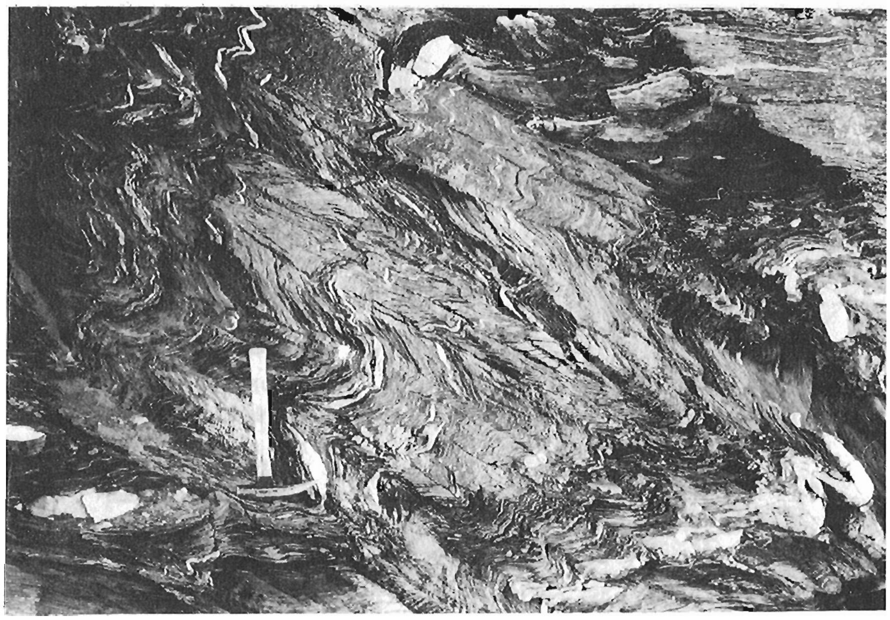

Figure 21.10.

Transposed bedding in dark grey, thinly laminated, calcareous phyllite (Cambro-Ordovician). Crystalization of mica along the crenulation cleavage is seen locally. This style of deformation is characteristic of Cambro-Ordovician rocks in much of the Omineca Crystalline Belt.

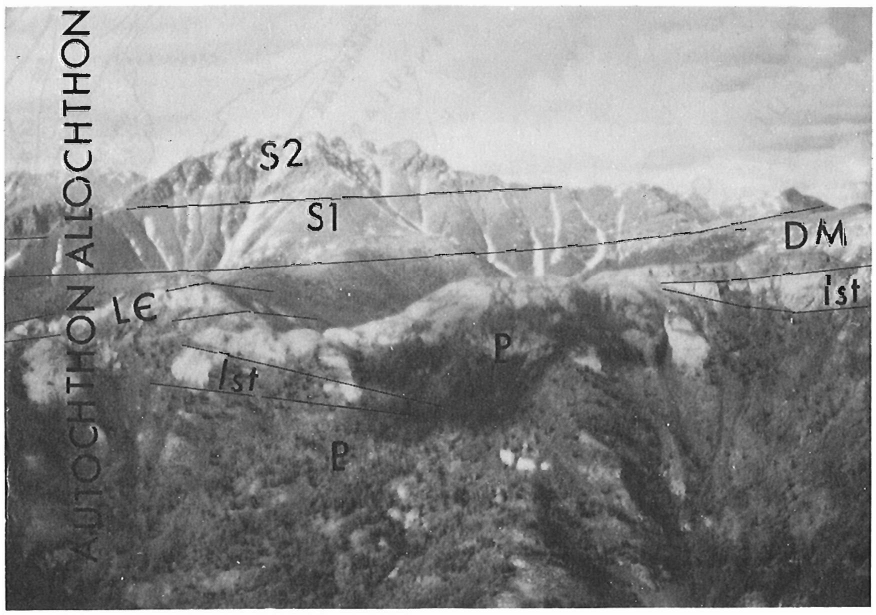

Figure 21.11.

Serpentinized peridotite forms the resistant outcrops on the top of Dunite Mountain and lies as a flat sheet on the miogeoclinal sequence. The periodite mass is interpreted as allochthonous here and its lower contact is thought to be a regional thrust fault, modified in many other places by folding, faulting and intrusion.

- P - Late Proterozoic mica schist (Green argillite unit) with limestone lenses (1st);
- LЄ - limestone;
- D-M - Black slate with lenses of orthoquartzite and dolomite;
- S^1 - Serpentinite;
- S^2 - Serpentinized peridotie.

Regional Considerations

A generalized sketch map of a part of the northern Cordillera (Fig. 21.12) shows the boundary between the allochthonous eugeoclinal rocks and the miogeoclinal strata on which they rest as interpreted from present work and from previous mapping in the region. From its position in the northern Omineca Crystalline Belt the boundary is offset in a right lateral sense along the Tintina Fault. The boundary is a gently dipping thrust fault modified in places by later faults, deformation, metamorphism and plutonism. The contact is probably not a simple sole thrust, but an imbricate zone of faults and the allochthonous sheet, rather than being a single coherent mass is probably faulted and imbricated and may incorporate slices of the overridden platform.

The area of low to medium grade metamorphic rocks in the northern Yukon Crystalline Terrane is shown as part of the allochthonous sheet, but the area may include large windows through the allochthon or important slices of the overridden miogeoclinal strata. Complication by later and possibly earlier structural events and superposed metamorphism has prohibited delineation of such features with present data. This structural

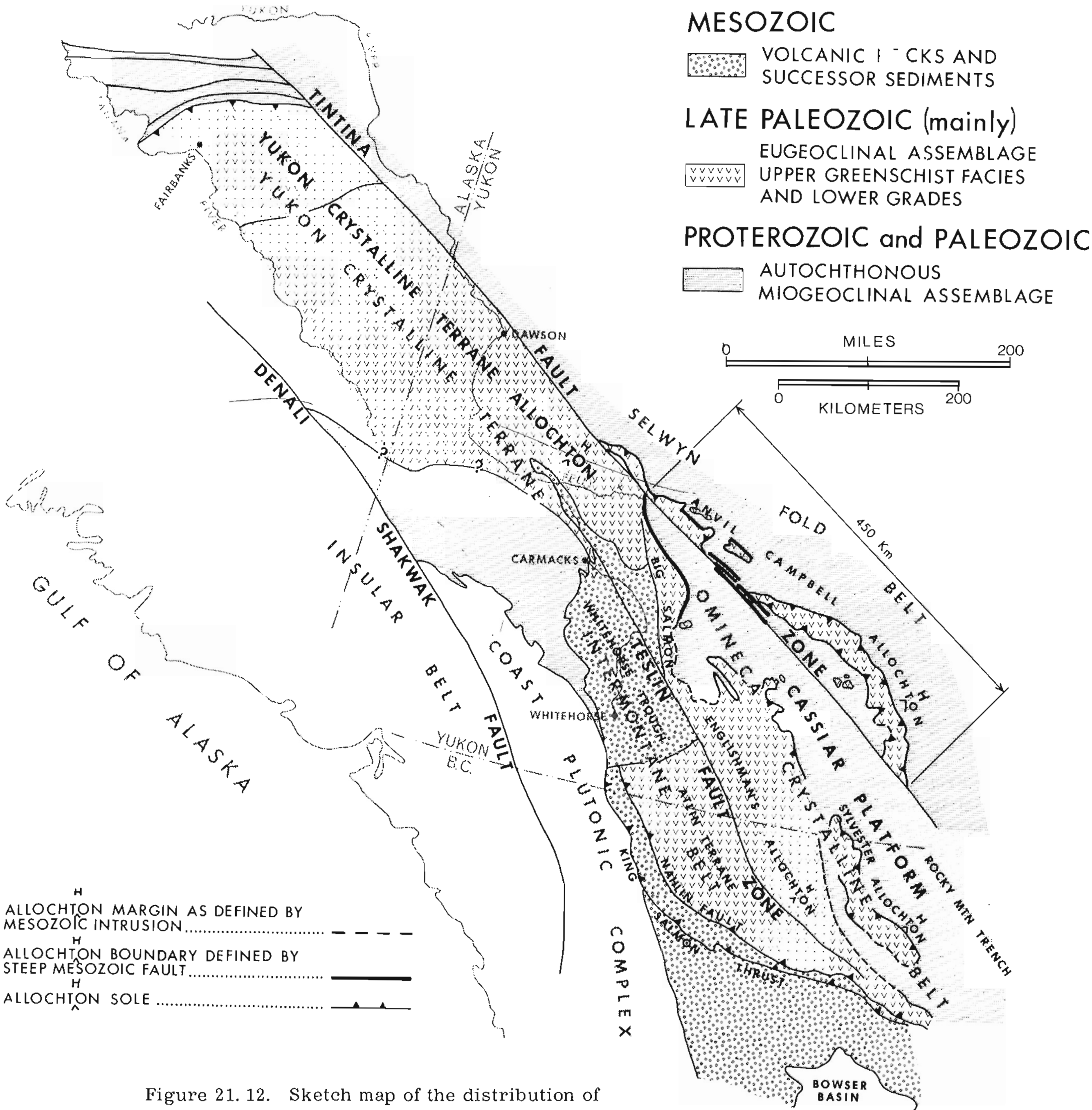

Figure 21. 12. Sketch map of the distribution of eugeoclinal and miogeoclinal strata in the northern Omineca Crystalline Belt. Mesozoic intrusions are omitted. In the Pelly Mountains the eugeoclinal strata are thrust over the miogeoclinal sequence and a similar relationship modified in many places by younger phenomena is postulated on trend. The allochthonous eugeoclinal rocks are offset by mid-Cretaceous movement on the Tintina Fault a distance of about 450 km. Rocks in the Yukon Crystalline Terrane are the continuation of those in the allochthonous part of the Omineca Crystalline Belt.

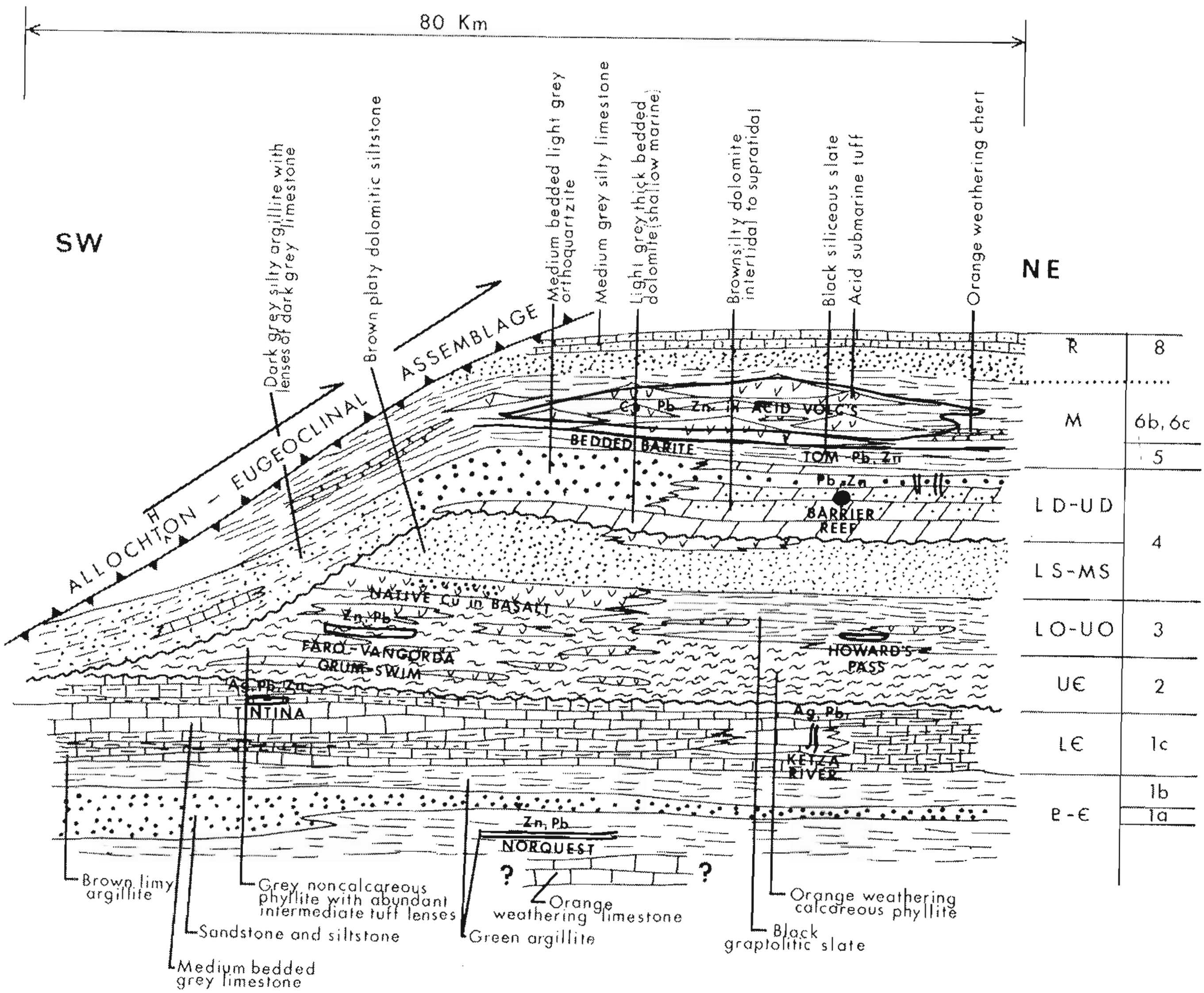

Figure 21.13. Mineral occurrences in southern Yukon that seem to have stratigraphic control are superposed on a schematic facies diagram of the Pelly Mountains to emphasize the presence of these stratigraphic targets for mineral exploration in this region.

interpretation of the northern Yukon Crystalline Terrane implies that the bulk of the rocks formerly included in the Yukon Group or Birch Creek Schist are late Paleozoic.

Displacement of the allochthonous sheet by about 450 km provides an approximate measure of the transcurrent movement on the Tintina Fault.

Guides for Mineral Exploration

A number of stratum controlled mineral occurrences in southern Yukon are shown superposed on the schematic facies diagram of the miogeoclinal rocks of the Pelly Mountains (Fig. 21.13). The diagram emphasizes the presence in the Pelly Mountains of strata of the same facies and age as those in which various mineral occurrences are found elsewhere in southern Yukon. By inference these stratigraphic units in the Pelly Mountains warrant careful examination.

One stratigraphic unit of particular interest to mineral exploration is the strongly sheared Mississippian acid and intermediate volcanic suite (Map-units 6b and 6c of Wheeler *et al.*, 1960a, b) that occupies the large area near the head of Cloutier Creek. These rocks locally contain as much as 10 or 15 per cent of finely disseminated pyrite and they form spectacular orange and red gossans in which the rocks are deeply leached. In places the rocks contain minor malachite on fracture surfaces and elsewhere zinc-lead mineralization is known. For example the zine-lead occurrence 6 km west of the mouth of White Creek known as the MM Claims (Sinclair and Gilbert, 1975) or the Arnold property lies entirely within the sheared acid volcanic suite and not within older rocks as was previously thought.

References

Campbell, R. B., Mountjoy, E. W., and Young, F. G.
1973: Geology of McBride map-area, British Columbia; Geol. Surv. Can., Paper 72-35, 104 p.

Gabrielse, H.
1963: McDame map-area, Cassiar District, British Columbia; Geol. Surv. Can., Mem. 319, 138 p.

Gabrielse, H. and Dodds, C. J.
1974: Operation Finlay; Geol. Surv. Can., Paper 74-1, Part A, p. 13-16.

Gabrielse, H. and Wheeler, J. O.
1961: Tectonic framework of southern Yukon and northwestern British Columbia; Geol. Surv. Can., Paper 60-24, 37 p.

Mulligan, R.
1963: Geology of Teslin map-area, Yukon Territory; Geol. Surv. Can., Mem. 326, 96 p.

Sinclair, W. D. and Gilbert, G. W.
1975: Mineral Industry report 1973, Yukon Territory; Dep. of Indian and Northern Affairs, EGS 1975-7, p. 169.

Tempelman-Kluit, D. J.
1972: Geology and origin of the Faro, Vangorda and Swim Concordant zinc-lead deposits, central Yukon Territory; Geol. Surv. Can., Bull. 208, 73 p.

Tempelman-Kluit, D. J., Abbott, G., Gordey, S., and Read, B. C.
1975: Stratigraphic and structural studies in the Pelly Mountains, Yukon Territory; Geol. Surv. Can., Paper 75-1A, p. 45-48.

Wheeler, J. O., Green, L. H., and Roddick, J. A.
1960a: Finlayson Lake, Yukon Territory; Geol. Surv. Can., Map 8-1960.

1960b: Quiet Lake, Yukon Territory; Geol. Surv. Can., Map 7-1960.

CAPE FLATTERY MAP-AREA (92 C), BRITISH COLUMBIA

Project 73006

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Introduction

Reconnaissance mapping of the Canadian part of Cape Flattery map-area (92C), started in 1973, was almost completed in the 1975 field season. As in the other half of this project, the Victoria map-area (Muller, 1975), enough data are now at hand for the compilation of a regional geological map (Fig. 22. 1) but more detailed work on structure and stratigraphy are required to complete the project.

The area was the subject of a geological reconnaissance by C.H. Clapp (1912) and the Cowichan Lake region was mapped and studied in detail by J.T. Fyles (1955). K. Northcote (1972) examined the geology of the "Nitinat Triangle". In addition several mining company maps were useful, especially in locating small outcrops of limestone, so essential as marker beds in this otherwise mainly volcanic and crystalline terrane.

The map-area includes parts of the three main tectonic units of Vancouver Island. Figure 22. 2 shows the contrasting geological histories of these three blocks bounded by fundamental faults. The Vancouver Island block is bounded to the south and southwest by San Juan and Westcoast faults. The San Juan Block is juxtaposed along these faults and bounded to the south by Leech River faults. The Metchosin Block lies south of Leech River Fault. Another fault of major consequence is the Cowichan Lake Fault that separates a block of mainly Paleozoic rocks to the north from a block of Mesozoic rocks to the south. A brief review of some salient points regarding the formations of the area follows, but more extensive descriptions are available in papers by Fyles, 1955; Yole, 1969; Muller and Carson, 1969 and Muller, Northcote and Carlisle, 1974. Various points discussed in the earlier preliminary reports on the map-area (Muller, 1973, 1974 and 1975) are only briefly restated.

Pre-Devonian sediments

According to current interpretation the Sicker Group embraces all Paleozoic rocks of Vancouver Island From oldest to youngest it includes a volcanic, a clastic sedimentary, and a limestone formation. Recent field work and age determinations indicate that this succession is underlain by one or more other formations that are intruded by early Paleozoic granitic rocks. The author (Muller, 1975) has briefly described the Tyee Intrusions of quartz porphyry, converted to quartz-augen schist, and altered granitoid rocks. The rocks occur in the southwest part of Saltspring Island and continue across Sansum Narrows north of Maple Bay on Vancouver Island. They are emplaced into a greywacke-argillite sequence and both sediment and intrusion have subsequently been shearfolded. On the basis of this structural evidence and the total dissimilarity with Mesozoic intrusions in the region, the granitoid rocks were assumed to be late Paleozoic. R.K. Wanless has now obtained zircons from the quartz porphyry that have yielded only slightly discordant ages for two size-fractions. They indicate the porphyry is at least 390 m.y. old, or very early Devonian. In San Juan Islands (Mattinson, 1972 and pers. comm.) granitoid rocks of the Turtleback Complex have yielded roughly similar ages that, when plotted on a concordia diagram, give a primary age of 450 m.y. On Vancouver and Saltspring islands some shearfolded greywacke-argillite sequences are therefore of pre-Devonian age. On Orcas Island, although Turtleback Complex rocks are generally considered to be "basement" the writer has observed granitoid rocks, intruding basic volcanic rocks with well preserved pyroclastic and agglomeratic texture. The granitoid rocks are similar and probably continuous to those dated pre-Devonian. Thus the possibility that some Paleozoic "Sicker" volcanics are pre-Devonian is equally valid.

The scale of mapping has not yet permitted separation of Sicker sedimentary and volcanic rocks into pre- and post-intrusive formations. Lithological or paleontological criteria are not available to separate the sedimentary formations, comprising greywacke, argillite and minor limestone. Structural differences may indicate if folding accompanied or preceded early Paleozoic intrusions. Such differences are indeed present: some Paleozoic sediments are almost horizontal and undisturbed, while others are isoclinally folded and have slaty cleavage. The problem of separating the formations is further compounded by their similarity to the Jurassic-Cretaceous greywacke-argillite sequence (Leech River Formation) that was shearfolded in early Tertiary time. Dykes and sills cutting the sediments may also be useful criteria. In some areas sediments are extensively invaded by coarse grained uralite porphyry dykes (e.g. Shaw Creek area). The dykes are feeders of Sicker volcanic rocks and the sediments therefore belong to the older pre-volcanic sequence. Other areas of sediments (e.g. Reynhardt Creek area) are invaded by gabbroic dykes and sills, locally with plagioclase phenocrysts, not uncommonly in large clusters. These are almost certainly related to Triassic Karmutsen volcanics and suggest a late Paleozoic age for the sediments.

Sicker volcanics

Volcanics of the Sicker Group are breccias, agglomeratic lavas and massive and thinly bedded tuffs. Primary textures are best seen on weathered surfaces but are barely visible on fresh surfaces in new cuts on logging roads. Many rocks exhibit conspicuous uralite

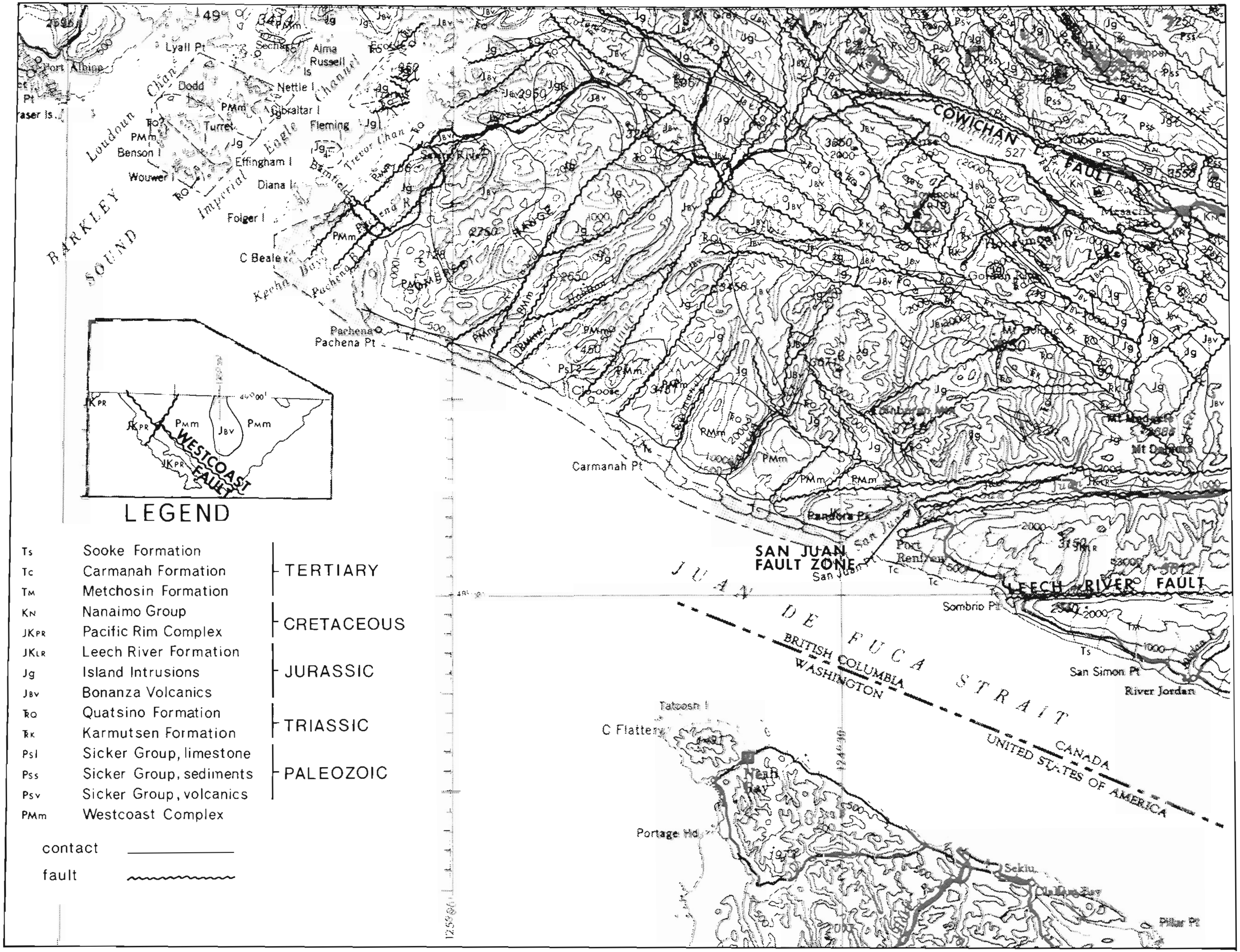
COWICHAN FAULT
LEECH RIVER FAULT
SAN JUAN FAULT ZONE
JUAN DE FUCA STRAIT
BRITISH COLUMBIA
WASHINGTON
CANADA
UNITED STATES OF AMERICA
BARKLEY SOUND
Loudoun Chan
Imperial Eagle Channel
Trevor Chan
Port Albion
Lyall Pt
Dodd
Nettle I
Alma Russell Is
Gibraltar I
Fleming
Turret
Benson I
Wouwer I
Effingham I
Diana I
Folger I
C Beale
Keeha Bay
Bamfield
Pachena Pt
Carmanah Pt
Port Renfrew
San Juan Pt
Sombrio Pt
San Simon Pt
River Jordan
Tatoosh I
C Flattery
Neah Bay
Portage Hd
Sekiu
Pillar Pt
Mt Dallas
WESTCOAST FAULT
LEGEND
Ts Sooke Formation
Tc Carmanah Formation
TM Metchosin Formation
TERTIARY
KN Nanaimo Group
JKPR Pacific Rim Complex
CRETACEOUS
JKLR Leech River Formation
Jg Island Intrusions
JBV Bonanza Volcanics
JURASSIC
ꞦQ Quatsino Formation
ꞦK Karmutsen Formation
TRIASSIC
PsI Sicker Group, limestone
PSS Sicker Group, sediments
PSV Sicker Group, volcanics
PMm Westcoast Complex
PALEOZOIC
contact
fault

Figure 22. 1.

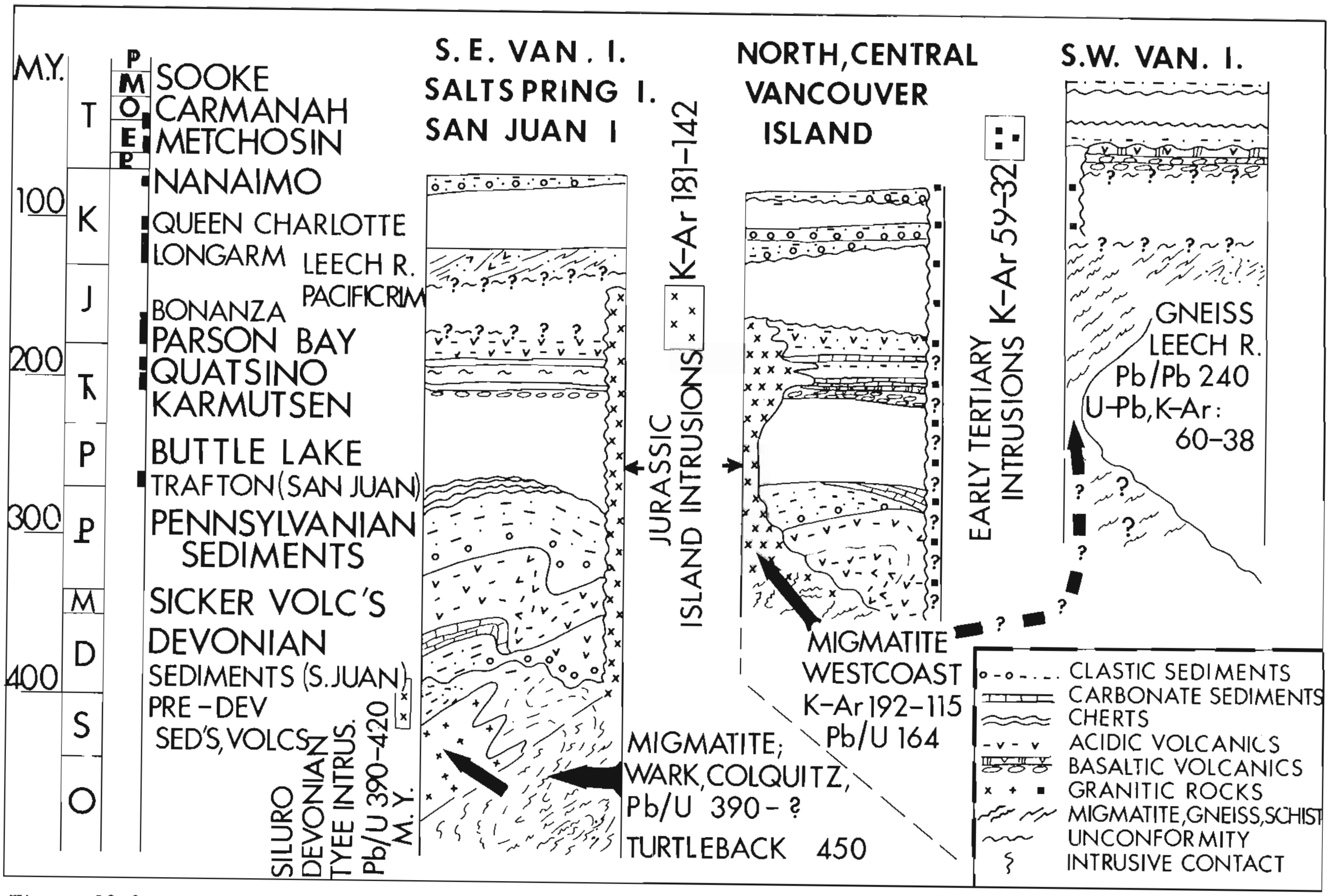

Figure 22. 2.

(actinolitized augite) phenocrysts and these are either crystal tuffs or agglomeratic lavas containing fragments of coarse grained uralite porphyry in a finer grained matrix. The rocks are commonly low-grade greenstones with epidote, actinolite, quartz and prehnite common in vugs and veinlets. In some areas the volcanics have been converted to schistose rocks almost completely devoid of primary structures. Some are hard siliceous and strongly rodded rocks, others are flaky chlorite-talc schists. Folding and crenulation of cleavage planes is common and indicate at least two phases of tectonism. So far no conclusive evidence has been found that the Sicker volcanics include rocks of pre-Devonian age.

Sicker sediments

Well-bedded turbiditic greywacke-argillite sequences of the Sicker Group underlie much of the northeast corner of the map-area. They have a characteristic light-dark banding on smooth joint faces, that are roughly perpendicular to bedding. They are siliceous and near intrusions hornfelsic although elsewhere slaty cleavage is common. In a few places they contain layers of breccia with scattered angular fragments probably representing rip-up clasts. The sediments have been intruded pervasively by sills of basalt-porphyry, commonly with conspicuous single or clustered plagioclase phenocrysts. As earlier suggested by Fyles (1955) the porphyries are probably comagmatic with overlying Karmutsen volcanics.

Sicker limestone

Sicker limestone (Buttle Lake Formation) was mapped by Fyles (1955) and Yole (1969) in the southeast part of the map-area. No other occurrence of these rocks was found. The limestone is commonly recrystallized but distinctly crinoidal. It is interbedded with clastic and cherty sediments. Recrystallized limestone exposed at the south end of Nitinat Lake ("Nitinat Formation" of Clapp, 1912) contains cherty layers and is associated with siliceous tuffs. The lithology is more like Paleozoic than Triassic Quatsino limestone, but could also represent the more siliceous Triassic Sutton limestone.

Karmutsen Formation

Karmutsen volcanic rocks form an almost continuous belt, only a few miles wide and broken by many faults, trending west-northwesterly through the ranges south of Cowichan Lake to Alberni Inlet. Their steep, rough-surfaced, red brown weathering bluffs are as distinct as their pillowed or thickly layered structure. In the map-area pillow lavas and aquagene breccias appear to be more common than layered amygdaloidal lavas. The narrow outcrop belt, narrowing further in Victoria map-area, suggests that the formation is much thinner than in the central part of Vancouver Island and probably contains only between 1000 and 2000 m of mixed pillowed and layered basaltic lavas.

Quatsino Formation

Upper Triassic post-Karmutsen sediments in central and northern Vancouver Island are divided into Quatsino Formation (limestone) and Parson Bay Formation (mixed fine clastic and carbonate sediments). Sutton limestone forms a member at the top of Parson Bay Formation. Although lithologies and diagnostic fossils of all three units occur in the map-area, Quatsino limestone dominates by far. As mapped it includes beds of Parson Bay and Sutton, commonly less than 10 m in exposed thickness. The greatest thickness of Quatsino limestone, estimated at about 300 m, occurs in the Gordon River area, south of Cowichan Lake. The limestone thins eastward and is missing in adjacent Victoria map-area. It also decreases in thickness westward, is very thin in Nitinat River area, but thickens again towards Alberni Inlet.

Typically, the rock is a dark grey to black, blue grey weathering fine crystalline limestone. Where the formation is thin bedding is commonly obscure, but thick sections exhibit distinct beds of 10 cm to 1 cm thick separated by calcarenitic layers. Corals have not been found and the beds are more probably algal reefs. The limited extent of the reefs suggests they may have been atoll-like structures on seamounts of Karmutsen volcanics, rather than basin deposits.

Black, brown weathering calcareous and shaly siltstone, commonly carrying imprints of *Halobia* or *Monotis*, and typical of Parson Bay Formation, are rare in the map-area and are found mainly in areas where Quatsino limestone is thin. Likewise, Sutton limestone, with type locality on Cowichan Lake, is uncommon and its stratigraphic relationships with Quatsino limestone cannot be established in this area. Farther north (Muller *et al.*, 1974) Sutton limestone locally forms the top of Parson Bay Formation that overlies Quatsino limestone, but it seems to be a lateral equivalent of Quatsino limestone in the Cape Flattery map-area. Where exposed the limestone is detrital and largely composed of shell fragments.

Bonanza Volcanics

Bonanza volcanics underlie large parts of the map-area south of Cowichan Lake Fault and, although accurate measurements are lacking, their thickness probably exceeds that of Karmutsen Formation and is estimated at between 2000 and 3000 m. They are composed largely of massive to poorly bedded tuff and breccia, commonly with deep maroon and dark green colours, weathering to dull brown. Flows and crystal tuffs with feldspar and pyroxene phenocrysts are also common and similar to those in Sicker volcanics, but less altered. Chemical analyses of Bonanza volcanics have shown that tholeiitic basalt is rare and alkali basalt and tholeiitic dacite and rhyolite are the more common compositions. Clastic sediments occur locally, apparently in the lowest part of the formation and may contain abundant pelecypod shells, in many instances entirely decomposed.

Island Intrusions and Westcoast Complex

The Westcoast Complex and Island Intrusions are exposed in a large area between Cowichan Lake and San Juan faults and probably underlie all other formations in the remainder of that block. The Westcoast rocks are various types of agmatite, composed of dark amphibolite and diorite and lighter coloured quartz diorite. They were probably derived from basic volcanic rocks and minor sediments by partial migmatization. The resulting granitic fractions migrated to form the Island Intrusions of mainly quartz diorite, granodiorite and minor quartz monzonite. Accordingly, all contacts of the two crystalline formations with one another and with the older formations are poorly defined and gradational. Even contacts with sediments are arbitrary. Some small islands in Barkley Sound consist of highly folded contact metamorphic bedded sediments, of assumed Triassic age, grading into amphibolite and agmatite.

Pacific Rim Complex and Leech River Formation

Pacific Rim Complex has been described in previous reports (Muller, 1973; Muller *et al.*, 1974) as a structurally chaotic assemblage of mainly greywacke and argillite underlying most of Pacific Rim National Park and adjacent Esowista and Ucluth peninsulas. Only the latter is part of the map-area (inset corner west of 124°30'W). *Buchia pacifica* Jeletzky found in a few places indicates an earliest Cretaceous, Valanginian age for some of these beds. Ribbon chert had tentatively been dated as Jurassic whereas pillow lava and limestone were assigned a probable Triassic age. Samples from unaltered, maroon coloured cherts from a small island east of Ucluelet Inlet and one mile northwest of Chrow Island contained good radiolarians. The locality was visited and sampled by A.E. Pessagno Jr. of the University of Dallas, B.E.B. Cameron and the writer, and Dr. Pessagno kindly supplied the following report on the contained fauna and listed the following species from 6 samples taken across a section of about 10 m thick:

Eucrytidium (?) *ptyctum* Riedel and Sanfilippo

Praceonocaryomma mamillaria (Rust)

Praeconocaryomma magnimamma (Rust)

Hsuium maxwelli Pessagno

Hsuium cuestaensis Pessagno

Hsuium obispoensis Pessagno

Pantanellium riedeli Pessagno

Archaeodictyomitra rigida Pessagno

Parvacingula turrita (Rust)

Paronaella sp.

Spongocapsula palmerae Pessagno

Parvacingula sp.

Emiluvia sp.

Biostratigraphic determination: Subzone 2A; early Tithonian/late Kimmeridgian

The Subzone 2A forms part of a new zonation of the uppermost Jurassic, (Tithonian) sequence of the Coast Ranges of California, introduced by Pessagno (in press). The new dating thus lends further support to correlation of Pacific Rim Complex with the Franciscan Terrane of California.

Leech River Formation occupies the belt between San Juan and Leech River faults and consists of shear-folded greywacke and argillite. Metamorphic grade increases from slate and phyllite in the north near San Juan Fault to garnetiferous quartz-biotite schist in the south near Leech River Fault. Though previously assumed to be Paleozoic the rocks are now considered more likely to be Jurassic to Cretaceous and correlative to Pacific Rim Complex (Muller, 1975). Several K-Argon datings of biotite in the schist have yielded 40 m.y. ages indicating late Eocene metamorphism of the sediments. Muscovite pegmatite intruding the schists contains zircons that have been studied by R. K. Wanless. The preliminary results are highly discordant ages that suggest the possible remobilization of Paleozoic material to form an early Tertiary intrusion.

The writer believes that Pacific Rim Complex and Leech River Formation were deposits on the Late Jurassic to Early Cretaceous continental slope and the adjacent trench. These formations and the Nooksack Group of Cascade Mountains and San Juan Islands form a link between coeval slope-trench deposits of the Pacific Margin from California to Alaska.

Tertiary Formations

The area south of Leech River Fault is underlain by Eocene Metchosin basalts. Along part of the coast the basalts are overlain by a narrow strip of sandstone and conglomerate of the Sooke Formation of Miocene to Pliocene age. North of Leech River Fault Tertiary clastic sediments, exposed in a narrow zone along the coast, form the Carmanah Formation of Eocene to Oligocene age. No further work was done on these rocks during the field season.

Structure

Most of the discussion on structure in the report on Victoria map-area (Muller, 1975) is relevant to Cape Flattery map-area. Blockfaulting of the crystalline and volcanic rocks is dominant. Shearfolding only occurred in less competent sediments and was subsidiary to faulting.

In the discussion of pre-Devonian sediments the probability of a mid-Paleozoic folding phase affecting certain parts but not all rocks of the Sicker Group has been mentioned. It is also apparent that Leech River rocks were shearfolded in late Cretaceous to Eocene time while Pacific Rim sediments were compressed into a melange. The writer interprets the latter deformation as a result of northward underthrusting of the late Mesozoic slope and trench sediments along San Juan and Westcoast Fault zones. Carmanah sediments unconformably overlying Leech River schists indicate that folding ceased before the end of Eocene time.

The fundamental east-west striking Leech River and San Juan faults separate tectonic blocks of different history and crustal structure. The Cowichan Lake Fault, though less important, separates Paleozoic rocks to the north from Mesozoic rocks to the south and its throw is at least the estimated thickness of Karmutsen volcanics, between 1000 and 2000 m. The faults were probably active during Late Cretaceous to Eocene time as they affect rocks of that age. One part of the Westcoast Fault was definitely inactive since then as it is overlain unconformably by undisturbed upper Eocene Carmanah beds. Where exposed the faults are steeply dipping, commonly to the northeast, but also to the southwest, or vertical. Juxtaposition of formations clearly shows that the movement had a vertical component. Oblique and even horizontal slickensides indicate a horizontal component of movement as well.

Secondary faults appear to occur in sets, more or less characteristic of the formations affected. Sicker and Karmutsen rocks north of Cowichan Lake are cut by a set of northwesterly faults. They could be relatively old faults that are transected by a later westerly striking fault along Meade Creek that offsets Upper Cretaceous Nanaimo beds. Karmutsen and Quatsino formations south of the Lake are offset by a set of closely spaced west-northwesterly faults that terminate abruptly to the east in the Fleet River granitic body. Possibly these faults are directly related to the rifts along which Karmutsen basalts were extruded. A detailed study of lithology and thickness of the formations might show if the faults were active in Upper Triassic time. Crystalline areas show more widely spaced, almost rectangular fault patterns. Northeasterly directed faults may be the youngest and are mainly present in the granitoid rocks near the coast between Barkley Sound and Port San Juan. The lakes in the "Nitinat Triangle" including Nitinat Lake and Nitinat Valley are controlled by these faults. Northeasterly trending faults with considerable offset are also present in Eocene to Oligocene Carmanah sediments.

References

Clapp, C.H.
1912: Southern Vancouver Island; Geol. Surv. Can., Mem. 13.

Danner, W.R.
1966: Limestone resources of Western Washington; Wash. Div. Mines, Geol. Bull. 37.

Fyles, J.T.
1955: Geology of Cowichan Lake area, Vancouver Island, British Columbia; B.C. Dep. Mines, Bull. 37.

Mattinson, J.M.
1972: Ages of zircons from the northern Cascade Mountains, Washington; Bull. Geol. Soc. Am., v. 83, p. 3769-3784.

Muller, J.E.
1973: Geology of Pacific Rim National Park; *in* Report of Activities, Part A, Geol. Surv. Can., Paper 73-1A, p. 29-37.

1974: Victoria map-area (92B, C), Pacific Rim National Park, Vancouver Island, B.C.; *in* Report of Activities, Part A, Geol. Surv. Can., Paper 74-1A, p. 21-23.

1975: Victoria map-area, British Columbia (92B); *in* Report of Activities, Part A, Geol. Surv. Can., Paper 75-1A, p. 21-26.

Muller, J.E. and Carson, D.J.T.
1969: Geology and mineral deposits of Alberni map-area, British Columbia (92F); Geol. Surv. Can., Paper 68-50.

Muller, J.E., Northcote, K.E., and Carlisle, D.
1974: Geology and mineral deposits of Alert Bay — Cape Scott map-areas; Geol. Surv. Can., Paper 74-8.

Northcote, K.E.
1972: The geology of the Nitinat Triangle; B.C. Dep. Mines Pet. Resour., Ann. Rept. for 1972 p. 243-255.

Pessagno, E.A., Jr.
Upper Jurassic radiolaria and radiolarian biostratigraphy of the California Coast Ranges; Inst. Geol. Sci., Univ. Texas at Dallas, Contr. 287. (in press)

Yole, R.W.
1969: Upper Paleozoic stratigraphy of Vancouver Island, British Columbia; Geol. Assoc. Can. Proc., v. 20, p. 30-40.

SUCCESSOR BASINS OF THE WESTERN CORDILLERA

Project 750013

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Introduction

During the past field season a study of structure and sedimentation in Cordilleran successor basins was continued with reconnaissance work along the northwestern rim of the Bowser Basin, British Columbia. The sedimentary rocks there belong to the predominantly marine clastic Bowser Group as defined by Duffell and Souther (1964) and Souther (1972). Most of the Bowser Group seems to have been deposited during Late Jurassic

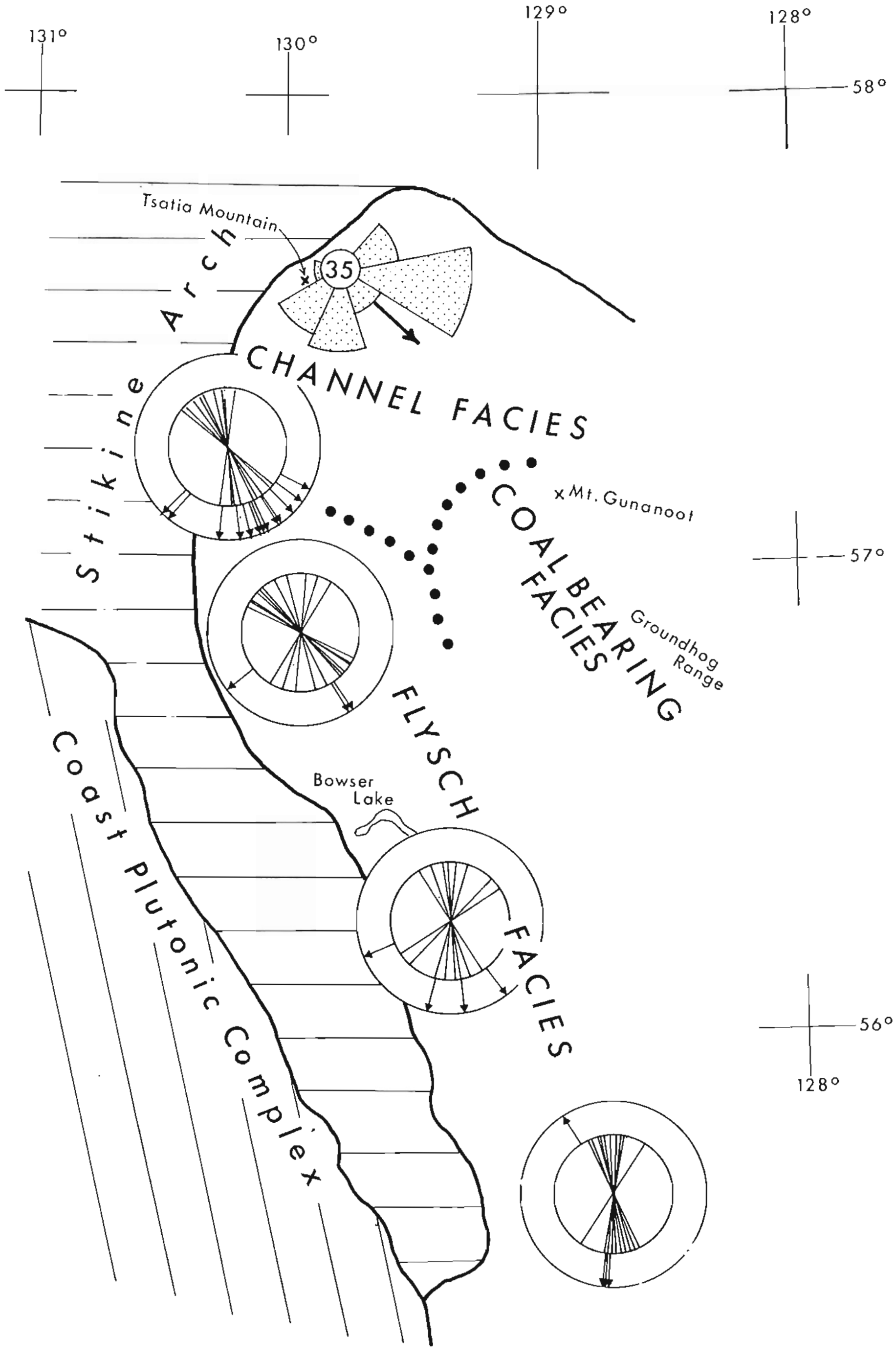
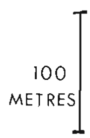
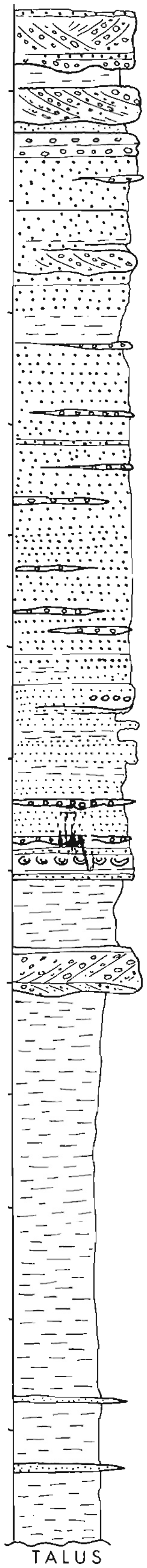


Figure 23.1.

Principal facies of Upper Jurassic – Lower Cretaceous Bowser Group, northwestern Bowser Basin. Direction of foresets of large scale cross-beds such as those shown in Figure 23.2 is plotted near Tsatia Mountain. Linear sole marks on sandstone beds of the 'flysch facies' are shown in four diagrams. Current directions are indicated as arrows where direction of transport could be ascertained unambiguously or as lines where direction of sediment could not be clearly established.





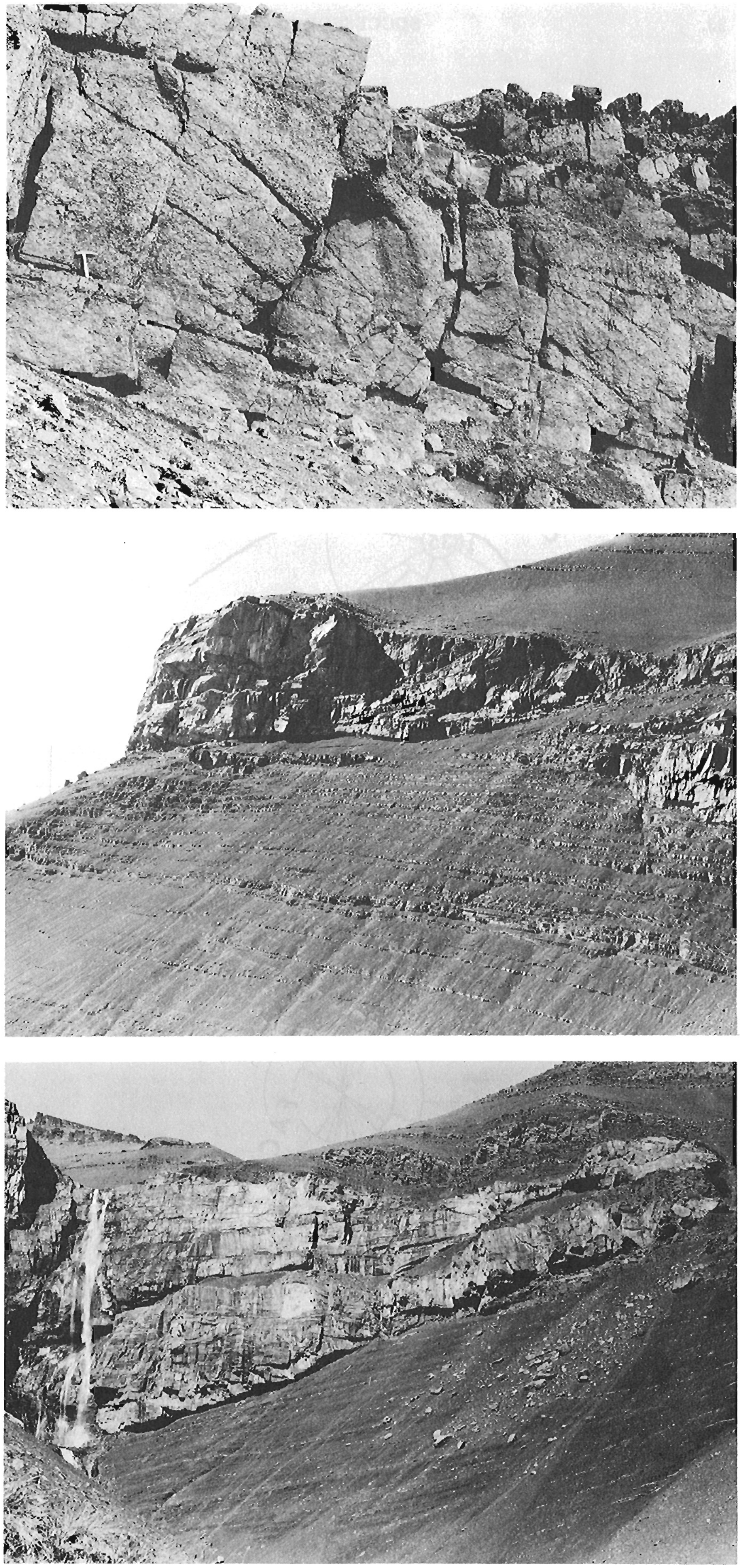


Figure 23.2

Measured section of Upper Jurassic Bowser Group on Tsatia Mountain (Fig. 23.1). Photos illustrate large lenticular channel systems comprising chert-pebble conglomerate with characteristic internal cross-bedding. These channel systems are interpreted as having formed along a rapidly accreting submarine slope.

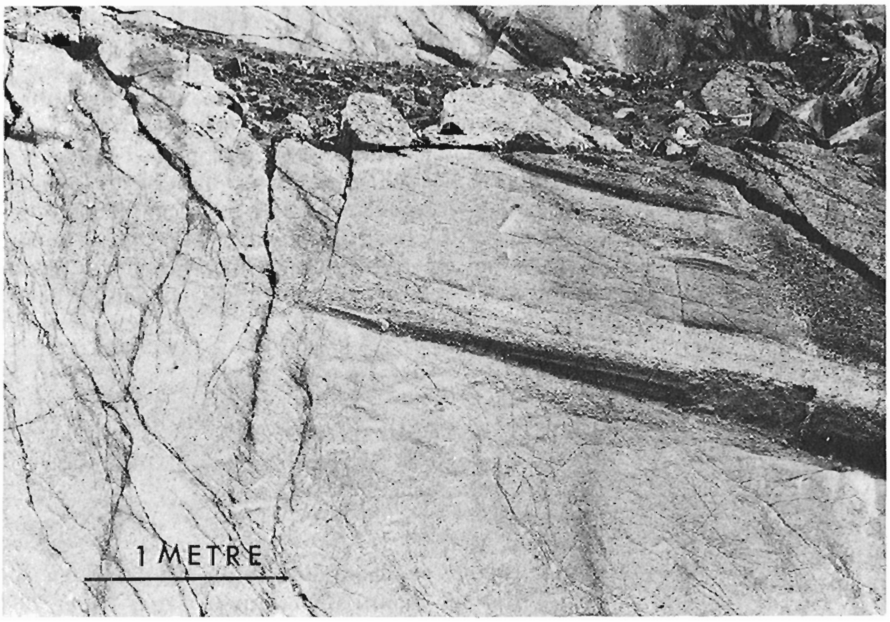
and earliest Cretaceous time although locally it may be older (Grove, 1971). It comprises lithic sandstone, dark siltstone and chert pebble conglomerate with only minor volcanics. It rests on and locally grades into predominantly volcanogenic sequences of the Cordilleran eugeosyncline.

Previous air photo interpretation and ground traverses during Operation Stikine (Geol. Surv. Can. Map 9 – 1957) revealed a bewildering array of folds along the northwestern margin of the sedimentary basin. Some of these fold systems intersect at angles up to 90 degrees. It was suspected that sedimentary facies changes and the geometry of the basin may have controlled the fold pattern, but the nature of the controls was not known.

Sedimentation

Three major, and probably diachronous, facies were recognized during the reconnaissance:

(a) Lenticular conglomerate channels, massive internally laminated sandstone and siltstone occur in a broad north-northeasterly trending belt. The conglomerate channels show large scale cross-bedding and cut deeply into fossiliferous siltstone and sandstone (Fig. 23.2). The abrupt lateral termination of many conglomerate channels suggests vigorous cutting, filling and dislocation of a submarine channel system. Individual foreset strata within the conglomeratic units, which in places exceed 5 m in length, indicate only the direction in which a previously eroded channel was



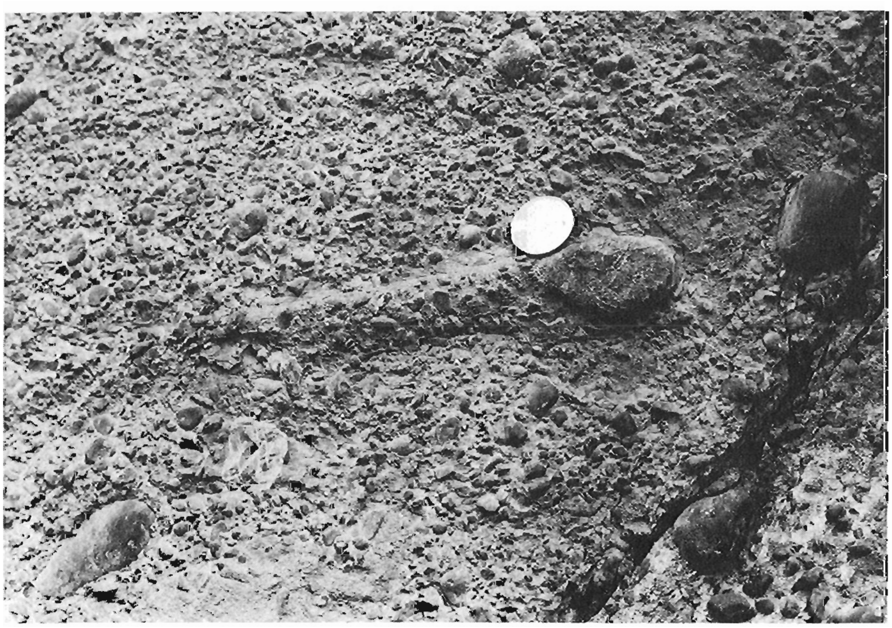


Figure 23.3.

Sole marks at the bottom of conglomeratic deposits carved by stones dragged along the base of submarine grain flows or turbidites. Current direction in both photographs is from left to right.

filled (Fig. 23.2); their mean direction indicates a southeasterly dipping paleoslope (Fig. 23.1). Local angular discordances imply syndepositional tilting of the sedimentary pile. The finer grained rocks between and along trend with the conglomerate channels are arranged in thick amalgamated sandstone bodies with a distinct internal lamination on the scale of centimetres or millimetres. Shell fragments and plant hash are generally found with pebbly lenses scattered throughout these sandstone units.

A section 1400 m thick was measured on Tsatia Mountain (Fig. 23.2). Generally the section coarsens upward and the southeasterly directed channel flow suggests that this coarsening may be related to the basinward progradation of the shelf. This "channel" facies is similar to another facies belt along the northeastern margin of the Bowser Basin (Eisbacher, 1974a). It probably represents a relatively steep submarine slope cut by several shifting channels down which most of the coarser clastics were transported into deeper parts of the basin.

(b) Conglomerate sheets, graded sandstones, and shale underlie a large area east and west of the Eddontenajon – Stewart – Terrace road system. In many sections and road-cuts the sequence and character of sedimentary structures within sandstones and conglomerates suggest deposition by turbidity currents and debris flows. Sole marks are well exposed at several localities along the Stewart-Cassiar road (Fig. 23.3). In strong contrast with the conglomerates of the "channel" facies, the conglomerates of the "flysch" facies do not display large scale internal cross-bedding. This facies extends to the westernmost outcrops of the Bowser Group (Grove, 1971). Judging from the limited number of solemarks plotted in Figure 23.1, the basin axis may have had a southerly trend, parallel with the present erosional margin of the Bowser Basin.

(c) A sequence of cross-bedded conglomerate and sandstone, brown siderite concretions, siltstone and thin coal seams is younger than the two previously described facies. This "coal-bearing" facies is coarse in the north (Mount Gunanoot) and finer grained towards the centre of the basin (Groundhog Range). It probably is the record of a southerly prograding delta system that covered the basinal flysch and channel facies.

Structure

Folds in the area are closely spaced and generally cannot be followed far along trend. Three types of folds were found in the region:

(a) Northeast and north-northeast trending folds with no definite direction of vergence.

(b) Northwest trending folds overturned to the northeast.

(c) Northwest trending folds overturned to the southwest.

The northeast trending folds are parallel with the trend of the paleoslope for the "channel" facies. Because the northeast trending folds are essentially confined to the channel facies they may have formed in response to gravitational movements down the basin slope contemporaneous with or shortly after clastic deposition. The northeasterly trending folds are clearly overprinted by younger northwesterly trending northeastward verging folds. This later overprinting was caused by Late Cretaceous – Eocene uplift within the Coast Plutonic Complex possibly accompanied by some crustal shortening of the basin (Eisbacher, 1974b). The significance and relative age of the northwesterly trending and southwesterly overturned folds has not yet been determined.

References

Duffell, S. and Souther, J.G.
1964: Geology of Terrace map-area; Geol. Surv. Can., Mem. 329, 117 p.

Eisbacher, G.H.
1974a: Deltaic sedimentation in the northeastern Bowser Basin; Geol. Surv. Can., Paper 73-33, 13 p.

1974b: Sedimentary history and tectonic evolution of the Sustut and Sifton Basins, north-central British Columbia; Geol. Surv. Can., Paper 73-31, 57 p.

Grove, E.W.
1971: Geology and mineral deposits of the Stewart area, B.C.; B.C. Dep. Mines Pet. Resour., Bull. 58, 219 p.

Souther, J.G.
1972: Telegraph Creek map-area; Geol. Surv. Can., Paper 71-44, 38 p.

PROTEROZOIC RAPITAN GROUP AND RELATED ROCKS, REDSTONE RIVER AREA, DISTRICT OF MACKENZIE

Project 750014

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Introduction

Large areas of the Mackenzie Mountains are underlain by Proterozoic sedimentary rocks. In view of their importance as host rocks for iron, copper, lead, zinc and possibly other metals, a study of the Proterozoic stratigraphy, tectonic framework and sedimentology was begun during the field season of 1975 in the Redstone River region (Fig. 24.1). This area has been mapped in reconnaissance by Gabrielse *et al.* (1973).

Stratigraphic setting and terminology

The stratigraphic terminology of Proterozoic rocks in the Canadian Cordillera has been discussed by Gabrielse (1972) and Aitken *et al.* (1973). These discussions have clearly demonstrated that in the northern Cordillera the use of the time-stratigraphic terms Helikian (for rocks below the base of the Rapitan Group) and Hadrynian (for rocks above the base of the Rapitan Group) is based on little evidence as to the time of deposition of the Rapitan Group and formations immediately above or below it. Until such evidence is forthcoming the use of the terms Hadrynian and Helikian for Proterozoic strata of the Mackenzie Mountains seems to be unwarranted.

In the Redstone River area the Proterozoic sequence is locally up to 7000 m thick and has been subdivided into a number of formations (Gabrielse *et al.*, 1973). The present author has added three informal names for the mappable subdivisions of the Rapitan Group (Fig. 24.2).

The Proterozoic record of the area studied consists of four major depositional cycles, each commencing with deposits of a deepening basin and ending with gradual shoaling of the basin margin. In the first cycle a sequence of shale and siltstone (Tsezotene Formation) grades upward into shallow-water quartzites (Tigonankweine Formation). In the second cycle basinal carbonates (Little Dal Formation) grade upwards into redbeds and evaporites (Redstone River Formation). In the third cycle basinal carbonates (Coppercap Formation) are overlain by "Maroon" siltstone, "Diamictite", and "Shale" formations (Rapitan Group) which grade upward into shallow-water quartzite and carbonate (Keele Formation). The fourth cycle begins abruptly with an extensive blanket of shale (Sheepbed Formation) that grades upward into shallow-water sandstone and carbonate of sub-Cambrian to Cambrian (?) age (Backbone Ranges Formation).

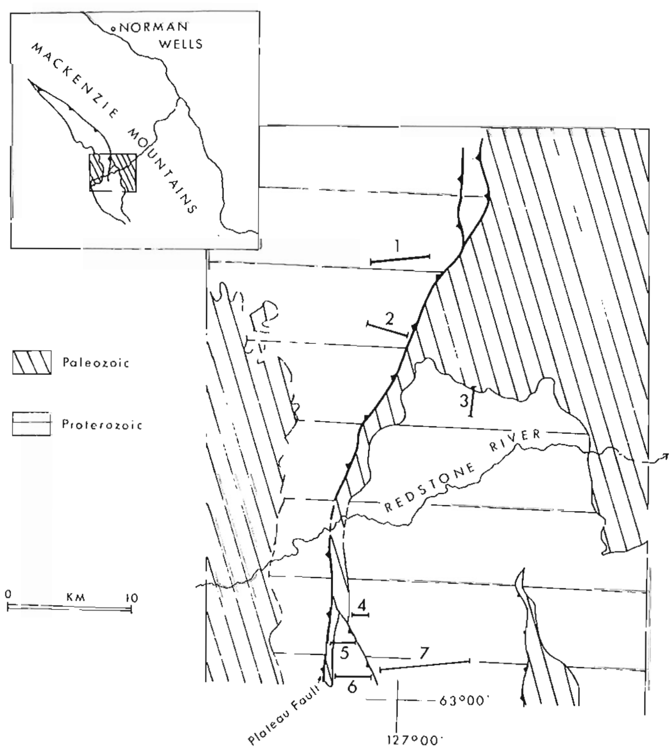


Figure 24.1. Index map of the Redstone River area and location of measured sections in Proterozoic sedimentary strata.

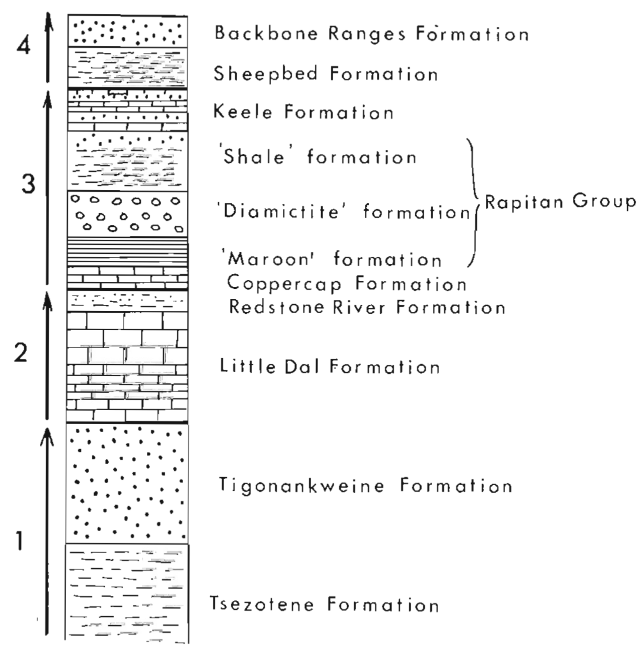


Figure 24.2. Proterozoic stratigraphy of the Redstone River area, Mackenzie Mountains, N.W.T. The four sedimentary cycles are indicated by arrows.

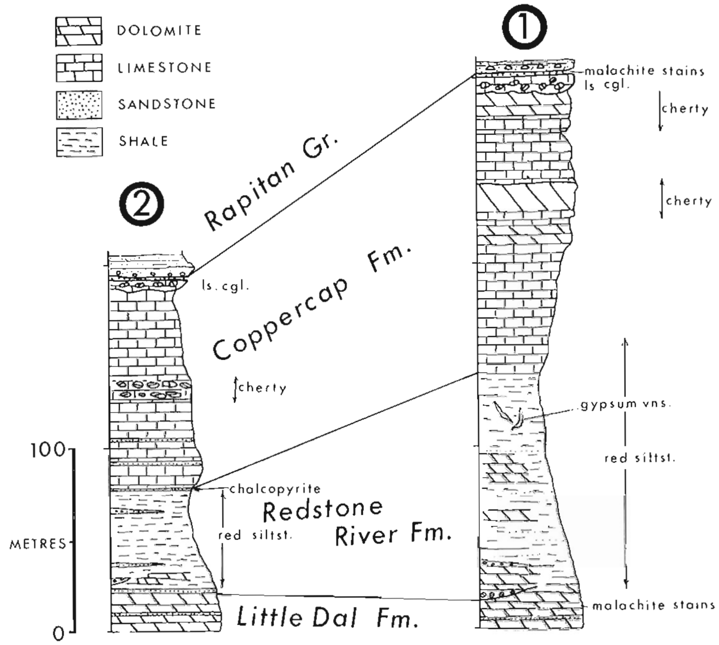
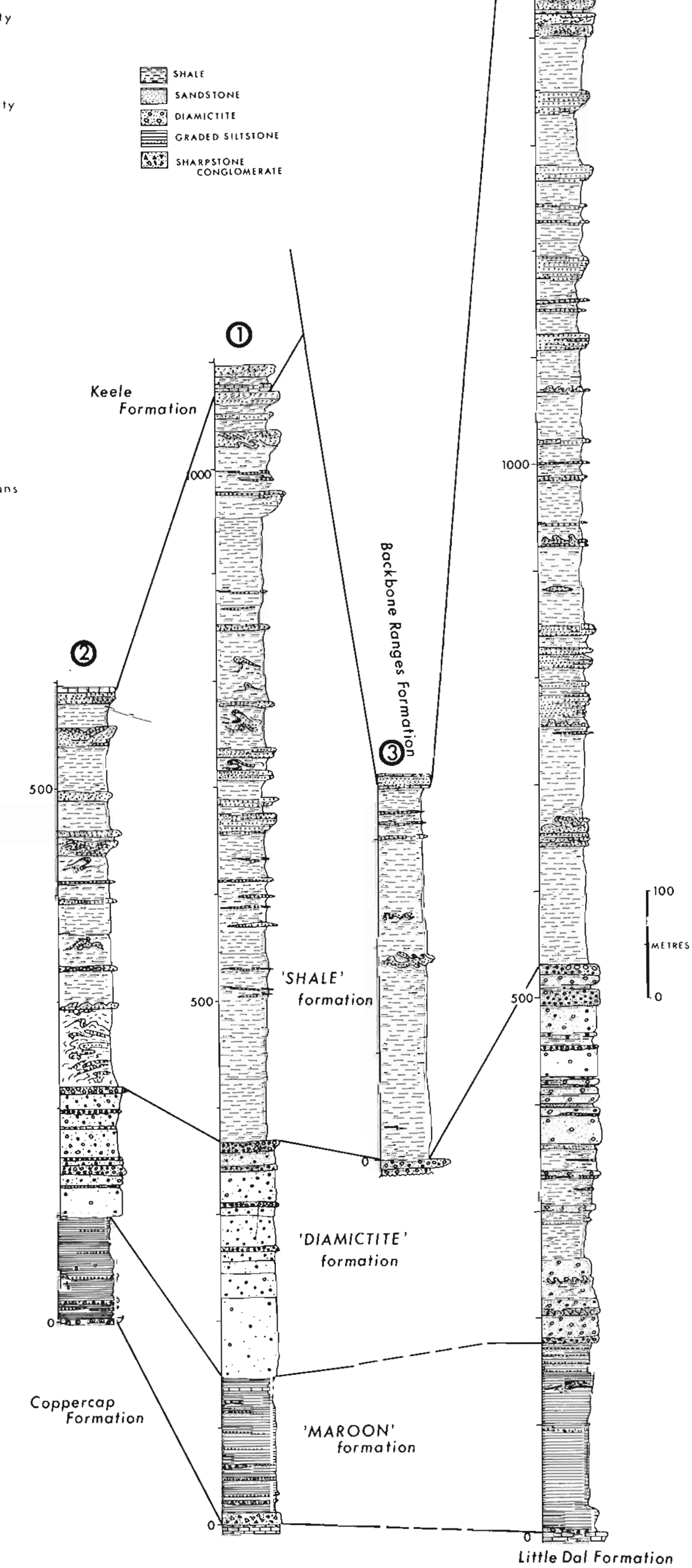


Figure 24. 3. Columnar sections of Redstone River and Coppercap Formations (location shown in Fig. 24. 1).

Figure 24. 4

Columnar sections of Rapitan Group. Locations are shown in Figure 24. 1.



Lateral thinning and facies changes are common throughout the succession and apparent unconformities are due to rapid thickening down-basin or lateral facies changes along a complicated miogeoclinal hinge. An economically important contact within the sequence is near the base of the third cycle where the copper-bearing Coppercap Formation underlies the Maroon Formation of the Rapitan Group. This contact has been interpreted as a major unconformity (Gabrielse *et al.*, 1973; Aitken *et al.*, 1973) but it may be questioned whether the Redstone River and Coppercap formations, where missing, had been eroded prior to deposition of the Rapitan Group in these areas or whether they are simply lateral facies equivalents of the uppermost Little Dal Formation.

Redstone River Formation and Coppercap Formation

The formations have been previously described by Gabrielse *et al.* (1973) from the Glacier Lake map-area. A long, narrow hitherto unrecognized outcrop belt of Redstone River and Coppercap formations containing copper mineralization was discovered north of Redstone River (Geol. Surv. Can., Open File 298 and Fig. 24. 3). The base of the Redstone River Formation has been placed at the base of the first distinct red mudstone unit, overlying massive dolomite, laminated grey dolomitic siltstone, and cross-bedded quartzite of the Little Dal Formation. Interbedded with the redbeds are beds of brownish weathering dolomitic grit and sharpstone conglomerate and buff mottled to nodular dolomite in a matrix of red siltstone. The grit layers are 3-10 cm thick and are commonly graded. Upwards this sequence grades into red siltstone and mudstone with scattered

Figure 24.5

Sharpstone channels in the Maroon formation of the basal Rapitan Group.

A. Coarse debris of limestone, dolomite and quartzite filling a scoured channel cut into finer grained sands.

B. Distinctly imbricated sharpstone pocket in laminated siltstone. Current direction from left to right.

veins of gypsum. The contact between the red mudstone of the Redstone River Formation and basal limestone of the Coppercap Formation is invariably recessive. Basal beds of the Coppercap Formation are either fetid, dark grey, thinly bedded and internally laminated limestones or buff, medium-bedded limestones commonly separated by grey-green siltstone containing sporadic copper mineralization. Most of the dark grey limestone is detrital and either graded or distinctly laminated. Towards the top it is interbedded with light laminated dolomite and replacement chert. The uppermost beds are lensoid bodies of limestone pebble conglomerate, which in turn are overlain in sharp contact by maroon sharpstone conglomerate and green or red siltstone of the Rapitan Group.

The sedimentary succession of the Redstone River Formation and the Coppercap Formation records shoaling along a basin margin followed by renewed submergence which continued with the deposition of the Rapitan Group.

Rapitan Group

The lithologic character of the Rapitan Group signifies an abrupt change in the regional sedimentary pattern. Previous workers in the Mackenzie Mountains have recognized a threefold subdivision of the Rapitan Group (Gabrielse *et al.*, 1973; Aitken *et al.*, 1973). The author refers to these three units informally as "Maroon" formation, "Diamictite" formation, and "Shale" formation.

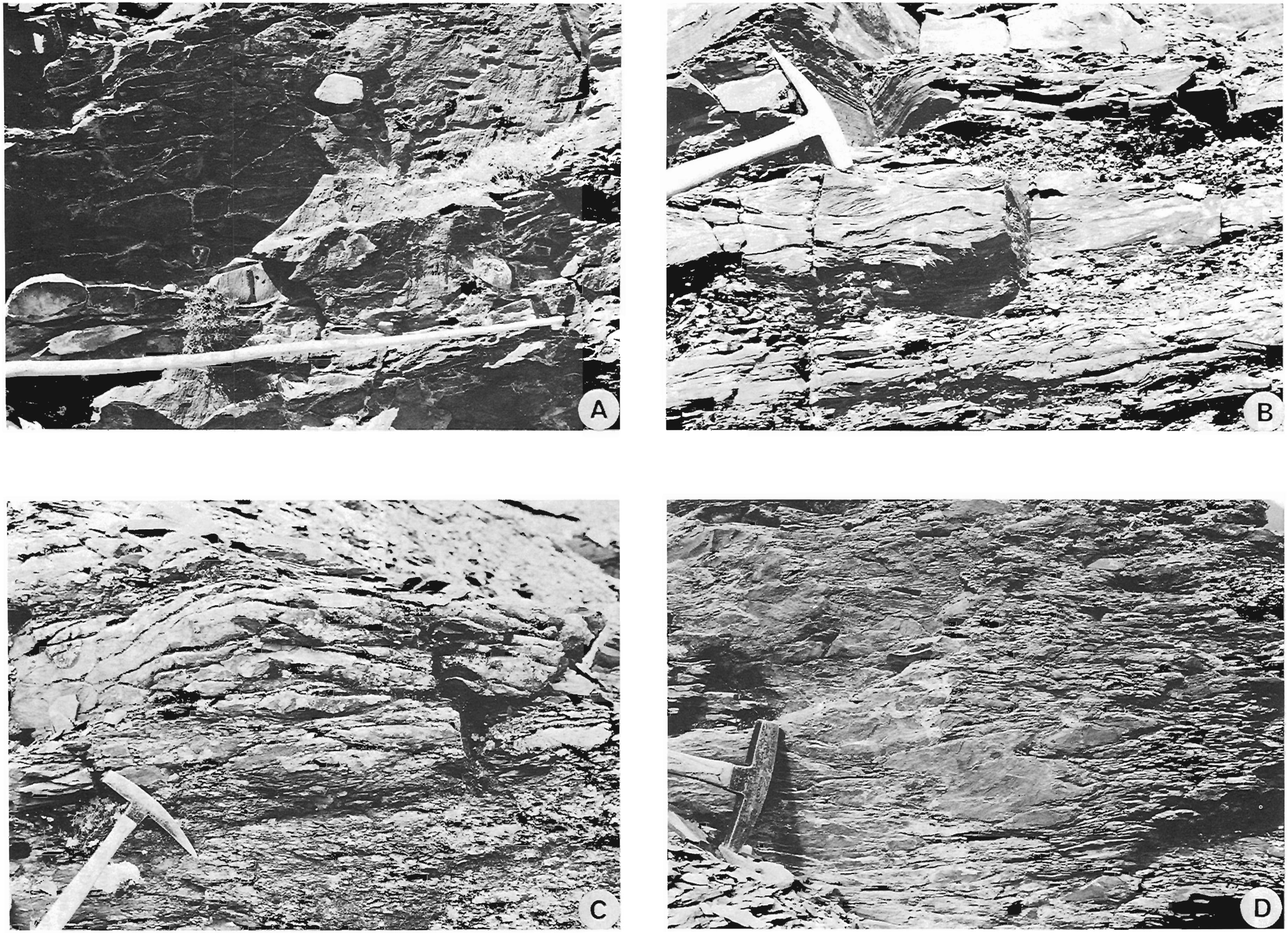


Figure 24.6. (A). Loosely dispersed diamictite with characteristic scaly matrix. The matrix ranges in composition from (B) ripple-drift cross-laminated siltstone and interbedded shale to (C) smeared out siltstone layers in a finer grained scaly matrix or (D) poorly structured pods of siltstone in a scaly matrix. The groundmass of the diamictite seems to have behaved plastically shortly after deposition of the sediment (Pole in Fig. 24.6A is 1.5 m long).

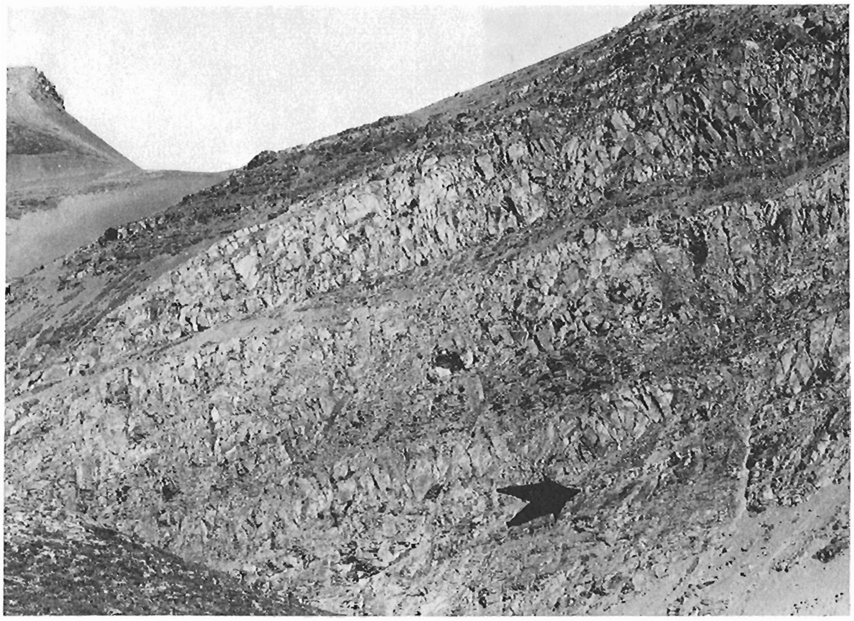


Figure 24. 7

Three distinct sheets of diamictite separated by shale or sandstone lenses. The picture shows the uppermost part of the Diamictite formation and the recessive basal part of the Shale formation of Section 4. The layer of quartz-sandstone shown in close-up in Figure 24. 8 and discussed in the text, is indicated by the arrow.

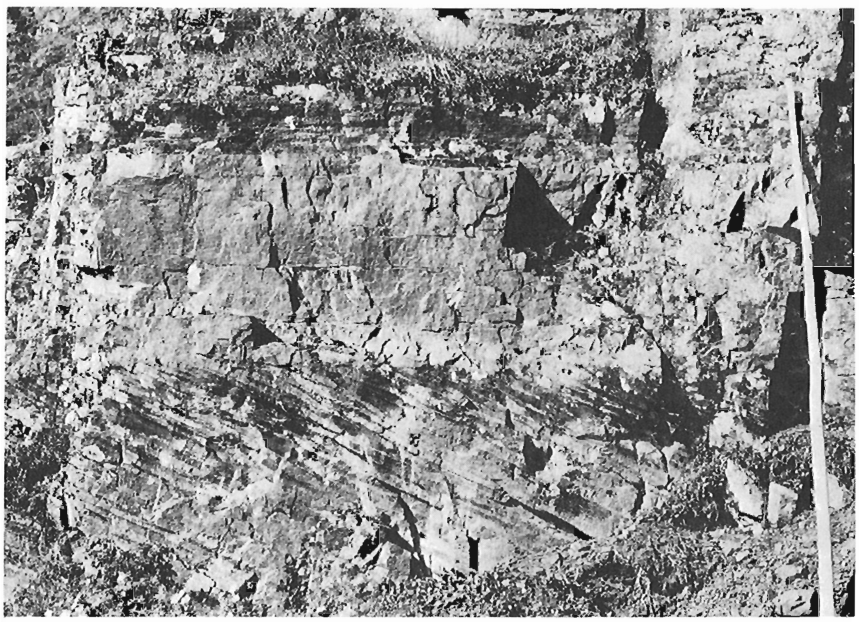


Figure 24. 8

Quartzose sandstone unit within the Diamictite formation of Section 4. Note large-scale planar foresets of cross-bedded interval (pole is 1. 5 m long). This sandstone is interpreted as a relatively shallow water deposit.

a) Maroon formation

The basal contact of the Maroon formation is sharp and defined by a striking change in composition and rock colours. In the area near Redstone River the Maroon formation overlies limestone-pebble channels of the Coppercap Formation or massive dolomitic rocks of the Little Dal Formation (Fig. 24. 4). Its basal member is a sequence of green, dark red or maroon graded siltstone, graded lithic arenite, and lenticular sharpstone conglomerate. The lithic arenites display many features typical of turbidites, such as graded bedding, ripped-up shale clasts, and the Bouma sequence of internal sedimentary structures. Graded sharpstone channels comprise limestone, dolomite, quartzite and minor greenstone clasts which are commonly imbricated (Figs. 24. 5A, 5B). In the central part, the most common sediments are graded rhythmites consisting of siltstone laminae 5 to 10 mm thick, with thin partings of dark red argillite. Ripple-drift cross-lamination occurs sporadically throughout. Iron-formation, common in other parts of the Rapitan basin, is found locally as thin beds within fine clastics. Towards the top, graded lithic arenites interbedded with maroon siltstone and dark red argillite increase in abundance. The graded arenites are intricately amalgamated with the underlying argillite and bedding planes are generally not exposed. Where coarse sediments overlie finer red silts many forms of flame

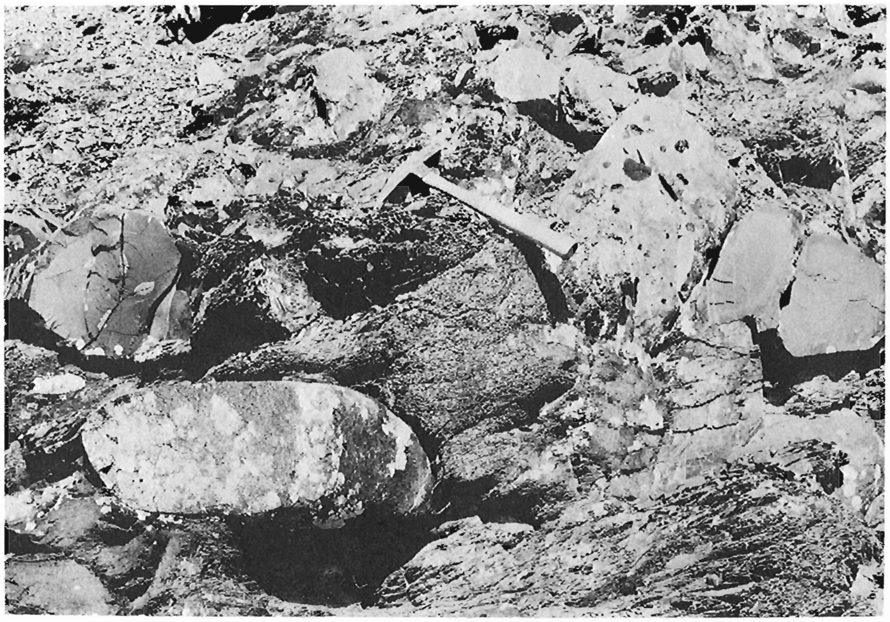


Figure 24.9

Densely packed diamictite near the top of the Diamictite formation of Section 4. Clasts are diorite (lower left), greenstone (right and left centre), and limestone (right of pick handle).

Figure 24.10

Slumped shale and siltstone of the Shale formation, Rapitan Group. Slumped units are as much as 50 m thick.

structures, folds, and micro slumps mark the passage from one bed to the next. The thickness of the Maroon formation ranges between 50 and 200 m in areas where clastic beds are fine grained. Sections with coarse sharpstone channels are thicker.

The abrupt change from carbonate to clastic deposition at the base of the Maroon formation was most likely caused by tectonic accentuation of the basin margin, rapid subsidence, and possibly cooling of the climate. The angularity and composition of clasts suggests that most of the clastic material was derived locally, perhaps from submarine cliffs.

b) Diamictite formation

This unit of the Rapitan Group was first described and interpreted by Ziegler (1960) in the Snake River region as a glacial-marine deposit (Snake River Tillite).

The base of the Diamictite formation is generally sharp and is marked by the appearance of well rounded greenstone boulders and a colour change from red-maroon to green. In several places isolated rounded greenstone clasts appear below sharpstone channels and light maroon siltstone bands reoccur within the lower part of the Diamictite formation. The top of the unit is sharp and well defined by recessive shales and

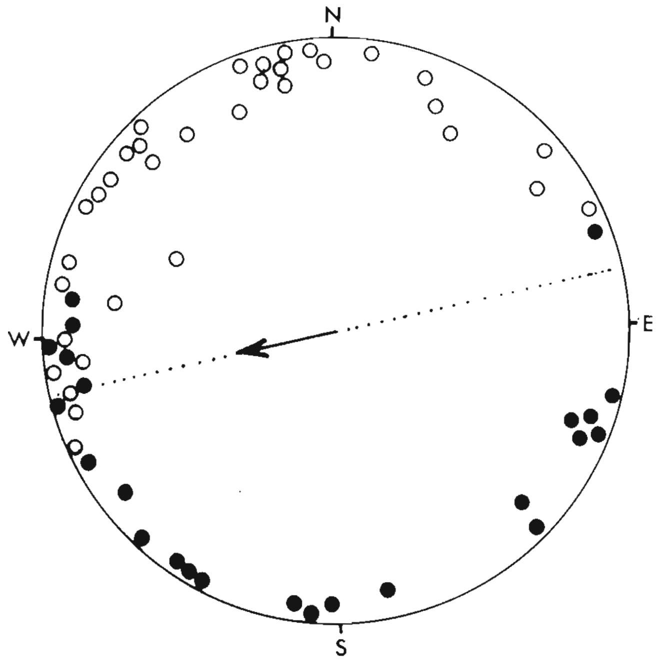


Figure 24.11. Stereo-plot of slump-fold axes with sense of short-limb rotation (dots signify clockwise rotation, circles counterclockwise rotation). Line of separation indicates a westerly directed mass transport and paleoslope.

siltstone of the overlying Shale formation. In the Redstone River area the Diamictite formation is between 130 and 350 m thick. A strong facies change is apparent between sections east and west of the Plateau Fault. East of Plateau Fault distinct members of diamictite (densely, loosely or sporadically dispersed clasts within a silty matrix) can be separated from shale and sandstone members (Fig. 24.4), whereas to the west the sequence is more uniform and mainly composed of diamictite with rather diffuse borders. Only local pebble pockets interrupt the monotonous lithology.

Several features of the Diamictite formation are striking in section and outcrop. The clasts, ranging from pebbles to boulders, show only a weak preferred orientation with respect to bedding, and clasts "standing on edge" are fairly common. Locally densely packed pockets or clusters of clasts with deep load casts at their basal contact occur within loosely dispersed diamictite. The clasts, in contrast to those of the sharpstone channels in the Maroon formation below, are well rounded to subrounded, and consist predominantly of greenstone, with smaller amounts of dolomite and limestone. The silty matrix of the loosely dispersed diamictite may range from composite, cross-laminated siltstone beds to smeared out lenses of siltstone in a "scaly" matrix (Fig. 24.6, A to D). East of Plateau Fault several thin sandstone members are interbedded with diamictite (Fig. 24.7 and 24.8). One unit of sandstone which can be traced for several kilometres along strike shows large scale cross-bedding (1.5 m). Of particular interest in the interpretation of the diamictites are sparse pebbles with surface striations characteristic of glacial polish. Most striations observed are, however, due to minor tectonic shear between matrix and clasts, as described previously from Huronian diamictites in northern Ontario (Eisbacher, 1973). The largest clasts (up to 160 cm in diameter) occur invariably close to the top of the formation in densely packed units (Fig. 24.9). These units, in turn, are overlain in sharp contact by shale above which no more clasts are found.

The author, with Ziegler (1960), favours a glacial-marine environment for the deposition of the Diamictite formation. The analogy with the Plio-Pleistocene Yakataga Formation of coastal Alaska, as drawn by Ziegler, may indeed have merit in understanding the setting of the Diamictite formation. An interpretation in terms of non-glacial mud- and debris flows as reported from orogenic flysch basins around the world seems unlikely because the associated sandstones do

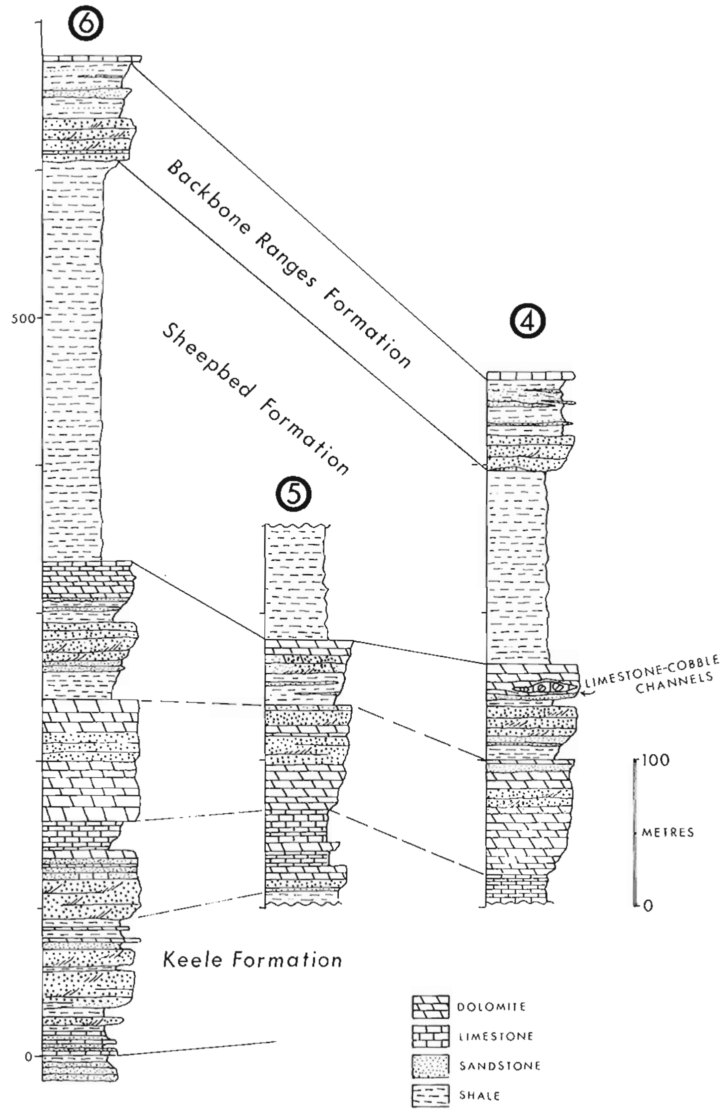


Figure 24.12. Columnar sections of the Keele Formation south of Redstone River (*see* Fig. 24.1 for location). Note rapid facies and thickness changes.

Figure 24.13

A. Laminated limestone of the Keele Formation. Some of the lamination is possibly of organic origin.

B. Pure quartz arenite of the Keele Formation. Distinct 'herring-bone' arrangement of cross-beds indicates shallow water depositional environment, possibly migrating tidal bars (pole is 1.5 m long).

not show turbidite or debris flow characteristics. In contrast, they are well sorted quartzose arenites with local large-scale cross-bedding. The fact that the coarsest boulder diamictites (containing greenstone, diorite, limestone and dolomite) occur at the very top of the formation and that they are overlain by a thick shale unit without clasts could be explained by assuming a rapidly decaying ice mass at the end of a glacial period. At this stage icebergs larger than usual would break away from glaciers and carry coarse morainal debris with them. Following complete melting of the floating ice masses eustatic rise in sea level would reinstate normal deposition of clay.

c) Shale formation

Shale and thinly laminated siltstone overlies the coarse diamictite described above in generally conformable contact. Apparent unconformities reported previously from the Redstone River area (Gabrielse *et al.*, 1973) are probably due to mass flow and sliding visible in several sections (Fig. 24.4). Higher in the section thin- to medium-bedded sandstones occupy intervals of 5 to 20 m. Characteristic sedimentary structures associated with these sandstones are load casts, parallel or ripple-drift lamination, and well developed current ripple-marks on bedding surfaces. Slump folds and other indications of down-slope mass movements are common throughout the section and account for internal stratigraphic discordances (Fig. 24.10). In the uppermost part the quartzose sandstones become coarser, thicker bedded, and display distinct cross-bedding on a scale of about one metre. The total thickness varies from about 400 to 1000 m.

Within the Shale formation the distribution of sandstone beds is not rhythmic or monotonous. Beds of sandstone with parallel laminations occur in broadly lensoid packets 5 to 30 m thick, with bedding thickness commonly decreasing upwards.

It is probable that most of the sandstone units within the Shale formation were deposited within widely meandering channels on a subtidal shelf and that individual layers were reworked by ocean currents. Deposition by turbidity currents was probably not significant. Slumping indicates that the shelf region was unstable and measurement of slump folds indicates a westerly directed paleoslope (Fig. 24.11). Since the Shale formation grades upward into carbonates and channeled quartz arenites of the Keele Formation, the basin was probably progressively filled by sediments of shallow water origin in the youngest parts of Rapitan depositional cycle.

Keele Formation

The Keele Formation in the Redstone River area consists of a heterogeneous suite of grey, laminated limestone, laminated or recrystallized dolomite, cross-bedded quartzite and minor shale (Fig. 24.12). The sequence seems to be made up of several cycles shoaling upward and grading from fine siltstone or limestone to quartzite or dolomite (Fig. 24.13, A to C). Within the highest member of the Keele Formation an orange weathering laminated dolomite is interbedded with a few channels of grey limestone cobble conglomerate. The cobbles of these conglomerates are similar to limestone of the Keele Formation about 100 to 200 m stratigraphically below. The reworking of older Keele units into conglomerate higher in the section suggests a very active basin hinge during deposition.

Inferred geological history of the Rapitan depositional cycle

The Rapitan depositional cycle in the Redstone River area began with subsidence of a carbonate shelf underneath a rapidly accumulating sequence of sharpstone conglomerate channels, maroon siltstone laminites, and turbidites. During a subsequent glacial phase coarse debris was carried offshore by icebergs resulting in the deposition of glacial-marine diamictites interstratified with nearshore quartzarenite. After wasting of the glaciers an unstable shelf sequence of clay, sand and carbonate gradually filled the margin of the Rapitan basin until carbonate sedimentation again became important.

References

Aitken, J.D., Macqueen, R.W., and Usher, J.L.
1973: Reconnaissance studies of Proterozoic and Cambrian stratigraphy, lower Mackenzie River area (Operation Norman), District of Mackenzie; Geol. Surv. Can., Paper 73-9, 178 p.

Eisbacher, G.H.
1973: Deformationsmechanik eines offenen Konglomeratgefüges; Tschermaks Min. Petr. Mitt., v. 20, p. 249-256.

Gabrielse, H.
1972: Younger Precambrian of the Canadian Cordillera; Am. J. Sci., v. 272, p. 521-536.

Gabrielse, H., Blusson, S.L., and Roddick, J.A.
1973: Geology of Flat River, Glacier Lake, and Wrigley Lake map-areas, District of Mackenzie and Yukon Territory; Geol. Surv. Can., Mem. 366, 153 p.

Ziegler, P.A.
1960: Frühpaläozoische Tillite im östlichen Yukon-Territorium (Kanada); Eclog. Geol. Helvetiae, v. 52, p. 735-741.

NEPHRITE JADE OCCURRENCES IN BRITISH COLUMBIA AND YUKON TERRITORY

Project 490038

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During the 1975 field season, two months were spent completing investigations of nephrite jade occurrences in British Columbia and Yukon Territory. Visits were made to active properties, one property was examined for the first time, and studies were made of serpentinite belts in Campbell Range, Yukon Territory, Cry Lake map-area and the Shulaps Range in British Columbia. In addition, some new locations for other lapidary materials were examined.

Annual production of jade in British Columbia is about 320 tons from the following producers (*see* Fig. 25.1):

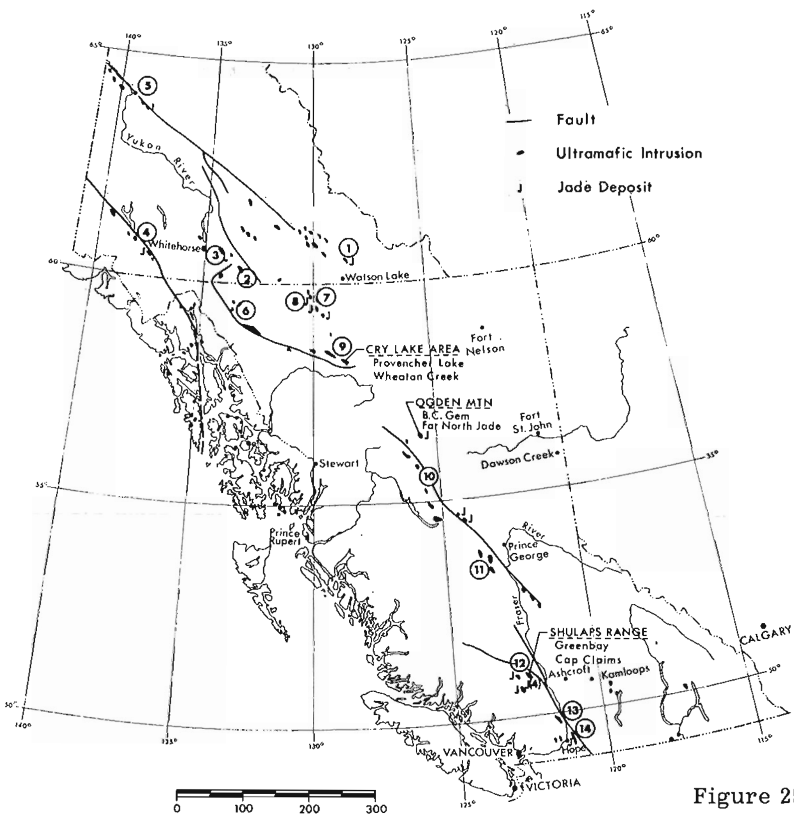
Greenbay Mining	200 tons
Far North Jade - Provencher Lake Property	50 tons
Ogden Mountain	50 tons
Cap Claims	15 tons
Minor other producers	5 tons
Total	320 tons

Minor production of possibly 5 tons comes from recovery of small boulders along Fraser River, and in alluvial deposits at Cassiar, Wheaton Creek and the Omineca area.

The Provencher Lake deposit in the Cry Lake map-area was visited for the first time. Production over the past three years has been entirely from alluvial boulders which lie singly or in clusters along the valley floor. Erosional blocks close to in situ deposits lie higher up the valley slopes, but these have not been exploited. The company estimates that more than 600 boulders contain 10 000 tons of raw jade of variable quality, including unsalable material. The boulders range from 200 pounds to more than 40 tons, with an average of about 5 to 6 tons. Few of the smallest boulders contain commercial material because of fractures or poor colour.

Five men were employed in 1975 to core-drill all boulders which appear to be sound, homogeneous and good colour. Boulders which reveal good material in the cores are marked for removal to Dease Lake by crawler tractor trains after freeze-up. There, wire saws are used to cut the boulders into blocks small enough to be cut by diamond saws. The blocks are then trucked to Vancouver.

Four in situ occurrences of jade were seen on the property. Three are on the crest of ridges east of the lake, at elevations of about 6500 feet. The fourth occurrence lies about one-quarter mile west of the north end of the lake about 100 feet above the lake level. The



PRELIMINARY ESTIMATES OF JADE RESOURCES IN YUKON - BRITISH COLUMBIA

AREA	TONNAGE (all grades) PROVED	PROBABLE	POSSIBLE
YUKON			
(1) CAMPBELL RANGE		100	500
(2) TESLIN AREA		50	100
(3) WHITEHORSE AREA		100	300
(4) KLUANE AREA		50	100
(5) DAWSON AREA		50	100
MISCELLANEOUS		50	100
Sub-total		**400**	**1,200**
BRITISH COLUMBIA			
(6) ATLIN		50	150
(7) CASSIAR	100	2,000	5,000
(8) DEASE LAKE	10	150	500
(9) CRY LAKE	50	10,000	20,000
(10) OMINECA	100	2,000	10,000
(11) PRINCE GEORGE			100
(12) SHULAPS RANGE	50	1,000	2,000
(13) FRASER BELTS		100	500
(14) COQUIHALLA RANGE		50	200
Sub-total	**310**	**15,350**	**38,450**
TOTAL	**310***	**15,750**	**39,650**
Saleable Jade (approx. 10%)		1,600	4,000

* Does not include inventory held by major producers in warehouses in Vancouver, Chilliwack and Lillooet.

Figure 25.1

deposits are typical of nephrite jade occurrences elsewhere in that they occur in serpentinite near the contact with metasediments of late Paleozoic age. The nephrite jade is associated with other contact reaction zone minerals — talc, diopside, and hydrogarnet. In one of the deposits, the jade blocks are clearly xenoliths which lie in random orientation within the serpentinite close to the metasediments. In the other deposits, the jade is restricted to the serpentinite — sediment contact and therefore still in the place of origin.

It is postulated that all the alluvial boulders came from erosion of local, in situ deposits, some of which may have been completely eroded or covered by alluvium. In Pleistocene time, some concentration of the boulders into trains and clusters on the valley floor was effected by glacial transport. Active erosion is presently slowly releasing blocks from the in situ deposits.

Possible reserves include the remaining undiscovered boulders in the valley floor, and recoverable parts of the in situ lodes now known only from outcrops.

A small amount of jade, not exceeding a few hundred pounds, was produced on placer leases along the lower part of Wheaton Creek. A search was made for in situ deposits from the mouth of Wheaton Creek south to the first tributary entering the creek from the east, but none were found. The large number and size of the boulders present suggest that they were locally derived. One in situ deposit occurs along Wheaton Creek about 4½ miles south of the confluence with Turnagain River. Others may be mantled by drift. Little or no jade was produced in the Dease Lake area in 1975. The principal occurrence is on the Seywerd claims at Sawmill Point on the north end of Dease Lake. The property was inactive in 1975. The quantities shown in Figure 25.1 include reserves on Thibert Creek and its tributaries.

The Ogden Mountain claims, formerly held by New World Jade Limited, were sold to B. C. Gem Supply (H. K.) Limited in 1974. The latter company took over production in 1975 and have been cutting jade from the remnants of material mined in previous years, and have been engaged in a program of exploration for further supplies. An X-ray diamond drill was used to probe the main contact reaction zone between serpentinite and Cache Creek Group sediments, the principal source of in situ jade mined in past years. Some production also came from alluvial boulders and residual blocks on or near outcrops.

Far North Jade Limited holds claims on Ogden Mountain northwest of the B. C. Gem claims. All past production came from alluvial material including one huge 80-ton block. No jade was produced in 1975. Stripping of overburden in several places was designed to locate in situ deposits, and alluvial boulders. Some contact reaction zones were revealed but it is not known if any jade was discovered.

Jade occurs in the Axelgold Range about 10 miles northwest of Ogden Mountain. The deposit was not visited but reliable information confirms that good quality nephrite has been found, although no information on the quantity is known. Specimens from the deposit are of good quality, typical dark green nephrite.

Mining of in situ jade on the Brett Creek property of Greenbay Mining resulted in production of 200 tons in 1975. The main occurrence is a vein-like lode developed with other contact reaction zone minerals, between a very large inclusion of chert and enclosing serpentinite. The vein is 1.5 - 3 feet wide, extends along strike for 80 feet and for as much as 30 feet down the dip. The jade has an outer talc rim against the hanging wall serpentinite with which it is in fault contact. Contact reaction minerals include carbonate, diopside, hydrogarnet and disseminated thulite.

Most of the mined material has been removed from the pit and only limited exposures are left to permit any appraisal of quality. Hence proved ore on the property is small.

The CAP claims owned by Mr. C. McEwen at the head of Jim Creek yielded about 15 tons of in situ jade in 1975. Work on a contact reaction zone between chert and serpentinite was started in 1974, but only a small sample was removed. Most of the jade is apple green with much magnetite (chromite and some chrome garnet ?). The full range of colour and texture has not yet been revealed.

Traverses on the ridge running west from Shulaps Peak and north of the CAP claims revealed two small "veins" of botryoidal jade of inferior colour. Although of mineralogical interest, they are of no commercial significance.

Traverses on the north side of the Shulaps Range near the Elizabeth Mine encountered some contact reaction zones. They contain a green mineral thought to be vesuvianite but no jade was seen.

The possibilities for the discovery of jade have not been exhausted. The Shulaps Range in general is highly rated as a potential jade source. Not only are numerous occurrences known and further discoveries likely, but the area offers the advantage of relatively easy access so that the costs of operating a camp and shipping the product is a fraction of that incurred in far northern areas.

Estimates of Ore Reserves

The estimation of jade ore reserves in the Canadian Cordillera (*see* Fig. 25.1), is attempted with the quantities expressed in the three categories — proved, probable, and possible.

Proved ore should include only those boulders or outcrops tested by core drilling or sawcuts sufficiently close that no unsalable material is included in the dimensions used to calculate the volume. Fully proved ore would include only rectangular, sawn blocks. In practice, one or more drillholes or sawcuts may be expected to allow a reasonable appraisal.

Probable ore may include all boulders and outcrops which may be recognized as jade in the field and which appear to be homogeneous, free of fractures and of good colour. Not all jade from boulders and outcrops is salable and therefore a reduction factor must be applied to calculations in this category. This factor may be as low as one per cent in some boulder fields, but perhaps as much as seventy-five per cent in others.

Possible ore includes jade expected to be found in alpine-type serpentinites with known or likely contact reaction zones, but without any proven jade occurrences. It includes a consideration of the number and areal extent of the serpentinite bodies shown on Figure 25. 1. The estimate is subject to the same reduction factor as with probable ore.

In the tables included in Figure 25. 1, the ratio of proved to probable ore is very small. This results from the low estimate of proved reserves, compared to the large number of known boulders and outcrops. The ratio might be raised by including a higher percentage of jade from the Cassiar stockpile which amounts to about 300 tons, held at the mine site. This material, although untested by drillholes or sawcuts, has been indirectly sampled by use of similar material at the McDame Lapidary Shop. Thus the figure of 100 tons considered in the proved category (Cassiar — Fig. 25. 1), is probably low.

A very attractive porphyry occurs along Copper River just east of Terrace. This rock is amygdaloidal and consists of reddish feldspar phenocrysts and green epidote-filled vesicles. The best collecting area is a small creek locally known as Kelly Creek, which flows into Copper River 17. 8 miles up the logging road along the south side of the river. The logging road may be found with reference to the Highway Bridge (Highway 16). It is one-half mile west of the bridge on the south side of the highway, about 6 miles east of Terrace.

Project 730069

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Field work in 1975 encompassed parts of several 1: 250 000 map-areas, in particular those regions underlain by black shale and chert of Selwyn Basin. Because of incompetent rock type and glacial drift cover, exposures are limited especially east and northeast of Ross River.

As far as can be determined, black shales in Nahanni (105 I), Sheldon Lake (105 J) and Tay River (105 K) map-areas are underlain everywhere by Cambro-Ordovician 'wavy-banded' silty limestone or its more argillaceous, transitional equivalents (Fig. 26.1, section A B). The black shales and associated strata comprise two main groups: 1. a lower group of early Ordovician to middle Devonian age assigned to the Road River Formation, the equivalents of carbonates in the Mackenzie Mountains to the east; and 2. an upper informal group known as the 'Black Clastic' group, containing much coarse clastic debris derived from the west and southwest. As outlined by facies changes in the lower shale group, the Selwyn Basin is more properly referred to as a trough, toward the axis of which flanking sediments comprising carbonaceous shale, shaly limestone and minor argillaceous calcarenite, change facies to grey, green and black ribboned chert and variegated shale of presumably deep water origin (*see* section A-B). Figure 26.1 shows in a general way the axis of this trough and the area of maximum chert development. In these rocks bedded zinc-lead deposits are conspicuously absent. By far the greatest amount of zinc and lead known is in two bedded deposits of about the same size, or groups of deposits, at two widely separated localities in carbon-rich shales flanking the trough. These include the Canex – Placer deposit of Howard's Pass and the Vangorda – Swim belt, both probably in rocks of Ordovician-Silurian age. Shales of similar lithology with unusual amounts of zinc are present just east of Ross River, at Fortin Lake and just northwest of Cantung. The area of potentially productive ground is large and until further geologic guides, in addition to stratigraphic position, are developed, lead-zinc exploration targets will necessarily be governed by access and availability of outcrop.

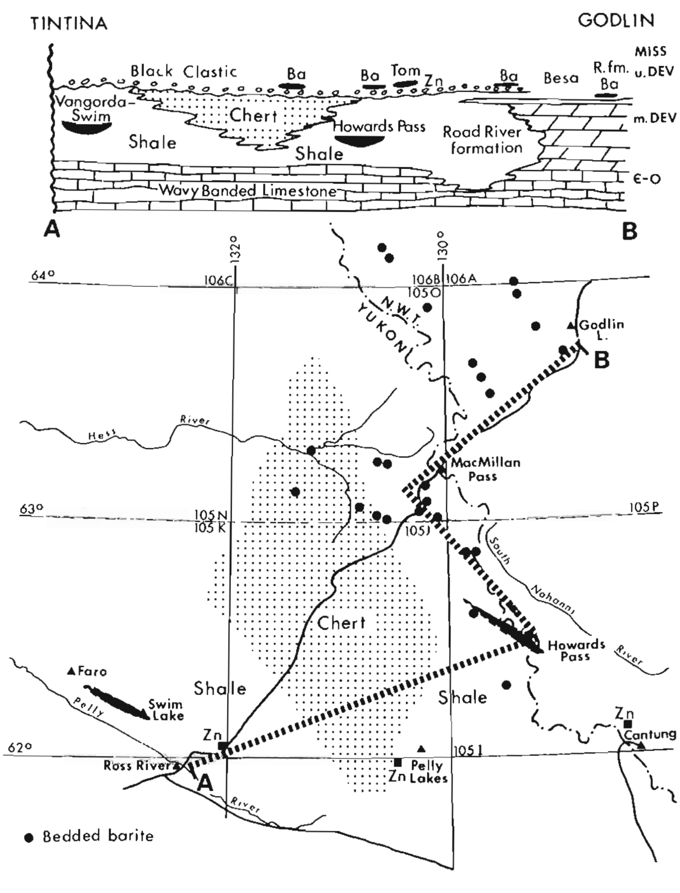

Figure 26.1. Sketch map showing generalized distribution of Ordovician – Silurian shale and chert facies (stippled), and locations of Upper Devonian bedded barite.

The upper shale division or 'Black Clastic' group extends across the region as far as Caribou Pass, where it correlates with the Besa River Formation of the Mackenzie Mountains. The Tom Property at Macmillan Pass, a significant deposit of bedded barite, lead and zinc, occurs in a distinctive silvery-weathering, coaly-black shale member within the lower part of the group. Similar looking and stratigraphically positioned laminated barite deposits are present in numerous places over a wide area extending from north-central Coal River map-area in the south, to at least the head of Mountain River in the north, and across a width of more than 100 miles (160 km). These deposits range in thickness up to 500 feet (150 m) or more, and in length to over 5 miles (8 km). Only a few have been examined for lead and zinc, as most are new finds. A unique feature of the Tom Property is an abrupt thickening of the host shale unit coincident with the limits of the deposit. The mineralization occurs near the base of the thickest, least laminated sequence of shale filling an east-west graben-like structure, suggesting an origin of the Red Sea type. A similar structure occurs a few miles northwest of the Macmillan River crossing, in north-central Sheldon map-area. Erosion has removed the 'Black Clastic' group throughout much of Sheldon and Lansing map-areas, probably accounting for the limited barite occurrences in these areas.

A helicopter-mounted scintillometer proved helpful in some instances for distinguishing main black shale units, provided constant relation to topography was maintained, preferrably in hover or at skid height.

In the course of this work no particularly radioactive localities were found in either of the two black shale groups, or in the Upper Proterozoic 'Grit' unit. Highest background readings came from red shales of the 'Grit' unit, whereas syenitic intrusive rocks of the Rogue Range in Central Niddery map-area, produced the highest readings. Of special interest are two localities in Ogilvie Mountains, which provided specimens of tuffaceous rocks anomalous in U_3O_8. The rocks were collected in a previous field season and no follow-up work has been done. The localities are 65°04'20"N, 134°14'30"W, and 65°02'40"N, 134°38'20"W. These rocks are associated with copper deposits in unit Ho of Norris, 1974, which extends throughout much of central Ogilvie Mountains as unit 1, Green, 1972, and into northwestern Nadaleen map-area as units H_{sc} and H_{cs} of Blusson, 1974. The unit consists predominantly of dark argillite, slate and phyllite with an abundance of mostly fine grained acidic volcanoclastic calc-silicate rocks in the lower part. The copper and uranium occur in fractures within the volcanic suites. These rocks are highly deformed, strongly cleaved and overlain unconformably by orange-weathering algal carbonates of probable Helikian age. There are no obvious correlative sequences of these argillites in the Mackenzie Mountains. They are either Aphebian in age or, more probably, represent a pelitic-volcanic facies of the oldest Halikian rocks in Mackenzie Mountains, the Katherine-Tsezotene clastic rocks. The unconformity at the base of the Rapitan Group, unit Hl_s of Blusson, 1974, limits the Helikian strata on the east, and on the south at the head of Rackla River. Rocks mapped as units 1 and 2 between Kathleen Lakes and the head of Rackla River in Nash Creek map-area, are now correlated respectively with the Hadrynian units Hl_s and Hl_{c1} of the Nadaleen map-area.

References

Blusson, S.L.
1974: Nadaleen River map-area of Operation Stewart; Geol. Surv. Can., Open File No. 205.

Green, L.H.
1972: Geology of Nash Creek, Larsen Creek and Dawson map-areas, Yukon Territory; Geol. Surv. Can., Mem. 364.

Norris, D.K.
1975: Wind River map-area, Yukon Territory; Geol. Surv. Can., Open File No. 279.

27. GEOPHYSICAL SURVEYS WEST OF VANCOUVER ISLAND, BRITISH COLUMBIA

Project 710049

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A multi-disciplinary program of systematic mapping of geophysical and geological parameters in the offshore region of British Columbia was continued in 1975. A hydrographic priority in the approaches to Strait of Juan de Fuca due to imminent deep draft tanker traffic set the area, and survey limits were defined as 48^{o}-49^{o}N, from coastal limit of navigation to approximately 127^{o}W. Gravity, magnetics and bathymetry were obtained in this region from CSS "Parizeau" in early spring. A second ship, the CSS "Vector" made use of the Minifix and Trisponder navigation systems to obtain surficial bottom sediment samples, continuous seismic profiles, and suspended particulate matter in the surface water in the same area. These latter data are reported by J. D. Milliman.

The magnetics program was continued north from 49^{o} to 50^{o} on the continental shelf by the Canadian Hydrographic Service during the summer, under Geological Survey supervision. An opportunity to obtain further magnetic and gravity data in this area was afforded in September, and a gravimeter was placed aboard "Parizeau" again for that purpose by Earth Physics Branch. The survey limits were extended off the continental shelf to approximately $128^{o}30'$ West. Preliminary maps of the 48^{o}-49^{o} area indicate a number of regions of interest:

1). The oceanic crust is well defined by characteristic linear magnetic stripes, some of which are present under the continental slope up to the shelf break.

2). A major continental type magnetic anomaly, the Prometheus anomaly (Shouldice, 1971; MacLeod *et al.*, in prep.) extends along the shelf parallel with Vancouver Island and with the trend of the Eocene Crescent volcanic belt on the Olympic Peninsula, but offset to the south of an extension of that belt. The Prometheus anomaly is a magnetic anomaly that does not appear to have an associated gravity anomaly (Riddihough, pers. comm.). As with the Crescent Volcanics on Olympic Peninsula, sediments are possibly underthrust below a volcanic slab which is the source of the magnetic high. A preliminary estimate by Riddihough based on a gravity model indicates a possible 2 km of sediment below the volcanics.

3). A magnetic quiet zone 20 to 30 km wide on the shelf between the Prometheus Anomaly and the shelf edge separates the continental and oceanic type anomalies.

Three reversed Deep Seismic Sounding profiles were obtained by R. M. Clowes, Department of Geophysics and Astronomy, University of British Columbia, in the Winona Basin west of the north end of Vancouver Island under contract to the Geological Survey. Two profiles cross the width of the basin, and one follows the length of the basin. The DSS profiles are complemented by single channel continuous seismic profiles using a 300 cubic inch air gun source. Velocity and structural information will be interpreted from the data through the upper and lower crust, and possibly upper mantle from which the thickness of the sediments in Winona Basin and the nature of the basement beneath it may be determined.

References

MacLeod, N. S., Tiffin, D. L., Snavely, P. D., and Currie, R. G.
Major structures of the Strait of Juan de Fuca as interpreted from gravity and magnetic data, and their relation to the geology of the Olympic Peninsula, Washington, and Vancouver Island, British Columbia. (in preparation)

Shouldice, D. H.
1971: Geology of the Western Canadian Continental Shelf; Can. Pet. Geol. Bull., v. 19, p. 405-436.

LARGE-SCALE BLOCK FAULTING DURING DEPOSITION OF THE WINDERMERE SUPERGROUP (HADRYNIAN) IN SOUTHEASTERN BRITISH COLUMBIA

E.M.R. Research Agreement 1135-D-13-4-157/75

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Introduction

Detailed mapping of about 350 km^2 in the western Purcell Mountains between Crawford Bay and Columbia Point, British Columbia has demonstrated that the Late Precambrian (Hadrynian) Windermere Supergroup in this part of the western flank of the Purcell anticlinorium occurs as a simple, steeply dipping to southwesterly overturned homoclinal succession that is now up to 9 km thick. The plane of flattening of deformed clasts in conglomerates within the Windermere Supergroup coincides with a pervasive schistosity that is almost parallel with the bedding; and accordingly, the original thickness may have been substantially greater. The conglomerates which occur at several stratigraphic levels, including the top of the Windermere Supergroup, consist mainly of distinctive rock types that occur in the upper part of the Middle Proterozoic (Helikian) Belt-Purcell Supergroup. This implies that the upper part of the Purcell Supergroup was exposed nearby as the thick succession of Windermere strata was being deposited. Stratigraphic relationships at the unconformity beneath the overlying Lower Cambrian quartzites indicate that the St. Mary Fault, a major northwest-dipping, right-hand, reverse fault, follows the focus of the Late Proterozoic fault which had the northwest side downthrown with a maximum stratigraphic separation of up to 13 km or more.

Windermere Supergroup

The Windermere Supergroup (Young *et al.*, 1973) along the west flank of the Purcell anticlinorium east of the southern part of Kootenay Lake, consists of up to 500 m of conglomeratic strata that have been assigned to the Toby Formation, overlain by up to 8500 m of interbedded pelite, feldspathic quartz wacke and pebble conglomerate, quartzite, limestone and dolomite, and boulder conglomerate that were assigned to the Horsethief Creek Group (Rice, 1941).

The Toby Formation is dominantly a polymict conglomeratic mudstone and conglomerate in which the majority of the boulders, cobbles and pebbles consist of distinctive dolomites and quartzites derived from the underlying Mount Nelson and Dutch Creek formations of the upper part of the Purcell Supergroup. The texture and composition of the conglomeratic strata vary markedly along and across the formation. In many exposures the occurrence of poorly sorted and free floating larger clasts in a massive mudstone matrix is suggestive of deposition as mudflows. In others, the clasts are well sorted and closely packed in distinct beds with a sandy mudstone matrix, and appear to represent fluvial gravels. The Toby Formation unconformably truncates the Mount Nelson Formation and the upper part of the Dutch Creek Formation at a low angle toward the south.

The Horsethief Creek Group is about 8500 m thick near Kootenay Lake. There it contains thick pelite and quartzite units in the lower part, but northward both internal unconformities and erosional truncation at the base of the Cambrian succession have cut out some of the units within it and the total thickness is reduced to about 4500 m. Three distinct polymict conglomeratic units, 20, 80 and 50 m thick, occur in the middle, middle upper and uppermost parts of the Horsethief Creek Group. The sedimentary fabrics of these conglomeratic units and the rock types represented by phenoclasts within them are similar to those in the Toby Formation, and accordingly, they also appear to have been deposited as mudflows and fluvial gravels derived from the upper part of the Purcell Supergroup. However, the fact that boulder conglomerates of this type occur at the top of a sequence of Windermere strata that is up to 9 km or more thick implies not only that the Mount Nelson and Dutch Creek strata from which the boulders were derived must have been exposed nearby, but also that the thick sequence of Windermere strata upon which they were deposited must have occupied a deep structural basin adjacent to the uplifted source area. This leads to the suggestion that the conglomeratic units represent fanglomerates which accumulated adjacent to a fault scarp that separated the uplifted source area from the deep structural basin.

Late Proterozoic Faulting

Regional relationships beneath the sub-Cambrian unconformity (Fig. 28.1) support this interpretation, and moreover, indicate that the St. Mary Fault, a major northwest-dipping, right-hand reverse fault (Rice, 1941; Leech, 1957, 1962), follows the locus of an older fault across which there was up to 13 km or more of downthrow to the northwest during the Late Proterozoic. Northwest of the St. Mary Fault, at the south end of Kootenay Lake, Lower Cambrian quartz sandstones of the Hamill Formation lie unconformably on a section of the Windermere Supergroup that is more than 9 km thick. From there the sub-Cambrian erosion surface gradually cuts deeper into the Windermere Supergroup toward the north, where the Windermere Supergroup is only a few kilometres thick, and toward the northeast, where it is unconformably overlapped in the

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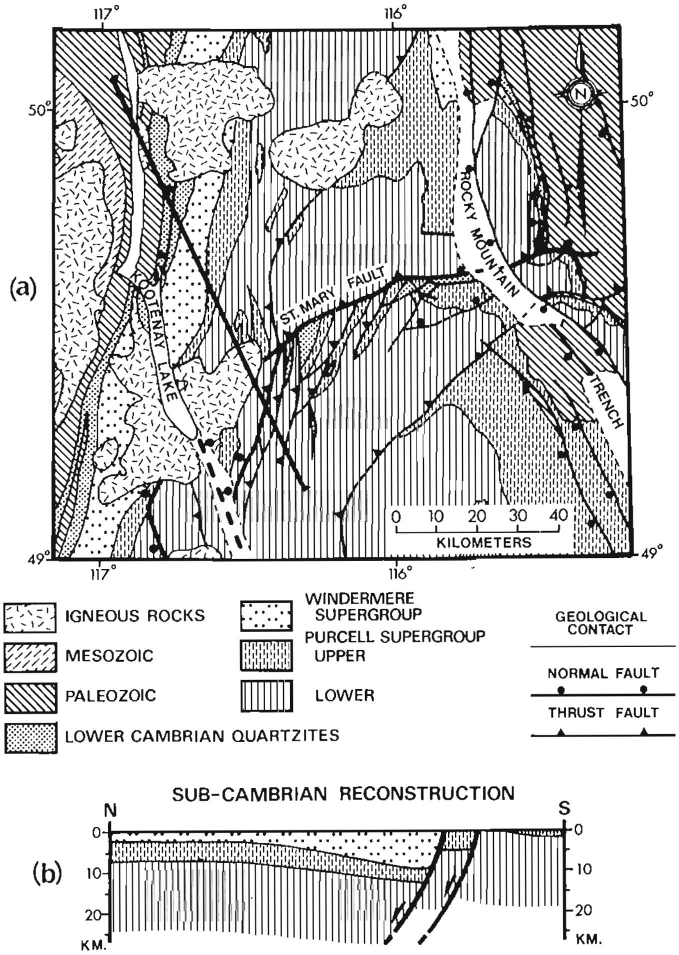

Figure 28. 1

a. Generalized geological map of the Purcell anticlinorium and environs in southern British Columbia.
b. Reconstruction of relationships beneath the sub-Cambrian unconformity along a north-south section across the St. Mary Fault.

western Rocky Mountains by the Lower Cambrian quartzites (Leech, 1954). Southeast of the St. Mary Fault outliers of Lower Cambrian quartz sandstones of the Cranbrook Formation lie unconformably on the Middle Proterozoic strata of the Belt-Purcell Supergroup. The Windermere Supergroup is missing completely and the sub-Cambrian unconformity, when traced from one outlier to another, cuts progressively deeper into the Belt-Purcell succession southwestward from the Rocky Mountain Trench to where the St. Mary Fault is cut by Mesozoic granitic plutons. There, the Cranbrook Formation lies unconformably on the Creston Formation, and some 4 km of Middle Proterozoic strata, which are present beneath the sub-Cambrian unconformity farther northeast, near the Rocky Mountain Trench, have been eroded away. The fact that the deepest pre-Lower Cambrian erosion of Middle Proterozoic rocks on the block southeast of the St. Mary Fault (about 4 km) occurs adjacent to the thickest section of Late Proterozoic rocks that was deposited, and preserved, in the block northwest of the St. Mary Fault (more than 9 km) is noteworthy because it implies that the Late Proterozoic fault displacements involved a combination of uplift and southeasterly tilting of the block on the southeast side and of subsidence and northeasterly tilting of the block on the northwest side, and that the fault displacements were hinged about a point to the northeast in the western Rocky Mountains where the stratigraphic separation beneath the sub-Lower Cambrian unconformity decreases essentially to zero.

The magnitude of the stratigraphic separation associated with the Late Proterozoic faulting indicates that it represents a structure of crustal dimensions. This structure has had a profound influence on the nature and distribution of Late Proterozoic deposits in this part of the Cordillera and is a feature that warrants further investigation in view of its proximity to important mineral deposits in southeastern British Columbia and to various phenomena that has been ascribed to the East Kootenay Orogeny (White, 1958; Leech, 1962).

References

Leech, G. B.
1954: Canal Flats, British Columbia; Geol. Surv. Can., Paper 54-7.

1957: St. Mary Lake, British Columbia; Geol. Surv. Can., Map 15-1957.

1958: Fernie, west half, British Columbia; Geol. Surv. Can., Paper 58-10.

1960: Fernie, west half, British Columbia; Geol. Surv. Can., Map 11-1960.

1962: Metamorphism and granitic intrusions of Precambrian age in southeastern British Columbia; Geol. Surv. Can., Paper 62-13.

Rice, H. M. A.
1941: Nelson map-area, east half, British Columbia; Geol. Surv. Can., Mem. 228.

White, W. H.
1959: Cordilleran tectonics in British Columbia; Am. Assoc. Pet. Geol. Bull., v. 43, no. 1, p. 60-100.

Young, F. G., Campbell, R. B., and Poulton, T. P.
1973: The Windermere Supergroup of the southeastern Canadian Cordillera. Belt Symposium, 1973, University of Idaho, Moscow, Idaho.

29. THE PETROLOGY AND STRUCTURAL FABRIC OF SOME LOWER PALEOZOIC CALCAREOUS PELITES IN THE PORCUPINE CREEK ANTICLINORIUM NEAR GOLDEN, BRITISH COLUMBIA

E. M. R. Research Agreement 1135-D-13-4-157/75

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Introduction

Detailed field mapping and sampling of an area of some 150 km^2 along the Blaeberry River, 30 km northeast of Golden, British Columbia, during the 1974 and 1975 field seasons has provided new information on the stratigraphy and sedimentology, geologic structures, metamorphism, and tectonic evolution of the Lower Paleozoic calcareous pelites in part of the Porcupine Creek anticlinorium and Split Creek anticlinorium of Cook (1966, 1970) and Balkwill (1969, 1972). These Lower Paleozoic rocks comprise two thick successions of slate with subordinate limestone: the upper part of the Chancellor Formation (Upper Cambrian), and the lower part of the McKay Group (Upper Cambrian and Lower Ordovician), which are separated by a thick massive limestone unit, the Upper Cambrian Ottertail Formation.

The principal goal of this study is the elucidation of the nature, evolution, and tectonic significance of the deformation and metamorphism of these rocks. Particular attention is devoted to the effects of primary sedimentary fabric, nature and magnitude of total strain, and specific metamorphic processes relating to the development of the pervasive slaty cleavage and of the various types of veins and related structures which occur in these rocks.

Stratigraphy

The oldest rocks exposed in the area of study belong to the upper part of the Chancellor Formation (Cook, 1966, 1970) and can be assigned to the upper four of the five subunits that were established by Balkwill (1969, 1972). The lowest, subunit 2, consists of green-grey, very slightly calcareous, grey to red-brown weathering slate which contains thin beds of microcrystalline limestone in proportions of up to 10 per cent (*see* Fig. 29.1). The upper part of the subunit is characterized by cycles 3 to 10 m thick in which non-calcareous slate grades upward through nodular slate containing 10 to 50 per cent thin limestone beds to a light grey, argillaceous, microcrystalline limestone, 20 to 40 cm thick. This is overlain abruptly by non-calcareous slates marking the base of the next cycle. Although the base of subunit 2 is not exposed in the area of the present study, preliminary structural interpretations indicate that the unit must be more than 850 m thick.

Subunit 3 consists of about 230 m of regular interbeds, 1 to 3 m thick, of slate and resistant, massive limestone. The base of subunit 3 is chosen to correspond with the bottom of the lowest massive limestone bed that is 50 cm or more thick. The limestone beds consist of: (1) light grey microcrystalline limestone with irregular, dark grey, argillaceous, silty laminae; (2) occasional light to medium grey, thinly laminated calcisiltites; or (3) dark grey intraoobiomicrite containing tabular intraclasts of these previous two rock types in an oomicrite, oosparite, or oobiomicrite matrix. Although the limestone beds are thin, and some of the intraoomicrite beds show cut-and-fill relationship along their lower contact, individual beds apparently extend over very large areas without change in thickness.

Subunit 4 is remarkably similar to subunit 2. It consists of a lower slate part and an upper cyclic slate-limestone part and is about 150 m thick. The base is taken at the top of the highest massive limestone bed of the subunit 3 type that is 50 cm or more thick.

Subunit 5 is about 120 m thick and is transitional to the massive limestone sequence of the Ottertail Formation. It consists of green-grey calcareous slates that are interbedded with increasing amounts of limestone towards the top. The limestone occurs as: (1) massive, dark grey intraoomicrite in beds 20 to 40 cm thick; (2) dark grey to black, microcrystalline limestone in interbeds 1 to 3 cm thick; (3) coarsely cleaved to massive, light grey, laminated, micritic, silty limestone; and (4) finely laminated and cross-laminated calcisiltite and calcite-cemented quartz siltstone. The base of subunit 5 is chosen at the base of the lowest massive carbonate bed that is 50 cm or more thick and is distinct from the types of limestone involved in the cycles of subunit 4.

The Ottertail Formation consists of an estimated 240 to 260 m of dense, resistant limestone. It is divisible into a lower unit, 150 to 190 m thick, of thick bedded to flaggy, black microcrystalline limestone which contains 20 to 40 per cent by volume of thin beds, irregular layers and nodules of buff weathering, argillaceous and silty, black, fine crystalline limestone; and an upper unit, 100 to 130 m thick, of flaggy, mottled, fine crystalline limestone and finely laminated silty limestones which contain stromatolitic algal laminates and lenses of biomicrite and intrabiomicrite. Within the upper unit, slump folds, load casts, mudcracks and cut-and-fill and rip-up structures are common. Cross-lamination and medium scale cross-bedding occur locally. Fragments of trilobites and brachiopods occur in coarse detrital beds. A thin buff weathering argillaceous calcisiltite at the top of

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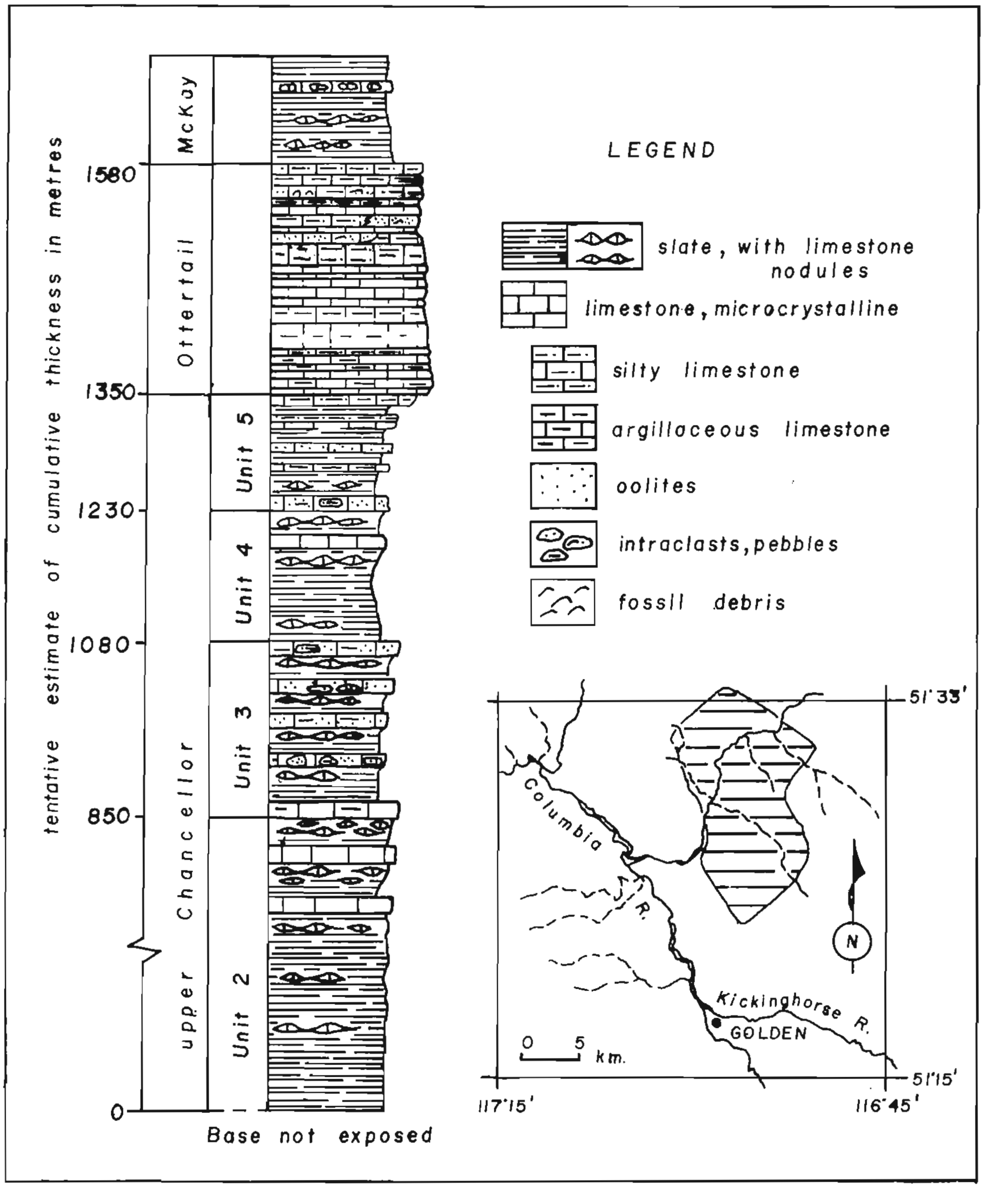

Figure 29.1.

Index map and summary of Cambro-Ordovician Stratigraphy in the Blaeberry River Area, Western and Western Main Ranges, British Columbia. Stratigraphic units numbered to comply with Balkwill (1969, 1972).

the Ottertail Formation is overlain abruptly by the grey-green slates of the McKay Group along a contact which is suggestive of a disconformity.

The lower part of the McKay Group, which is exposed in the overturned southwest limb of the Porcupine Creek anticlinorium and the core of the Split Creek synclinorium, is characterized by green-grey slates with thin 1 cm interbeds of light grey microcrystalline limestone that comprise from 5 to 20 per cent of the succession. Although these interbedded slates and limestones are similar to those in the upper part of the Chancellor Formation, the McKay Group also contains distinctive interbeds of flat pebble and edgewise conglomerate, 10 to 20 cm thick, which comprise abundant pebbles of microcrystalline limestone and occasionally fossil fragments and oolites in a matrix of microcrystalline limestone.

One particularly noteworthy feature of the upper part of the Chancellor Formation and the lower part of the McKay Group is the widespread abrupt interbedding of (1) finely laminated, terrigenous clay rocks, and (2) coarse rudaceous, bioclastic and oolitic limestones which contain sedimentary structures indicative of shallow water deposition. These stratigraphic units which are the westerly "basinal" equivalents of the shallow-water carbonate facies of the Cambro-Ordovician succession in the eastern Main Ranges of the Canadian Rockies (Aitken, 1966; Cook, 1966, 1970), obviously do not represent deep-water deposits. Rather, it is suggested that deposition took place on a shallow, gently sloping marine shelf, over which variations in the rate of supply of terrigenous sediment and small-scale fluctuations in the water level strongly influenced the type of sediment that was deposited at any particular time.

Structural Analysis

Heterogeneities in style of folding and in the nature and magnitude of the strain, which are characteristic of deformation in these rocks, can be attributed to effects of high contrasts in ductility between beds of pelitic rock and limestone and between stratigraphic units comprising different proportions of these two lithologies. Dominantly pelitic units form tight, harmonic, passive flow folds characterized by a microscopically pervasive slaty cleavage. Units consisting mainly of limestone, in particular limestones of the Ottertail Formation, form tight to open flexural-slip folds characterized by bedding-glide features, mesoscopic-scale faulting, jointing and local development of widely spaced "fracture" cleavage planes. In the lithologically mixed units (subunits 3 and 5 of the Chancellor Formation), folding is of a flexural-flow style and the folds are generally of the tight, flattened concentric type.

Faults that occur in conjunction with the folding exhibit complex relationships the details of which are partly obscured by the paucity of good stratigraphic markers. Contraction faults (Norris, 1956) which occur in both overturned limbs of the fan-shaped Porcupine Creek anticlinorium have been externally rotated, along with the beds which they cut (Balkwill, 1972). Many of them have been overturned and are now marked by dip separations characteristic of normal faults. On each side of the fan structure one major fault separates the upright, relatively open structures of the core from the overturned, tight structures of the margins. Offsets due to contraction faulting are most conspicuous in the limestone units. In the pelitic units faulting appears to be represented mainly by distributed shear parallel with the cleavage. Locally those displacements are concentrated within narrow shear zones that contain 10 to 100 cm thick massive, quartz-calcite-chlorite-siderite veins and are cut by massive, sigmoidal, *en echelon* tension gashes.

The dominant element in the tectonic fabric of the area is a pervasive slaty cleavage. The cleavage is parallel with the axial surfaces of the folds and outlines a broad cleavage fan which is symmetrical about the hinge zone of the anticlinorium. On both of the overturned limbs of the anticlinorium a subvertically oriented younger cleavage, of strain-slip or crenulation type, occurs locally. This younger cleavage is probably the same as that observed by Balkwill (1972) in stratigraphically equivalent rocks along the tectonic strike to the southeast at Mount Hurd.

Measurements of deformed calcite veins, thin limestone beds and other "strain markers" made in conjunction with detailed studies of mesoscopic fabrics in the slates of the upper Chancellor at 10 localities throughout the area have been analyzed, using a modification of the method described by Talbot (1971), to obtain estimates of the nature and magnitude of the minimum total strain in the rocks. The strain appears to be homogeneous within the domain of individual sample localities and is essentially of the pure flattening type (Flinn, 1959) with axial ratios of $a = \frac{x}{y} \approx 1$ and $b = \frac{y}{z} = 3.8$ to 4.2 $(x > y > z)$, and with the plane of flattening aligned with the cleavage. The estimated decrease in volume during deformation is about 20 per cent. This implies that there has been approximately 45 – 55 per cent shortening perpendicular to the cleavage, and that the pelitic beds, which comprise 85 per cent of the upper Chancellor have been tectonically thickened by at least 175 per cent.

The slaty cleavage in the pelitic beds of the upper Chancellor Formation and the McKay Group is characterized by a microscopically pervasive subparallel alignment of fine chlorite and muscovite flakes in a matirx of fine crystalline quartz and calcite. Several samples of slate from subunit 2 of the upper Chancellor which have been examined by whole rock X-ray diffraction analysis on a 4° – 70° scan (*see* Fig. 29.2) are similar in composition and consist of the mineral

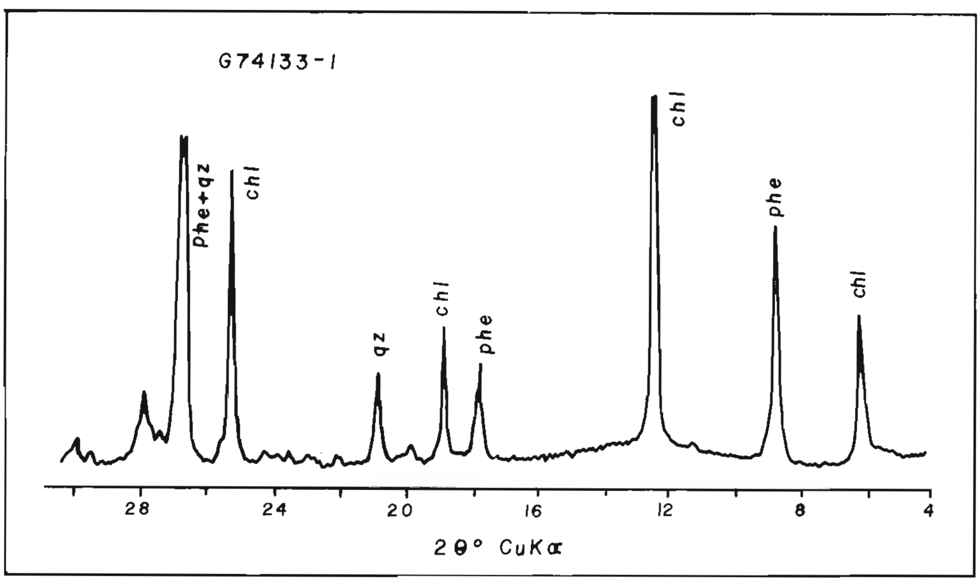

Figure 29.2.

Diffraction scan [4-30°] of sample G74133-1 of subunit 2 of the upper Chancellor Formation. Abbreviations: chl – chlorite; phe – phengite, qz – quartz.

assemblage chlorite-illite/muscovite (phengite) -quartz-(calcite). Estimates of the illite crystallinity [1.8 – 2.1] (Kubler, 1967) and the ratio of (001) to (002) reflection intensities for mica [0.4 to 0.5] (Esquevin, 1969) place the slates within the epimetamorphic zone of Frey (1970).

In thin section, calcite commonly appears to be a late replacement mineral after quartz. Silicate minerals and other insoluble materials which are disseminated as fine flecks throughout the slate, appear in greater concentrations as thin films along regularly spaced cleavage planes. Higher concentrations of chlorite, muscovite and iron oxides along these planes suggest that they are loci of: (1) selective removal of quartz and calcite; (2) recrystallization of sheet silicates, and; (3) concentrated movement of intergranular fluids during metamorphism and deformation.

In limestone-rich units, widely spaced fracture cleavage planes form the only generally visible cleavage. Here stylolites, dissolved grain contacts and films of insoluble residue along cleavage planes indicate that the fracture cleavage is due to localized pressure solution and removal of calcite.

Veining, which is abundant in many units, provides evidence of the high mobility of quartz and calcite during the deformation. Calcite veins are ubiquitous, but more abundant locally in limestone beds. Veins containing quartz appear to be restricted to slate horizons. This suggests local sources for the vein minerals and supports the concept that there was widespread pressure solution and recrystallization during the deformation and the development of the cleavage.

References

Aitken, J.D.
1966: Middle Cambrian to Middle Ordovician cyclic sedimentation, southern Rocky Mountains of Alberta; Can. Pet. Geol. Bull., v. 14, p. 405-441.

Balkwill, H.R.
1969: Structural Analysis of the Western Ranges, Rocky Mountains, near Golden, British Columbia; unpub. Ph.D. thesis, Univ. Texas at Austin, Austin, Texas.

Balkwill, H.R. (cont.)
1972: Structural geology, lower Kicking Horse River region, Rocky Mountains, British Columbia; Can. Pet. Geol. Bull., v. 20, p. 608-633.

Cook, D.G.
1966: Structural style influenced by a Cambrian regional facies change in the mount Stephen-Mount Dennis area, Alberta-British Columbia; unpub. Ph.D. thesis, Queen's Univ., Kingston, Ontario.

1970: A Cambrian facies change and its effect on structure, Mount Stephen-Mount Dennis area, Alberta-British Columbia; *in* Wheeler, J.O., *et al.*, Structure of the southern Canadian Cordillera, Geol. Assoc. Can., Spec. Paper no. 6, p. 27-40.

Esquevin, J.
1969: Influence de la composition chimique des illites sur leur cristallinité; Bull. Centre Rech. Pau. – S.N.P.A., v. 3, p. 147-153.

Flinn, D.
1962: On folding during 3 dimensional progressive deformation; Geol. Soc. London, Quart. J., v. 118, p. 385-428.

Frey, M.
1970: The step from diagenesis to metamorphism in pelitic rocks during Alpine orogenesis; Sedimentology, v. 15, p. 261-279.

Kubler, B.
1967: La cristallinité de l'illite et les zones tout a fait supérieures du métamorphisme; Etages Tectoniques, Coll. Neuchâtel, p. 105-122.

Norris, D.K.
1956: Structural conditions in coal mines of the Canmore area, Alberta; Alberta Soc. Pet. Geol., Guidebook, 6th Ann. Field Conf., p. 59.

Talbot, C.J.
1970: The minimum strain ellipsoid using deformed quartz veins; Tectonophysics, v. 9, p. 47-56.

SHALLOW COREHOLE DRILLING ON THE BAFFIN ISLAND SHELF

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A program of shallow drilling to obtain samples of the bedrock underlying parts of the southeastern Baffin Island Shelf was undertaken from CSS HUDSON during the period September 15-17, 1975 (cruise 75-009, phase V). The samples were collected by means of the Bedford Institute of Oceanography's underwater electric rock core drill. This instrument is designed to penetrate to a maximum depth of 6 m below seafloor. Operation of the drill was carried out by Huntec (70) Limited with the assistance of the Department of Energy, Mines and Resources and Department of the Environment personnel. The 1975 program served as a feasibility test for the more extensive bedrock sampling programs planned for this frontier area in future years.

Fifteen drilling attempts were made at nine different localities and six cores totalling 544 cm in length were recovered from four of these localities (Fig. 30.1 and Table 30.1). Cores from three localities consisted of limestone whereas the fourth consisted of gneiss.

The shelf southeast of Baffin Island (Hudson Strait to Cape Dyer) has been investigated mainly by Grant (1975) who inferred six distinct bedrock units on the basis of continuous seismic reflection and magnetic profiling. His investigations also revealed that the structure of the southern Baffin Shelf is considerably more

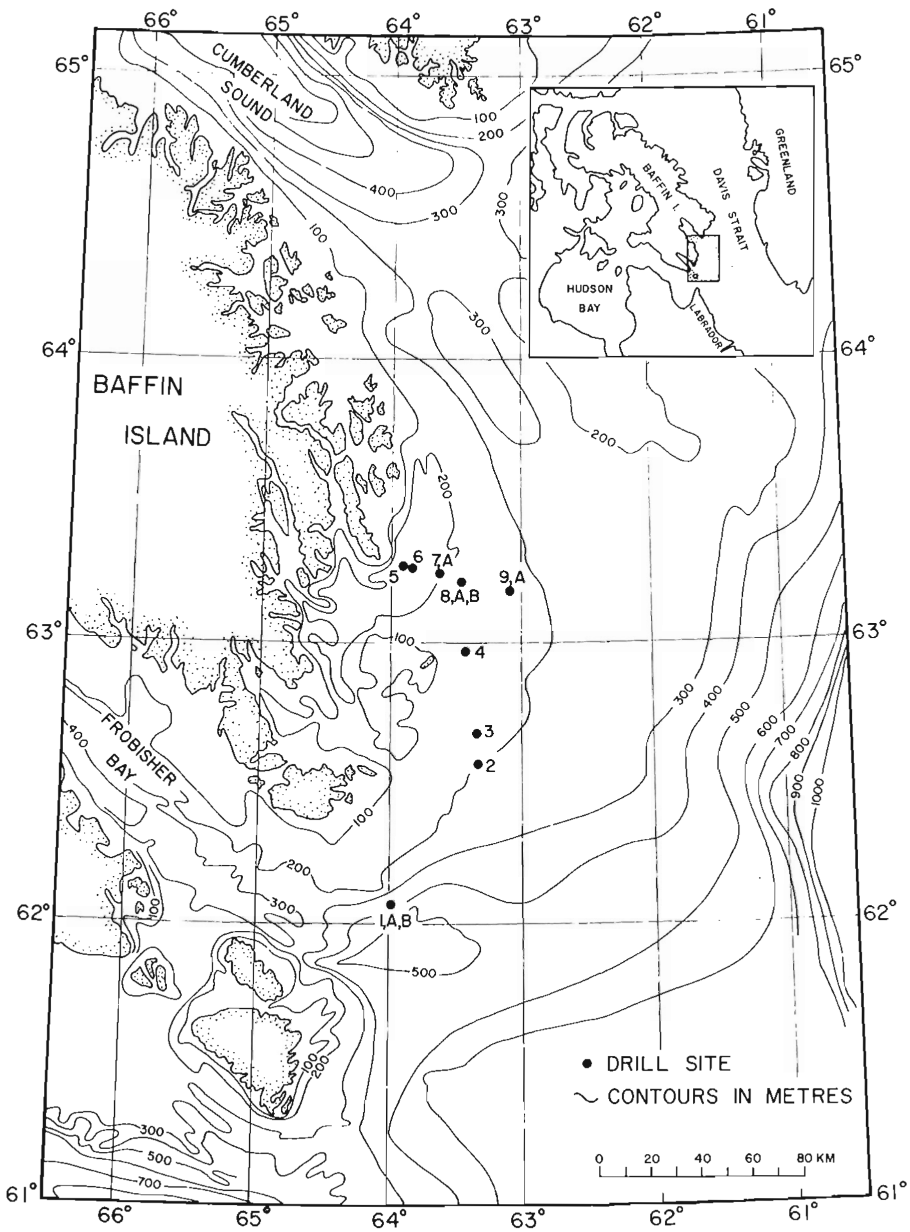

Figure 30.1

Map of program area showing drill site locations and regional bathymetry.

Table 30.1

Geological Sample Stations

Station No.	Lat.	Long.	Water Depth	Drill Extension	Nature	Results
1	62°04.2'N	63°58.0'W	289 M	39 cm	Drill	No sample
1A	62°04.0'N	63°59.4'W	190 M	80 cm	Drill, grab, camera	Gravel
1B	62°03.9'N	63°59.7'W	271 M	500 cm	Drill	No sample
2	62°34.2'N	63°19.5'W	201 M	540 cm	Drill, grab, camera	Gravel
3	62°40.9'N	63°20.2'W	183 M	260 cm	Drill, grab	Gravel
4	62°58.2'N	63°26.1'W	157 M	457 cm	Drill, grab	137 cm limestone core
5	63°16.2'N	63°54.6'W	358 M	181 cm	Drill	54 cm limestone core
6	63°15.7'N	63°50.6'W	430 M	525 cm	Drill, grab	Mud, sand, gravel
7	63°13.8'N	63°37.5'W	216 M	344 cm	Drill, grab	Gravel, mud
7A	63°13.9'N	63°37.5'W	216 M	320 cm	Drill	No sample
8	63°12.8'N	63°27.2'W	165 M	414 cm	Drill	Mainly gravel
8A	63°12.9'N	63°27.5'W	165 M	331 cm	Drill	45 cm limestone core
8B	63°13.2'N	63°27.6'W	165 M	222 cm	Drill	83 cm limestone core
9	63°11.1'N	63°06.2'W	176 M	179 cm	Drill, grab	66 cm gneiss core
9A	63°10.7'N	63°04.6'W	176 M	285 cm	Drill, camera	158 cm gneiss core

complex than that of the Labrador Shelf to the south. The importance of bedrock samples in this area is indicated by the fact that except for a short sandstone core collected by Srivastava (1974) at a site some 130 km northeast of locality 9 and samples of rocks of possible Silurian age dredged by Grant (1975) from Hudson Strait, no bedrock sample data has been available for control of the remote sensing information north of the Eastcan *et al.* Bjarni H-81 well on the Labrador Shelf, 520 km south of Hudson Strait.

Prior to the cruise several potential sites were selected on the basis of existing seismic reflection data. Specific drill sites were determined at sea on the basis of surveys conducted with shallow seismic (625 cm^3 air gun) and Huntec high resolution deep tow systems. The shallow seismic and high resolution systems, despite individual limitations, form a combination that is essential for the location of drill sites. Because of the very limited time available for the program, drilling efforts were concentrated at sites where seismic data indicated the rocks were likely to be well consolidated and therefore of a type that could be cored successfully by the present model of the drill, and where protracted site surveys were not likely to be required. As anticipated, site selection and locating the drill on target presented problems at some sites where overburden proved too thick. Also, mechanical and other problems led to the early termination of some holes.

The bedrock cores obtained during the program represent an important increase in geological data available in the region. It is hoped that these cores will yield definitive biostratigraphical data in addition to the lithostratigraphical and other data that will be obtained from laboratory studies. Structural complexities make regional seismic correlation difficult, but the limestone sequences cored during this program appear to occur lower in the stratigraphic succession than the sandstone section cored by Srivastava in 1974.

References

Grant, A.C.
1975: Geophysical results from the continental margin off southern Baffin Island; *in* Canada's Continental Margins and Offshore Petroleum Exploration, eds. C.J. Yorath, E.R. Parker, and D.J. Glass; Can. Soc. Pet. Geol., Mem. 4, p. 411-431.

Srivastava, S.P.
1974: Cruise Report Hudson 74-026; Phase IV; September 26-November 1; Bedford Institute of Oceanography, Dartmouth, N.S., unpubl. ms.

31. TERRAIN INVENTORY AND QUATERNARY GEOLOGY, SOMERSET, PRINCE OF WALES, AND ADJACENT ISLANDS

Project 750071

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Terrain studies were conducted on Somerset, Prince of Wales, and adjacent smaller islands in the eastern Canadian Arctic during summer 1975 to provide information on the distribution and characteristics of the surface materials and the processes affecting them. Geomorphic and geologic information was provided by the authors; soils and vegetation data will be the subject of a future report by S.C. Zoltai[1] and V. Woo[2]. Where possible glacial and nonglacial materials and landforms were examined in order to describe the Quaternary history of the area.

The field component of the project primarily involved traversing by helicopter from base camps situated on northwestern Somerset Island and central Prince of Wales Island (Fig. 31.1). This facet provided information on texture, lithology, origin, and moisture content of materials as well as their morphology, surface drainage, and depth to the frost table. Approximately 250 samples of material above the frost table were collected. Of these, roughly 75 per cent will be subjected to laboratory tests to determine their engineering and geochemical properties. The remainder consists of organic material for identification and radiocarbon dating.

Detailed investigations of the properties of frozen ground were carried out near Aston Bay, Cunningham Inlet, Cresswell Bay, and Stanwell-Fletcher Lake on Somerset Island and Back Bay on Prince of Wales Island. These studies utilized portable coring equipment and were oriented primarily towards supplying information on ground ice characteristics. Approximately 400 borehole samples and 170 pit samples were collected for determination of moisture content, Atterberg limits, and grain size distribution (Veillette, 1976). Some of the studies were complimented by a shallow seismic and resistivity program directed by J.A. Hunter.

Terrain Regions

The study area is divided into 31 terrain regions (Fig. 31.1), distinguished by differences in thickness, texture, and origin of materials and by relief and slope. Elements of one region may be found as minor constituents of another, and some of the boundaries are generalized and approximate. They have been drawn from field observations and interpretation of photomosaics, ERTS imagery, and topographic maps. Discussions of bedrock geology are based on the map of Blackadar *et al.* (1967).

[1]Department of the Environment.

[2]Department of Agriculture.

The relief and active layer properties of each region are summarized in Table 31.1. Active layer thickness and drainage were investigated by digging holes down to the frost table. The active layer thickness in Table 31.1 represent the range of typical depth to frost values encountered. Many measurements were made up to two months before the end of the melt season. Those made near the end of the melt season, however, rarely exceeded one metre. Patterned ground features were noted for the area surrounding each hole.

Many more data are available for some regions than for others; regions of Somerset Island were investigated in more detail than those of Prince of Wales Island. From these notes and observations made during traverses by helicopter it is concluded that nearly 100 per cent of the ground has been modified by periglacial processes. Large areas of slopes are characterized by parallel stone and vegetation stripes and more level areas by high-centre circles and stone nets. Desiccation cracks are common and ice-wedge polygons dominate areas of water-laid sediments and smaller areas of till.

Somerset Island

Region 1

Region 1 is the largest terrain unit on Somerset Island, covering most of the Paleozoic upland. Its boundaries are coincident with bedrock contacts and the entire area is underlain by limestone and dolomite assigned to the Read Bay Formation. Except along the western margin where strata are deformed into anticlines and synclines, the carbonates are flat lying. Their colour is a uniform grey to yellowish grey, in places interbedded with red. Bedrock outcrops are uncommon except in canyons.

The upland surface is mantled by a 1- to 3-m-thick accumulation of rubble, the texture of which varies from nearly pure sand and silt to large angular fragments with little or no fine grained matrix. Most of the accumulation, however, varies from moderately stony to very stony silt. Variations in stoniness appear to reflect variations in veneer thickness; nearly pure stone accumulations occur where bedrock lies close to the surface and nearly stone-free materials occur where the veneer is thickest. The latter occupy depressions into which fines have soliflucted from adjacent slopes.

The genesis of the rubble is problematical. Craig (1964) referred to it as a "silty rubbly glacial till" and pointed to the fact that Precambrian erratics are ubiquitously distributed upon its surface. Field work conducted in 1975 confirms the presence of these erratics. Thus, it is reasonable to assume that at least some of the rubble is till. Other observations, however, demonstrate that much of the material has been produced by weathering of bedrock as Blackadar (1967, p. 5) concluded. These observations are as follows: (1) Lithological changes across bedrock strata in many areas are duplicated in the overlying rubble. This shows that there has been no intermixing of rubble from different strata except for minor mixing due to solifluction. (2) Study of tors has permitted determination of the characteristics of material produced by subaerial weathering of carbonates. Layers of silt and sand between less weathered strata are exposed on tor sides, and the features are surrounded and partially overlain by silty rubble. This demonstrates that material identical to that which mantles the Paleozoic upland has been produced in this environment by weathering. Furthermore, the size (heights average 2 to 5 m) and widespread distribution of carbonate tors demonstrate that large volumes of bedrock have been altered to such material.

Stream networks are well integrated; ponds and boggy areas are rare. The drainage pattern is dendritic, except in the southeast where streams follow a trellis pattern. In the central portion, headwaters of large rivers flow in broad valleys. Middle and lower reaches of rivers are incised 100 to 150 m into bedrock, producing spectacular canyons. Many canyons portray excellent examples of entrenched meanders which testify to the existence of mature drainage systems prior to rejuvenation and canyon cutting. Little, if any, lateral widening of canyon floors has been achieved; rivers appear to be actively degrading their courses. Canyons and sections of canyons are diversely oriented but no section shows any signs of glacial modification. It is suggested, therefore, that canyon cutting postdates

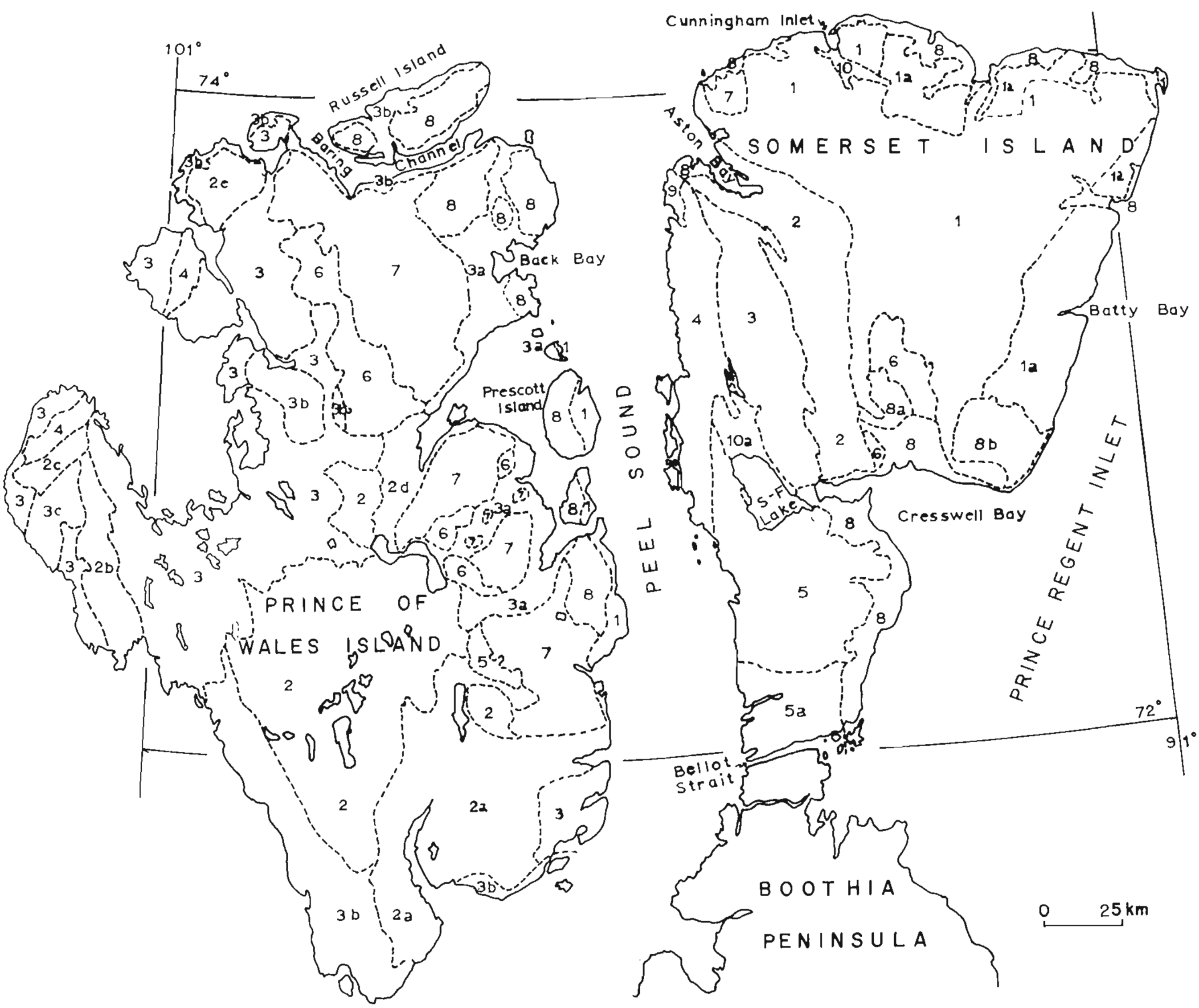

Figure 31. 1. Terrain Regions of Somerset and Prince of Wales Islands. Regions 1 to 8 on Somerset Island bear no relation to Regions 1 to 8 on Prince of Wales Island. (S-F=Stanwell-Fletcher)

Table 31.1

Relief and active layer properties of terrain regions

Terrain Region	Total Relief (m)	Local Relief (m)	Depth to Frost (cm)	Active Layer Drainage
SOMERSET ISLAND				
1	400	10-60	35-75	good
1a	150	10-30	43-54	good
2	300	10-20	42-68	good
3	500	50-150	35-62	moderate-good
4	400	100-150	58-89	moderate-good
5	425	30-150	60-70	poor
5a	500	150-200	75	moderate
6	180	15-30	-	good
7	300	40-60	-	-
8	150	2-10	40-73	good
8a	150	10-30	-	good
8b	60	10-20	68-82	moderate
9	130	10-30	65	good
10	120	30-50	40-60	good
10a	300	10-50	60-70	good
11	240	30-60	-	-
PRINCE OF WALES ISLAND				
1	350	70-175	-	poor
2	50	30-40	33-75	poor-good
2a	100	10-40	81-87	good
2b	35	10-20	30-70	poor-good
2c	30	5-20	-	-
2d	100	10-20	-	-
2e	105	10-40	-	-
3	200	20-80	43-80	good
3a	30	1-5	55-80	good
3b	115	2-10	57	moderate-good
3c	120	10-30	-	-
4	120	20-60	-	-
5	40	10-20	-	-
6	125	10-20	47-72	good
7	200	10-30	50-81	good
8	300	10-15	50-60	good

deposition of erratics on the surrounding plateaus. Alternately, the canyons predate glaciation and for some reason escaped modification.

Rivers following dendritic patterns flow without regard for bedrock structure. Adjacent to the canyons minor channels are eroded parallel to the strike of the bedrock and along joints. This structurally controlled element obviously postdates evolution of the main drainage pattern. Craig (1964) interpreted them as glacial meltwater channels. No evidence was found during the field study to support a glaciofluvial origin. If they are meltwater channels, however, their distribution indicates that the island supported an independent ice cap during the period of their formation. Many channels contain snowbanks which persist long into summer or throughout summer, and it is possible that nivation has played a large role in their evolution.

Region 1a

The surface here is underlain by the same bedrock type as in Region 1, and the nature of the rubble veneer is essentially identical to that discussed above. The character of the region differs in that it contains a large number of ponds and a hummocky surface. Unfortunately, field investigations were not sufficient to reveal definitely the reason for the presence of these features. On aerial photographs and from inspection during helicopter traverses, it appears that the ponds occupy bedrock basins and the hummocks have a kame-like morphology. Except for two mentioned below, the few hummocks examined proved to have bedrock cores which are almost entirely mantled by silty rubble. Apart from the lack of exposed bedrock sides they are identical to the tors discussed above and are likely to have developed by the same process.

Along the eastern coast canyons broaden into wide valleys near their mouths. These are partially filled with remnants of raised deltas, scree, and modern alluvium. Such a valley situated south of Batty Bay and part of an adjacent canyon are nearly filled with silts containing a few marine shell fragments. The silts and overlying sands and gravels extend up to an elevation of 227 m where they are associated with two kames. It is suggested that the valley mouths here and northward in Region 1 were excavated by different tongues of ice spreading landward from a trunk glacier in Prince Regent Inlet. The fact that the silts fill a portion of the canyon demonstrates that the canyon was cut prior to silt deposition. Association of kames with the deposit indicates that it may be glaciomarine or glaciolacustrine. The latter would require that the shell fragments are erratics.

Region 2

The boundaries of Region 2 coincide with bedrock contacts. Most of the surface is underlain by Lower Paleozoic dolomite and sandstone. The remaining area, between the southwestern shore of Aston Bay and the contact with Precambrian gneisses is underlain by Proterozoic dolomite, sandstone, siltstone, and cherty dolomites intruded in a few places by gabbro dykes.

The region differs from Region 1 mainly in that the sedimentary beds here are more tightly folded. Differential erosion of the strata has caused dominance of much of the topography by low parallel ridges and troughs. Craig (1964) has mapped some of these ridges as drumlins. This invokes a northward ice flow.

Bedrock outcrops are rare except along steep valley sides and canyons. The surface elsewhere has developed on a silty rubble veneer which appears slightly less stony than that of Region 1. Although the region borders on Precambrian gneisses along the west, the frequency of occurrence of gneissic erratics immediately east of the contact is no higher than in places far removed from it. Indeed, the contact is well expressed in the unconsolidated veneer and provides a useful navigational mark. The contact between felsenmeer developed upon the dark-coloured dykes and the carbonate rubble is equally well defined. These facts suggest that most of the surficial material has developed through weathering of bedrock rather than through deposition by glaciers.

Both dendritic and trellis drainage patterns occur; signs of derangement are absent. The only rock glaciers observed on the island are located immediately southwest of Aston Bay. These are composed of scree which has flowed towards the valley axes. A valley in the same locality contains an esker near its mouth. The valley mouth is broad but grades into a canyon a short distance upstream.

Region 3

While the eastern boundary of Region 3 coincides with the contact of gneisses with carbonates, the western boundary is drawn on the basis of topographic considerations. Bedrock comprises felsic and mafic gneisses except for the northern fringe which is underlain by Proterozoic quartzite. The topography is characterized by low rounded hills separated by broad valleys. The drainage network is well integrated, and the few small lakes which occur occupy basins in bedrock.

Most of the bedrock is concealed by a veneer of surficial materials but bedrock structure is detectable through this, particularly along the eastern margin. The veneer consists mainly of both bouldery till and felsenmeer; in places it is difficult to differentiate the two materials. Accumulations with abundant fines commonly were mapped as till. Dozens of large (2 to 4 m high) and well formed tors were observed throughout the region but are particularly common in the northeast portion. Studies of these (to be reported in a future paper) revealed that large quantities of sand-size material have been produced by weathering of gneiss. This, along with the discovery of significant amounts of sand in felsenmeer accumulations, leads to the difficulty in distinguishing tills from weathering products.

Although very scarce, carbonate and sandstone erratics were found at many stops. An esker occurs in the west-central portion of the region, and a line of kames runs parallel to the eastern shore of Stanwell-Fletcher Lake in the southern portion. Raised beaches attaining an altitude of 230 m occur both above and below the kames. On the southern margin of the region a fossiliferous marine delta extends to an elevation of at least 205 m.

Considerable areas of ground above the 400 m contour are nearly lichen-free. This probably represents lichen kill by a recent and persistent snow cover as suggested by Ives (1962) for similar areas on Baffin Island.

Region 4

Region 4 is underlain by the same types of bedrock as in Region 3. The primary difference between the two is one of topography. Here the surface is much more highly dissected with valleys and drainage following two dominant trends, namely: north-south, parallel to bedrock boundaries and foliation, and northwest-southeast, parallel to either a fracture or joint system. Numerous lakes and ponds occupy bedrock basins.

Surficial materials are thin and discontinuous. By far the most prevalent material is felsenmeer, interspersed with thin patches of bouldery till. The area west of Stanwell-Fletcher Lake has smooth bedrock surfaces exposed on valley bottoms and sides and felsenmeer-covered summits. Marine sediments ranging from silt to gravel fill many valley mouths. Three small eskers sit in valley bottoms in the west-central portion of the area and a series of kames and small kame moraines lie immediately west of Stanwell-Fletcher Lake. The latter are considered contemporaries of the kames above the eastern shore of the lake as they are at approximately the same elevation. Collectively, they outline the margin of an ice lobe which occupied the lake basin.

Region 5

Region 5 is an upland covering a large area south of Stanwell-Fletcher Lake. The boundaries are drawn on the basis of topography and drift origin and thickness. The topography, similar to that of Region 3, has developed mainly upon gneisses but the eastern portion is underlain by gently dipping sandstone, dolomite, and limestone. The drainage pattern is structurally controlled with streams and lakes oriented north-south and east-west. Rivers flowing across carbonate strata have eroded canyons to a depth of about 30 m.

Bedrock outcrops on about 25 per cent of the surface, mostly along valley sides and on steeper slopes. Mantling the bedrock over most of the remaining area is a moderately stony silty till, the thickness of which ranges from less than 1 to about 5 m. Unlike the northern portion of the island, the contact between gneisses and carbonates is largely obscured by the till sheet, and many gneissic erratics occur in till which overlies carbonates.

At elevations in excess of 200 m silt and stony silt deposits containing marine shells and shell fragments are found. At low elevations along the west coast, pockets of marine silts and sands occur as beach and deltaic deposits in valley mouths. Glaciofluvial sands have been mapped along the floors and sides of many valleys as well as groups of kames in the north and northeast.

The extreme northeastern area contains a number of well formed carbonate tors but none were found on gneissic bedrock.

Region 5a

The southern tip of the island has been designated Region 5a. Drift cover is thin or absent, and the topography is much more rugged than in Region 5. The bedrock here differs from that of Region 5 only in that small amounts of granite are present.

Over 80 per cent of the surface is bare bedrock with patches of felsenmeer and grus. Colluvium has accumulated in valley bottoms and depressions, while in large valleys minor amounts of glaciofluvial gravel and sand are found.

Region 6

The boundaries of Region 6 are primarily defined by the limit of the Peel Sound Formation in the area north of Cresswell Bay. The formation contains conglomerate, sandstone, and siltstone in mainly flat-lying beds. The bedrock is overlain by a silty rubble veneer ranging from slightly to very stony. The primary difference between this region and the main extent of the Paleozoic upland lies in the degree of dissection. Bedrock here is intensely dissected by minor gullies, most of which appear to follow joint systems.

Region 7

Region 7 includes all of the Peel Sound Formation, which outcrops north of Aston Bay, except for a coastal strip (Region 8) containing Quaternary marine sediments. As with Region 6 the area is outstanding for its advanced degree of dissection. A massif of small mountains, in which several valleys head in cirque-like features, is the result. On level areas a silty rubble veneers the bedrock from which it has been derived.

Region 8

This region covers a fairly wide coastal strip north and south of Cresswell Bay and along the northern coast. It is underlain by Lower Paleozoic carbonate rocks. Smaller areas of the same region occur along the northeast coast. The dominant materials are marine silts and sands along with many gravel beach ridges. The entire region lies below the 150 m contour. There is no apparent systematic change in elevation of the highest strandline features around the coast. Small areas of alluvial gravels and marine deltaic silts were mapped. The silts are commonly rich in ground ice, and display several flow-slide scars. A detailed investigation of the ground ice characteristics of the materials in the region around Cresswell Bay was made (Veillette, 1976).

Region 8a

This region has many features similar to Region 6 in that it is underlain by rocks of the Peel Sound Formation and is similarly dissected. It has, however, a closer affinity to Region 8 in that extensive marine silts and stony silts blanket the bedrock. These deposits are thin and lack constructional features.

Region 8b

The surficial material here is underlain by the same type of bedrock as is Region 1. The boundaries outline a deposit of fossiliferous marine silt which is slightly to moderately stony. Its thickness varies considerably but over large areas appears to be in excess of 3 m. Intensive development of high-centred ice-wedge polygons imparts a gently undulating form to the surface. Near the centre of the region a ridge, 7 km long and 30 m high, composed of gravel with a high concentration of Precambrian stones, is interpreted as a kame moraine. Several other smaller features nearby are thought to be composed of the same material.

Region 9

This region, situated near the northwest corner of the island, is underlain for the most part by granite and lies below the marine limit. The granite consists of approximately 65 per cent red feldspar, 30 per cent quartz, and 5 per cent biotite. It is separated from the main marine unit (Region 8) because the beaches here consist almost entirely of grus with a grain size approximating the size of crystals (2 to 5 mm) in the parent rock. Sand-size material derived by breakdown of individual crystals provides the matrix of this fine pebbly gravel whose depth varies from less than 1 m to more than 2 m. Grusification of the bedrock must have preceded marine inundation. The rest of the region, which makes up about 50 per cent of the total, consists of bedrock outcrops, felsenmeer, and till.

Region 10

Region 10 covers a small area of northern Somerset Island. It is basically a "bedrock" unit and outlines roughly the area of occurrence of Eureka Sound sandstone. The sandstone contains lignite, is compact, but is unlithified. It is overlain in many places by fossiliferous Quaternary marine gravel, sand, and silt. Due to the organic content of the formation, a lush vegetation has developed on it.

Region 10a

North of Stanwell-Fletcher Lake a second, larger area is underlain by rocks of the Eureka Sound Formation. The sandstone here is more highly lithified than in Region 10. Quaternary marine sand and gravel are present but over most of the area the sandstone outcrops. The ground ice characteristics of the materials in this area and also in Region 10 were studied in detail (Veillette, 1976).

PRINCE OF WALES ISLAND

Region 1

Lying along the extreme eastern portion of Prince of Wales and adjacent small islands, Region 1 is underlain by gneisses and minor amounts of sediments. The boundary is defined by the contact between these rocks and the overlying Peel Sound Formation. The drainage pattern is structurally controlled; rivers are oriented north-south, parallel to fold axes, and east-west, parallel to regional jointing. Numerous small lakes occupy bedrock basins.

More than 60 per cent of the surface is composed of intensely jointed and frost-cracked bedrock. The remainder has a veneer of stony rubble composed of pebbles and cobbles set in a silt and sand matrix. Depressions contain concentrations of silt and sand, washed and soliflucted from surrounding slopes. In large valleys at low elevations marine silt and sand form deposits up to 3 m thick. Except near sea level, bedrock outcrops through the marine deposits. On more gentle slopes near sea level, beaches composed of angular rubble have formed.

The glaciation of Prescott Island is documented by occurrences of roches moutonnées and striae indicating northward flowing ice cross-cut by striae with a northeast orientation. Marine shells were found up to 350 m above sea level but it is not known whether they were deposited during a high sea level stand or were brought to their present positions by glacier ice.

Region 2

The largest terrain unit on the island is a till sheet containing large well defined drumlins formed by north-northwestward flowing ice. The till is very stony with a silt and sand matrix. It is derived almost entirely from carbonate bedrock. The entire region has experienced marine inundation. For this reason, many drumlins have beach ridges perched on their tops and fine grained marine sediments in intervening depressions.

Regions 2a-2e

Regions 2a, 2b, 2c, 2d, and 2e all contain drumlinized or fluted till and lie below the marine limit. Except for the smaller size of individual drumlins and the diverse ice-flow directions indicated by them, the properties of these regions are similar to those of Region 2.

Region 3

Terrain Region 3 covers a large area of western Prince of Wales Island and a smaller area in the southeast. It is a complex unit consisting of a low-lying hummocky till sheet and a number of eskers. The entire area lies below the marine limit so gravel beaches and pockets of marine fines occur. Material texture varies with origin and extent of modification by marine activity. Thus, low relief till surfaces, when covered by a thin blanket of marine sediments and mixed with the latter by cryoturbation, exhibit a stony silt texture, whereas till ridges which have been extensively washed are relatively stony.

The drainage pattern is highly deranged. Lakes and ponds are numerous.

Region 3a

The area along the eastern coast, generally below the 30 m contour and lying seaward of a pronounced escarpment, has been designated Region 3a. Bedrock is not exposed. The till is composed of slightly stony silty sand with a red colour similar to that of rocks of the Peel Sound Formation. Overlying the till throughout most of the region are marine sands and silts which in places are slightly stony. These vary in thickness from less than 1 m to more than 10 m and thicken coastward.

Region 3b

This region, which covers large areas near the coast, differs from Region 3 only by a greater number of raised beaches. In many places beaches dominate entirely.

Region 3c

Region 3c is an area on the western part of the island which has a continuous cover of marine sand. No field data are available.

Region 4

Region 4 occurs as two small units in the northwest. The region was not investigated in the field. Craig (1964), however, mapped the area as a kame (sand and/or gravel) complex.

Region 5

From aerial observation and air photograph interpretation, Region 5 appears to consist of bedrock overlain by stony silt and sand. This cover must be thin as bedrock structure is detectable through it. On small areas thicker drift completely obscures structure. Drainage is joint controlled.

Region 6

Region 6 occupies part of the upland surface of the northern portion of the island. It is underlain by flat-lying carbonates which are blanketed by a veneer of silty rubble. The rubble reflects the topography and structure of the underlying strata and, in that respect, is similar to Region 1 of Somerset Island. Many kame-like features are present but are bedrock cored and are covered with silty rubble. The drainage pattern shows no signs of derangement.

Region 7

Region 7 is a lowland on the eastern part of the island. The bedrock, sandstone, sandy shale, and carbonates, outcrops only in river cuts. The most abundant surficial material is till, a moderately stony silty sand. Overlying this, as a continuous thick deposit or as a basin filling, is light brown to buff coloured marine sand and silt which supports an extensive growth of black lichens. Massive ground ice is exposed in a few flow-slide scars in the marine deposits.

Boundaries between marine sediments and till are gradational, because where the thickness of the latter is less than the active layer thickness, the two materials have been mixed by cryoturbation. The drainage pattern is deranged, with numerous shallow lakes.

Region 8

The upland surface along the eastern side of Prince of Wales Island and adjacent islands constitutes Region 8. It is underlain predominantly by the Peel Sound Formation, sandstone grading upwards into conglomerate. The northeast corner of the island is underlain by gently dipping dolomite. The horizontally bedded bedrock has been dissected to produce a series of mesas. These are underlain by conglomerate while the lowlands are underlain by dolomite and sandstone. The only large areas of bare rock occur along scree-lined cliff faces at the junction of mesas and lowlands.

On the mesa tops the conglomerate has weathered to a coarse, well rounded, bouldery gravel felsenmeer. Remnant blocks and tors are found along ridges. Where sandstones and carbonates occur, the surface veneer is a pebbly silty sand which is difficult to distinguish from till. Drainage is joint controlled and radial from the plateau tops. A series of about 15 cirques or large nivation bowls, oriented northwest and containing snow or firn banks, overlook Baring Channel.

Quaternary History

Previous work

Perhaps the earliest summary report of the Quaternary geology of Arctic Canada is that of Craig and Fyles (1960). On the basis of the regional distribution of freshly glaciated landscapes, they tentatively placed the Laurentide boundary on Prince of Wales Island between the "southern lowlands with fresh glacial landforms and (the) northern highlands that were reported to bear only indefinite evidence of glaciation" (Craig and Fyles, 1960, p. 2 and Fig. 1). Following further investigations on Somerset and Prince of Wales islands, Craig (1964, p. 4) cited the following criteria as evidence that the "northern highlands ... were overridden by the Laurentide (continental) ice-sheet at the same time as the region to the south": (1) a thin layer of silty, rubbly glacial till covers most of the highlands; (2) Precambrian erratics are distributed extensively over most of Somerset and Prince of Wales islands; (3) ground moraine and isolated kame hills occur at several localities; and (4) abandoned stream channels that "could only have been formed if the streams responsible for them were confined originally along one or both sides by glacial ice" occur along and subparallel to larger valleys and across ridge tops. Craig also reported marine shells at three localities above that inferred highest Holocene shorelines; one of these shell samples has been radiocarbon dated at $31\,960^{+2560}_{-1940}$ years B.P. (GSC-123). He speculated that the shells had been moved to their present positions during the last major advance of Wisconsin ice.

Bird (1967) made many references to the geomorphology of the two islands. His conclusions were that the perimeter of the Wisconsin Laurentide Ice Sheet on Prince of Wales Island coincided with the boundary between the lowlands and the uplands, and that southern Somerset Island was inundated by the ice sheet, while the northern portion was covered by a thin ice body which was dynamically separate from the ice sheet (Bird, 1967, p. 95, Fig. 30).

No other analysis of field data pertaining to the glacial history of the two islands is available. Two nation-wide syntheses by Prest *et al.* (1967) and Prest (1969) graphically display the prevalent concepts concerning the study area. Both sources place the boundary of the late Wisconsin Laurentide Ice Sheet along the axis of Barrow Strait while the latter source shows a local ice cap occupying northern Somerset Island at 9000 years B.P. Bryson *et al.* (1969) invoke an almost identical history. Miller and Dyke (1974) synthesized the available information concerning the extent of late Wisconsin Laurentide ice on Baffin Island, which lies immediately east of the study area, and proposed that the ice sheet was not as extensive as the above mentioned syntheses suggested.

The above comprises the background to the current study of the glacial history of Somerset and Prince of Wales islands. The purpose of this section is to consolidate the information on Quaternary history contained in the terrain region descriptions and to put forward an initial synthesis of these data along with published information.

Regional ice-flow pattern

Figure 31.2 summarizes known directions of ice flow for the field area and surrounding regions. It is derived from the Glacial Map of Canada (Prest *et al.*, 1967) but has been modified slightly for the field area; new measurements of striae, grooves, and crag-and-tail features have been added and the previously suggested northward flow across northern Somerset Island has been removed because the authors believe that the lineations on which that was based are structurally controlled landforms rather than glaciogenic elements. The main features of the ice-flow pattern shown in Figure 31.2 are: (a) convergence of ice towards a line following the axis of the Gulf of Boothia, and (b) divergence of ice along a line from southern King William Island to central Prince of Wales Island.

Evidence from the study area

Erratics

Precambrian erratics up to boulder size are sparsely scattered on the Paleozoic upland of northern Somerset Island and on both lowland and upland surfaces of Prince of Wales Island. Likewise both Paleozoic carbonate and sandstone erratics are sparsely scattered over the Precambrian terrain of the Boothia Arch. The ubiquitous distribution of erratics on all surfaces examined shows that at some time the entire region has been glaciated. Several east-west traverses on northern Somerset Island failed to detect any change in

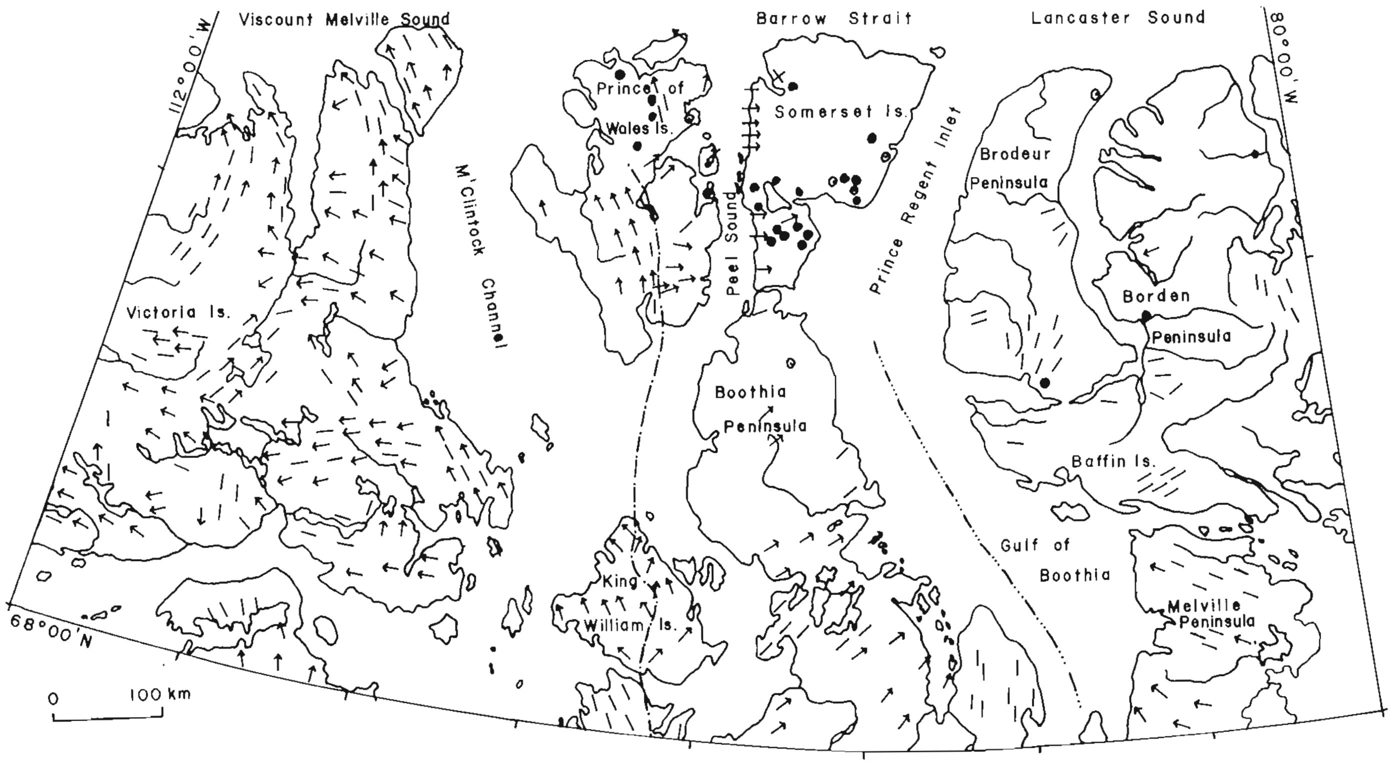

Figure 31.2. Regional Relationships. [Generalized ice-flow directions (→ sense known, – sense unknown) after Prest *et al.* (1967) and this report showing inferred axes of ice convergence (—•••—) and ice divergence (—•—). Marine shells (•) and fossiliferous marine deposits (⊙) above the postglacial marine limit after Craig (1964), Miller and Dyke (1974) and this report.]

frequency of their distribution. A source area cannot be delimited at this time. It is worth bearing in mind, however, that the Peel Sound conglomerate probably overlay the carbonates of Somerset Island and have since been eroded (J. Wm. Kerr, pers. comm., 1975). Some of the smaller Precambrian stones could be either true glacial erratics or a lag deposit derived from this formation.

Kames

On western Prince of Wales Island Craig (1964) mapped a zone of kames which appear to be ice-frontal. Scattered kames occur elsewhere on the island but may be confused with highly weathered bedrock knobs. On the uplands of northern and eastern Prince of Wales Island and on the Paleozoic plateau of Somerset Island, many features which had initially been interpreted as kames proved to be bedrock features.

On Somerset Island lines of kames occur south of Cresswell Bay and along both sides of Stanwell-Fletcher Lake valley. The latter mark a fairly well defined ice-marginal position and were deposited when a lobe of ice occupied the valley. East of Cresswell Bay a large kame moraine runs north-south; which of its sides is proximal is not known. South of Batty Bay two kames are in contact with marine or lacustrine deltaic gravel underlain by dipping sands and massive silts.

Grooves, roches moutonnées, striae, eskers

Little new information has been added to the known pattern of glaciation of Prince of Wales Island. Three new ice-flow directions are included on Figure 31.2. These are: (a) a northwestward flow across eastern Russell Island, (b) a north-northeastward flow on northeastern Prince of Wales Island, and (c) a northward flow across Prescott Island which is cross-cut by a later flow to the northeast or southwest.

Several additional striae orientations have been observed on Somerset Island. These, with one possible exception, record onshore movement of ice from Peel Sound. The possible exception is a set of striae oriented northwest-southeast near the northern coast of Aston Bay. It is unclear whether they were made by ice flowing down the bay from the island or by ice flowing onshore from northern Peel Sound. Craig (1964) recorded striae near this site with a northeast-southwest orientation. Except on the southern portion of the island striae, grooves, and roches moutonnées were found only near the western coastline. The same distribution of areas in which these features are preserved was noted by Blackadar (1967). Five eskers occur on western Somerset Island. Their orientations support the ideas outlined above.

Chronology

A relative glacial chronology for the field area has been devised. Samples have been collected for radiometric dating which will add absolute dating control and will provide a test of the ideas presented below. This tentative relative chronology is based upon observed morphostratigraphic relations between ice-marginal positions and raised marine sediments and upon the extent of alteration of the landscape by subaerial weathering processes during the periods since deglaciation.

Somerset Island

Phase 1. The earliest recognizable glacial phase is one of complete coverage by ice as shown by the ubiquitous distribution of erratics. The direction of ice flow remains unknown.

Phase 2. During a later period ice covered most or all of the southern peninsula of the island and probably terminated at the position marked by kame lines along the sides of Stanwell-Fletcher Lake basin. The kame moraine east of Cresswell Bay and the two kames south of Batty Bay probably were formed at this time as well. The silt deposits associated with the latter two kames may be contemporaneous with the ice margin at this locality. North of Cresswell Bay, high-level marine sediments were deposited after the ice withdrew. The highest distinct strandline features in the area of kames around Stanwell-Fletcher Lake extend up to 230 m above sea level. Near the kame moraine north of Cresswell Bay, marine shells at an elevation of 225 m have yielded an age of $31\ 860^{+2560}_{-1940}$ years B. P. (GSC-123). This is a likely minimum date on the phase 2 ice margin as the high-level strandlines must postdate this glaciation. Locations of sites where marine shells and sediments have been found at elevations above the inferred highest Holocene sea level stand, an inference which may change when additional data become available, are plotted on Figure 31.2. Similar observations from surrounding areas are added from other sources.

That a large portion of Somerset Island was free of actively eroding glacier ice during the phase 2 maximum is supported by recent observations of advanced weathering forms located on areas beyond the ice margin. Long periods of subaerial exposure of the landscape are indicated by: (a) tors developed on both gneissic and carbonate rocks and the related conical carbonated rock knobs; (b) extensive felsenmeer with associated grus; (c) widespread occurrence of solution features (relief ranging up to at least 15 cm) on carbonates; (d) occurrence of a large accumulation of pure grus derived from granite prior to marine inundation; (e) the large volume of silty rubble, which over at least much of the Paleozoic upland is a product of weathering rather than glacial deposition; (f) a well integrated drainage network showing no signs of derangement by glaciers; and (g) canyons eroded to depths of 100 to 150 m into the Paleozoic carbonates.

Later ice may have invaded southernmost Somerset Island but the authors can offer no stratigraphic proof of such an event. The length of time since deglaciation of the region immediately south of Cresswell Bay has been sufficient for formation of mature tors on the carbonate bedrock but apparently not on the adjacent gneisses. Time has been sufficient for generation of extensive felsenmeer on the gneisses and for erosion of canyons into the carbonates to a depth of 30 m. The surficial geology of northernmost Boothia Peninsula (Boydell *et al.*, 1975) appears to be similar to that of southernmost Somerset Island. The Quaternary geology of these latter two areas deserves further study.

Prince of Wales Island

Phase 1. The ubiquitous distribution of erratics leads to the same conclusion of complete ice coverage of Prince of Wales Island as for Somerset Island. It is likely that the periods of complete coverage were coincident.

Phase 2. The abrupt boundary between the area dominated by glacial landforms and the area where evidence of glaciation is much more rare, as well as the fact that the boundary coincides with an escarpment, suggest that the hypotheses of Craig and Fyles (1960) and Bird (1967) are valid. Furthermore, the upland surfaces of Prince of Wales Island contain evidences of subaerial weathering and possible marine inundation similar to those of Somerset Island. The authors do not imply any time correlation between phase 2 of one island and phase 2 of the other.

Future work

During the course of winter 1975-76 detailed air photograph interpretive maps will be prepared for the entire study area and samples will be analyzed with an eye to resolving some of the interpretive difficulties outlined above.

Acknowledgments

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References

Bird, J.B.
1967: The physiography of Arctic Canada (with special reference to the area south of Parry Channel); Johns Hopkins Press, Baltimore, Maryland, 336 p.

Blackadar, R.G.
1967: Precambrian geology of Boothia Peninsula, Somerset Island, and Prince of Wales Island, District of Franklin; Geol. Surv. Can., Bull. 151, 62 p.

Blackadar, R.G., Christie, R.L., and Taylor, F.C.
1967: Somerset Island and Prince of Wales Island, District of Franklin; Geol. Surv. Can., Map 3-1967.

Boydell, A., Drabinsky, .A., and Netterville, J.A.
1975: Terrain inventory and land classification, Boothia Peninsula and northern Keewatin; *in* Report of Activities, Part A, Geol. Surv. Can., Paper 75-1A, p. 393-396.

Bryson, R.A., Wendland, W.M., Ives, J.D., and Andrews, J.T.
1969: Radiocarbon isochrones on the disintegration of the Laurentide Ice Sheet; Arct. Alp. Res., v. 1, p. 1-14.

Craig, B.G.
1964: Surficial geology of Boothia Peninsula and Somerset, King William, and Prince of Wales Islands, District of Franklin; Geol. Surv. Can., Paper 63-44, 10 p.

Craig, B.G. and Fyles, J.R.
1960: Pleistocene geology of Arctic Canada; Geol. Surv. Can., Paper 60-12, 21 p.

Ives, J.D.
1962: Indications of recent extensive glacierization in north-central Baffin Island, N.W.T.; J. Glaciol., v. 4, p. 197-205.

Miller, G.H. and Dyke, A.S.
1974: Proposed extent of late Wisconsin Laurentide ice on eastern Baffin Island; Geology, v. 2, p. 125-130.

Prest, V.K.
1969: Speculative isochrones on the retreat of Wisconsin and Recent ice in North America; Geol. Surv. Can., Map 1257A.

Prest, V.K., Grant, D.R., and Rampton, V.N.
1967: Glacial Map of Canada, Geol. Surv. Can., Map 1253A.

Veillette, J.
1976: Assessment of terrain performance, Somerset and Prince of Wales islands, N.W.T.; *in* Report of Activities, Part A, Geol. Surv. Can., Paper 76-1A, rept. 59.

Project 730044

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Mapping commenced last year (Currie, 1975) in three localities in an attempt to gain new insight into the origin, method of emplacement and tectonic style of granitoid rocks. During the past field season mapping was continued at these localities and work was commenced in two new localities, one near Shelburne, Nova Scotia, and the other near Burlington, Newfoundland.

Progressive metamorphism, anatexis and metasomatism in the Cheticamp-Pleasant Bay region, Cape Breton Island

The rocks of this district fall into five main categories (Fig. 32.1): (1) post-orogenic rocks of Upper Devonian and/or Mississippian age (Fisset Brook Formation, and Horton Group); (2) late orogenic and post-orogenic granitoid rocks (Margaree and Cheticamp granites); (3) sedimentary and volcanic rocks first metamorphosed during the Acadian orogeny (Jumping Brook complex, substituted for Trout Brook complex of Currie, 1975); (4) older granitoid rocks on which the Jumping Brook complex was laid down; and (5) a presumed Precambrian basement complex containing schlieren and enclaves of diverse composition. Due to complexities of metamorphic and structural convergence, the relationships between these rocks have not been fully determined. Cataclastic rocks of post-Horton and older age commonly obscure the contacts between major rock groups. Although the derivation of some of these breccias and mylonites can be readily recognized, many

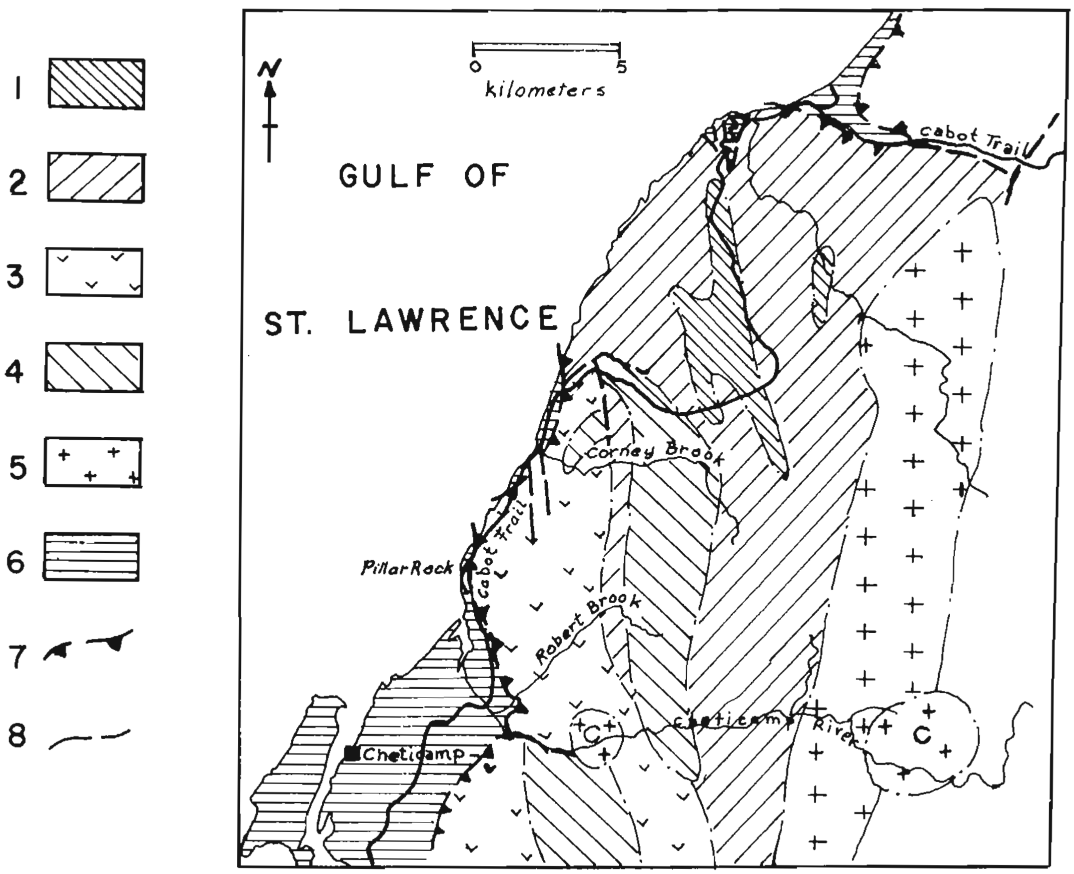

Figure 32.1. Geological sketch of the Cheticamp-Pleasant Bay region; (in part after Neale (1956) and MacLaren (1956). Units not in chronologic order). 1. Pleasant Bay complex, amphibolitic, paragneissic, and migmatitic portions; 2. Pleasant Bay complex, mainly biotite granodiorite portion; 3. Corney Brook complex, mainly hornblende granodiorite and alaskite; 4. Jumping Brook complex, mafic volcanics, greywacke, slate, and metamorphosed equivalents; 5. Post-orogenic granites, Margaree granite, and Cheticamp granite (C); 6. Upper Devonian and/or Mississippian rocks, Fisset Brook Formation and Horton Group; 7. Thrust fault, teeth on overthrust plate; 8. Fault.

Table 32.1

Chemical analyses of rocks from the Cheticamp-Pleasant Bay Region

	1	2	3	4	5	6	7
SiO_2	67.92	68.46	72.40	74.10	69.95	67.83	81.60
TiO_2	0.43	0.37	0.56	0.22	0.79	0.47	0.35
Al_2O_3	16.04	15.73	13.20	13.35	15.37	17.18	8.94
Fe_2O_3	0.75	1.71	1.50	0.70	0.25	0.19	0.75
FeO	2.54	1.60	1.15	0.62	3.15	1.94	1.35
MnO	0.09	0.13	0.05	0.03	0.12	0.04	0.05
MgO	1.15	1.21	0.65	0.33	1.30	0.79	1.08
CaO	2.19	2.51	0.48	0.48	1.89	1.80	0.31
Na_2O	4.33	3.63	3.25	3.35	4.80	5.32	1.32
K_2O	3.39	3.02	5.88	5.84	1.67	3.54	3.33
H_2O	0.83	1.37	0.75	0.40	0.95	0.42	0.80
P_2O_5	0.16	0.16	0.15	0.05	0.06	0.19	0.08
TOTAL	99.82	99.90	100.02	99.57	100.29	99.71	99.96

[1]Granitoid component of Pleasant Bay complex (average of 4)

[2]Hornblende granodiorite of Corney Brook complex (average of 2)

[3]Margaree granite (average of 4)

[4]Cheticamp granite (average of 3)

[5]Psammitic portion of Jumping Brook complex (average of 5)

[6]Granite gneiss developed from (5) (average of 3)

[7]Foliated cataclasites (average of 2)

Analyses by Geological Survey of Canada, Rapid Methods Group, 1975.

of the cataclasites are dark brown to green, compact, fine grained to aphanitic rocks with varying degrees of fissility, whose derivation is obscure. Because of their strong fissility some of these rocks were erroneously mapped last year as part of the Jumping Brook complex. In fact they seem to occur along the complex series of coastal faults characteristic of the northwest coast of Cape Breton Island. Preliminary petrographic and chemical data (Table 32.1, Analysis 7) suggest that they were derived from very quartz-rich rocks (SiO_2 >80 per cent), and that they have undergone a complex history, including ubiquitous late potash metasomatism.

At Pillar Rock the whole of the Jumping Brook and Corney Brook complexes can be seen to over-ride Horton Group strata along a thrust fault dipping inland at about 20^o. The older rocks are strongly shattered within about 300 m of the contact, and vestiges of this shattered zone can be traced at least 5 km inland in the valley of Robert Brook. The presence of numerous gypsum and barite veins in older rocks suggests that evaporites of the Windsor Group may be over-ridden somewhere in this interval. The cataclastic rocks described above thus appear to represent small scale disruption at the leading edge of a major thrust sheet which has carried all older rocks over Carboniferous strata. The full extent of this movement is unknown, but much of the western Cape Breton Highland may be allochthonous.

The Jumping Brook complex occurs within a single refolded synform. Metamorphic grade rises steadily across this structure from west to east, demonstrating that metamorphism post-dated at least two phases of folding. At least four isograds can be mapped within a pelitic sequence, and electron probe work by G.J. Pringle shows the presence of two other zones not recognized in the field. The most complete sequence of type minerals appears in Trout Brook and the upper part of Corney Brook, where the sequence is (diaspore/boehmite), chlorite, muscovite, epidote, biotite, garnet (chloritoid), staurolite, cordierite, andalusite. Phases in brackets were not recognized in the field, but were found during analytical work. In the upper part of Corney Brook the sequence from first appearance of biotite to first appearance of andalusite is almost completely exposed over a distance of 3 km. At the highest grade of metamorphism observed, the Jumping Brook complex is gneissic (rather than schistose) and distinctly pinkish. Each of the three distinctive portions of the complex converge toward a distinctive granitoid rock, the pelitic parts toward a porphyroblastic rock rich in muscovite and potash feldspar, the greywackes toward a gritty biotite-muscovite granite gneiss (Table 32.1, columns 5, 6), and the mafic volcanics toward a massive amphibolite with potash feldspar porphyroblasts. In addition to these locally developed granitoid rocks, a ubiquitous biotite-muscovite alaskite, commonly gneissic and leaf-textured, forms sills and schlieren in both the Jumping Brook rocks, and the older Pleasant Bay complex. All of these rocks are clearly older than the latest formation of gneissosity.

Two plutons clearly post-date this gneissosity. One lies in the valley of the Cheticamp River (Cheticamp granite) and comprises a distinctive reddish pink massive, medium grained granite with stellate amphibole and biotite. The exact form of this pluton is uncertain due to faulting, but it seems to form a subcircular mass about 3 km in diameter. The other late pluton is the distinctive megacrystic granite previously described (Currie, 1975) which forms the spine of the western highlands, extending more than 30 km from north to south (Margaree granite). As noted previously (Currie, 1975), there is strong evidence for metasomatism in the formation of this rock, although its form seems to require a basically igneous origin. Despite the diverse form, texture and size of the Cheticamp and Margaree granites, they are chemically indistinguishable (Table 32.1, columns 3, 4), which together with their distinctive age relations seems to imply some kind of genetic link.

At least three ages of granite generation seem to be indicated in this region. The generally granitoid matrix of the Pleasant Bay complex clearly pre-dates the other granitoid rocks. Metamorphism of the Jumping Brook complex led to development of granitoid rocks, but apparently on a minor scale. The emplacement of the Margaree and Cheticamp plutons clearly post-dates these granitoid rocks, and may post-date all other structural deformation in the region. The source of this late magma does not seem to have been close to the mapped region, and presumably lay at depth.

Metamorphism of the Meguma Group near Shelburne, Nova Scotia

The Lower Paleozoic Meguma Group of slate and fine grained greywacke, together with Devonian granitoid rocks, here collectively termed the Southern Nova Scotia batholith, comprise much of Nova Scotia south of the Chedabucto Fault. In the northeastern part of this area the granitoid rocks intrude the sedimentary sequence, producing hornfelsic aureoles characterized by andalusite and/or cordierite in pelitic rocks (Taylor and Schiller, 1966). Taylor and Schiller (1966) first pointed out that around Shelburne on the southern tip of Nova Scotia the Meguma Group has been regionally metamorphosed to amphibolite facies, and that migmatitic rather than hornfelsic aureoles occur around some plutons, while other plutons display no apparent contact effects. The region yields anomalously young potassium argon ages on granitoid rocks (Taylor, 1967) suggesting that the plutons remained above the K-Ar "blocking temperature" longer than those to the north. The occurrence of the experimentally studied assemblage garnet-cordierite-sillimanite-quartz (Currie, 1971; Hutcheon *et al.*, 1973; Hensen and Green, 1971) suggests that peak metamorphic conditions must have been close to those at which granitoid melts are generated from suitable hosts in the presence of excess water, while the presence of widespread primary muscovite suggests that large amounts of water were present. The Shelburne region may therefore display relicts of generation of anatectic granitoid rocks, and offer clues to the relation between such rocks and larger bodies of granitoid rocks.

High grade metamorphic rocks occur within a triangular area roughly 130 km from east to west by 90 km from north to south, bounded to the south and east by the Atlantic Ocean. To the west high grade rocks terminate abruptly against lower grade slates on the west side of Pubnico Harbour. Anomalous shearing, abrupt juxtaposition of different metamorphic grades, and slight discordance in strike make it probable that this contact is a fault. Taylor (1967, 1969) determined that the northern boundary is about latitude 44°, but due to poor outcrop nothing is known of the nature of this contact.

Three major units of metamorphic rocks can be distinguished. A monotonously uniform gritty, grey weathering, buff to olive quartz-feldspar-biotite gneiss, correlated by Taylor (1967) with the Goldenville Formation of the Meguma Group, forms most of the outcrop. Large outcrops commonly display necked down layers or boudins of pale grey, very fine grained granitoid rocks, with characteristic stellate, blastophitic amphibole. Three km west of Jordan Falls the cross-cutting nature of the granitoid rocks is partially preserved. They evidently represent dykes and/or sills deformed and metamorphosed with their host.

A second major unit contains varying amounts of the gneiss just described together with more pelitic layers. Near Middle Ohio psammitic and pelitic layers intercalate on scales that vary from a few centimetres to tens of metres. Although mapped as "metamorphosed equivalent" of Goldenville Formation by Taylor (1967), this unit generally shows more aluminous compositions than the typical quartz-feldspar-biotite gneiss, passing from mesocratic gneiss with scattered pin-head garnets, through muscovite- and/or chlorite-rich gneisses and muscovite-rich andalusite schists, to muscovite-rich schists containing spectacular porphyroblasts of andalusite, cordierite and biotite up to 10 cm in length. These spectacular pelitic schists form a distinctive third unit, correlated by Taylor (1967) with the Halifax Formation. With increasingly aluminous compositions the schists contain less quartz-feldspar-biotite gneiss, and the most aluminous rocks contain only small garnetiferous boudins. The transition from psammatic to pelitic compositions corresponds to the gradual transition from Goldenville to Halifax Formation observed elsewhere by Taylor (1967).

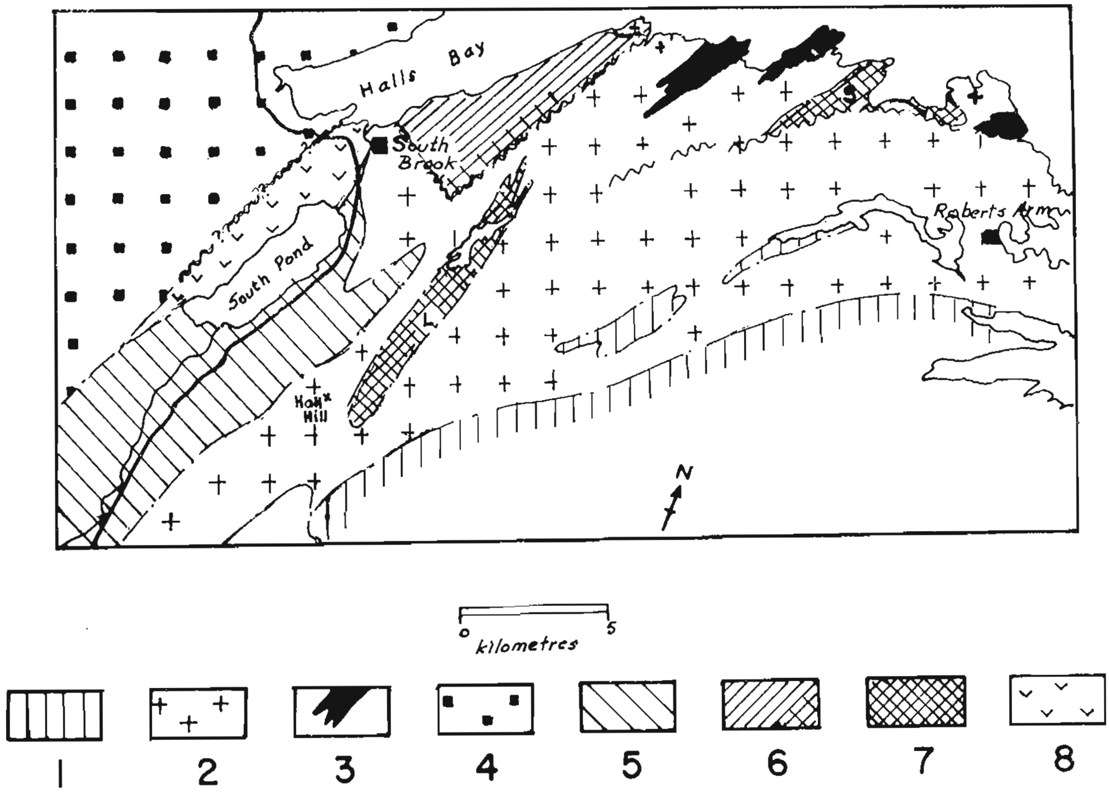

Figure 32. 2. Geological sketch of the South Brook-Roberts Arm area; (after Neale and Nash (1963), Williams (1964), Bostock (1975)). 1. Exploits Group, shale greywacke, conglomerate; 2. Roberts Arm Group, mafic portion, pillowed, massive and pyroclastic basalt; 3. Roberts Arm Group, felsic portion, rhyolite, felsite, dacite; 4. Springdale Group, mainly acid pyroclastic rocks; 5. Hall Hill complex, gabbro, hornblendite, agmatite of gabbro, basalt, and aplitic rocks; 6. Mansfield Cove block, mainly granodiorite and tonalite, cross hachured portion is younger felsitic and dioritic rocks; 7. Sunday Cove (S) and Loon Pond (L) plutons, inclusion-rich aplitic syenite, quartz syenite and felsite; 8. South Pond pluton, alaskitic granite.

In the most characteristic and spectacular pelitic rocks large, unoriented idioblasts transect a weakly schistose muscovitic matrix, giving the rock a superficially massive appearance, even though compositional banding is often preserved. Deformed or rolled porphyroblasts were observed only along Pubnico Harbour, where late shearing has converted some andalusite to disc-like masses.

A small amount of a distinctive variety of gneiss occurs within and around the major plutons. These coarse grained, uniform, grey rocks exhibit rather nebulous centimetre-scale banding, with some bands containing garnet, garnet and muscovite, or rarely, garnet-cordierite-sillimanite-quartz. The gneiss commonly shows plastic ptygmatic style deformation, but I found neither clear-cut truncation by granitoid rocks, nor inclusions of gneiss in granitoid rocks. Leucocratic parts of the gneiss invariably contain nebulous seams and masses of pegmatitic muscovite granite identical to those cutting the grey granodiorite of the major plutons, but dissection of boulders showed that most, if not all, of these masses are rootless, and must have formed *in situ.* The common grey gneiss of the matrix appears to grade into the ubiquitous grey granodiorite of the southern Nova Scotia batholithic complex, although all phases of this gradation cannot be observed at any one locality.

Taylor (1967) considered these migmatitic gneisses to be correlative with the Meguma Group, even though their textures differ markedly from those of the more common metamorphic rocks. The presence of sillimanite rather than andalusite in some bands suggests slightly higher temperatures and/or pressures. The presence of "sweats" in muscovitic parts of the gneiss strongly suggests that partial melting occurred in compositionally favourable portions. Under these conditions, the coarsening of grain size and development of plasticity may have been due to a convergence of physical properties between the compositionally similar Goldenville Formation and the granodioritic rocks of the Southern Nova Scotia batholith. The "floor" of the Goldenville Formation, which has never been observed, may be marked by gradual transition to granodiorite. Whether this transition represents intrusion of material from below into fortuitously similar compositions, or whether there is a genetic link between Goldenville Formation and granodiorite, cannot be decided without detailed chemical and isotopic studies. The undeformed character of the regionally developed pelitic assemblages suggests that a salient feature of the emplacement of the granitoid rocks was a rise in temperature in the absence of strong deformation.

A model for the origin of granitoid rocks in the Roberts Arm-South Brook area, Newfoundland

At least three different types of granitoid rocks outcrop in a small area near South Brook (Currie, 1975). The oldest component, termed the Mansfield Cove block (Fig. 32.2), consists mainly of grey gneissic tonalite with distinctive large quartz blotches. A variety of other rock types occur including an alaskitic pink phase of the tonalite, small areas of *lit par lit* gneisses, innumerable dykes of Roberts Arm basalts, and two late intrusive phases described below. The Mansfield Cove block displays well developed cataclastic foliation marked by smeared mafic minerals and local quartz augen. In sporadic 3- to 10-metre-thick zones this foliation passes to intense deformation with development of mylonite and flaser gneiss. Since this north-trending cataclasis does not parallel any other recognized structural element, it appears to be relict from some older event. The fault bounding the Mansfield Cove block to the southeast brecciates and shears Roberts Arm volcanic rocks to the southeast, but does not affect the felsite and diorite along the margin of Mansfield Cove block. The red, quartz-rich felsite veins and alters the tonalite, but the relationships of the massive, pale grey hornblende diorite are uncertain. Both units may be younger than the faulting, and related to the Loon Pond complex and the Hall Hill complex respectively. The southwestern margin of the Mansfield Cove block is deeply drift-covered over most of its length, but the southern corner of the block displays both northeast and northwest trending faults, suggesting that faults bound the block both to the southeast and southwest.

The volcanic Roberts Arm Group contacts the Mansfield Cove block. The lower (southeastern) part of this formation consists mainly of submarine basalts, with increasing amounts of felsic rocks to the northwest (Bostock, 1975). The Sunday Cove and Loon Pond plutons form very similar concordant plutons of fine grained pink syenitic rocks full of strongly assimilated mafic inclusions. The Sunday Cove pluton displays transitions to felsitic sheets within the volcanic pile on the one hand, and to rocks similar to Mansfield Cove alaskite on the other. Detailed study of the Loon Pond pluton showed it to be a silt-like mass about 1000 m thick and more than 8 km long. At its northern end it splits into three separate sills of rhyolitic appearance, two of which clearly cross-cut the surrounding mafic volcanics. These relationships suggest that considerable parts of the "acid volcanics" within the Roberts Arm Group may be intrusive. Such intrusive rocks are presumably less likely hosts of sulphide deposits than their extrusive counterparts, which may help explain the relatively disappointing prospecting results within the Roberts Arm.

Over a length of 8 km the Loon Pond pluton shows little change in character, although the number of inclusions appears to decrease southward. Although the Loon Pond pluton closely resembles the Sunday Cove pluton petrographically, and presumably petrologically, it does not form its faulted extension, as suggested last year. However the Loon Pond pluton does mark a subtle but significant change in the character of the rocks. To the west, fine grained mafic rocks show few or no volcanic features, and the proportion of gabbroic and dioritic rocks increases. Although many of the rocks are indistinguishable from the Roberts Arm Group, this mixture of basalt and gabbro is sufficiently distinctive that I informally term it the Hall Hill complex. Around South Brook the complex consists mainly of basaltic rocks with numerous gabbroic sills, which in some cases are themselves cut by basaltic rocks. Farther south around Hall Hill, basalt and gabbro compose equal amounts of the outcrops. Locally the rocks consist of alternating dykes of gabbro and basalt, while in others gabbro is cut by swarms of basaltic dykes, while in still other localities massive to schistose, possibly extrusive, basalt is cut by gabbro dykes. Along and west of the Trans-Canada Highway the situation is complicated still further by the admixture of Loon Pond type aplite, rendering most outcrops agmatitic. Many of the gabbros in this region appear quite fresh, but they invariably contain large hornblende crystals and grade to hornblendite. Whether this feature is primary or secondary remains to be determined. A distinctive type of granitoid rock consisting of aplitic plagioclase and quartz occurs in these agmatites. This rock type grades abruptly to gabbro in the agmatites, but a mass some hundreds of metres across outcrops east of Hall Hill.

The South Pond pluton forms an elongate homogeneous mass of leucocratic pink, medium grained, biotite granite, lacking either inclusions or dykes. A "near contact" with the Springdale Group is exposed on a hillock 3 km southwest of South Brook village. The granite becomes finer grained and porphyritic, with some shearing parallel to the contact, while the Springdale Group shows no signs of metamorphism or dykes, although some shearing is present. The fragmental volcanogenic rocks of the Springdale Group contain fragments of Roberts Arm type and Loon Pond type, but not of South Pond type. In gross aspect the South Pond-Springdale contact is perfectly linear, and may be faulted. The eastern contact of the South Pond pluton is totally drift covered in the valley of South Pond, which itself forms a continuation of Halls Bay. However at the south end of South Pond, Hall Hill rocks appear west of South Brook, showing that the South Pond pluton must be lenticular in shape, and suggesting that its eastern margin is not faulted. Since this pluton has escaped the pervasive shattering and alteration that characterizes the Roberts Arm Group and the Loon Pond pluton, it must presumably be younger, but no direct observational evidence has been found. The South Pond pluton seems to sit in splendid isolation from surrounding units.

Direct observation demonstrates the approximate contemporaneity of Loon Pond, Roberts Arm and Hall Hill magma. On one very favourable outcrop gabbro is broken up and veined by Loon Pond type aplite, which is itself cut by a basalt dyke of the amygdaloidal type typical of the Roberts Arm. This dyke is in turn boudined by the gabbro, hence all three components were mobile approximately simultaneously. Together with the observation that the Sunday Cove pluton grades to rocks of Mansfield Cove type, these facts suggest a speculative but plausible petrogenetic model. According to this model, the Mansfield Cove block forms a fragment of a sialic basement, presumably widespread beneath the western part of the Roberts Arm Group. This basement may be either Precambrian, or a relict of some earlier Paleozoic event as suggested by Williams and Payne (1975) for the Twillingate pluton, with which the Mansfield Cove block shares many petrographic features. The Roberts Arm Group was erupted through this basement along fissures now represented by the hundreds of dykes in the Mansfield Cove block. Beneath the current level of volcanic eruption, basalt magma was also emplaced as dykes, sills and plugs. I identify this zone with the northern part of the Hall Hill complex, which must thus be older than the pillowed, unintruded basalts to the east. Bostock (pers. comm.) observed that although the Roberts Arm Group generally faces west-northwest, in the vicinity of Loon Pond the facing reverses, so that the section from Loon Pond to the Trans-Canada Highway may represent a continuous east-facing section more than 5 km thick. As basaltic eruption proceeded, more and more heat leaked from the eruptive fissures into the surroundings until (hypothetically) the melting point of the granitoid rocks was reached along the eruptive fissures. These melts could maintain a separate identity and not be assimilated for the reasons outlined by Yoder (1973). The coexistence of basic and acid melts helps to explain the great numbers of highly assimilated mafic inclusions in the Loon Pond and Sunday Cove plutons, a feature otherwise difficult to understand in such high-level plutons. The confinement of acid rocks to the upper part of the Roberts Arm Group follows, on this model, from the long time required for their generation. Since acid melts are notoriously difficult to erupt as flows, the emplacement of these early granitoid melts as plutons at an approximate isotherm defined by emplacement of gabbro sills is not surprising.

If eruption of Roberts Arm rocks continued over a sufficiently long period, heat transfer to the sialic basement would eventually reach a point where wholesale melting would intervene, and gravitational instability would ensue. A first symptom of such instability might be upward migration of granitoid melt, breaking up and veining the overlying rocks. The agmatitic part of the Hall Hill complex fits this description. The full development of gravitational instability would lead to collapse of the overlying basalt, and eventual large scale eruption of the underlying acid magma. These stages can be identified respectively with the upturning and faulting of the Roberts Arm (and Hall Hill) rocks as described by Bostock (pers. comm.), and the subsequent emplacement of the Springdale Group, and presumably the South Pond pluton. Although many parts of this hypothesis are admittedly speculative, they are subject to chemical and isotopic testing, now in progress, which will hopefully test the model.

Notes on the Burlington pluton and nearby igneous rocks, northwestern Newfoundland

Some of the best exposed ophiolites in the world occur on the southeastern shore of Burlington Peninsula at Betts Cove and Nippers Harbour. Although the ophiolites are now well known and well described, divergent views abound on their emplacement and relations to surrounding rocks (*see* DeGrace *et al.*, 1975 for a brief summary of the debate). All the systematic field mappers agree that the Snooks Arm Group, of which the ophiolites form the lowest exposed part, is overlain with slight or negligible unconformity by the Cape St. John Group, mainly of basic and acid volcanic rocks, and that the Burlington pluton, which intrudes both, is in turn intruded by siliceous porphyrites related to the Cap Brule pluton, the youngest rocks in the region (Snelgrove, 1931; Baird, 1951; Neale, 1958a; DeGrace *et al.*, 1975). Since Lower Ordovician graptolites have been recovered from the Snooks Arm Group, both plutons must, on the basis of this model be younger, possibly much younger. The contrary view, that the Snooks Arm Group forms the youngest rather than the oldest rocks, rests on various structural arguments not germane to the present study (*see* DeGrace *et al.*, 1975 for details). However observation of plutonic debris in the Snooks Arm Group presumably derived from the Burlington pluton seems to give direct stratigraphic support to this argument. On this model the Burlington pluton, and all rocks which it intrudes, are pre-Ordovician. Since the Cap Brule pluton is deformed with the Cape St. John Group in its northern parts, yet intrudes the ophiolites in its southern parts, this model implies that this "pluton" must in fact be formed of at least two parts of markedly differing ages.

A test of these divergent views can be made on purely petrographic grounds, namely: (1) Is there another source for the Snooks Arm debris besides the Burlington pluton? and (2) is the Cap Brule pluton a single and distinctive unit? If the answer to both questions is yes, then the first hypothesis seems the only viable one, while if both questions are answered no, then the second hypothesis is upheld. With these questions in mind I examined the Burlington pluton and porphyries intrusive into it near the village of Burlington, and briefly looked at the main mass of the Cap Brule pluton under the guidance of John DeGrace.

Just west of Burlington (Fig. 32.3), the Burlington pluton consists of near massive, medium- to coarse-grained rocks weathering to dull pinkish buff. On fresh surface the rocks appear creamy, often with pinkish tinges. The rocks consist of roughly equal amounts of quartz, saussuritized plagioclase and hornblende, with biotite and potash feldspar very rare or absent. To the west the rock becomes grey-green as it acquires an increasingly prominent north-trending cataclastic

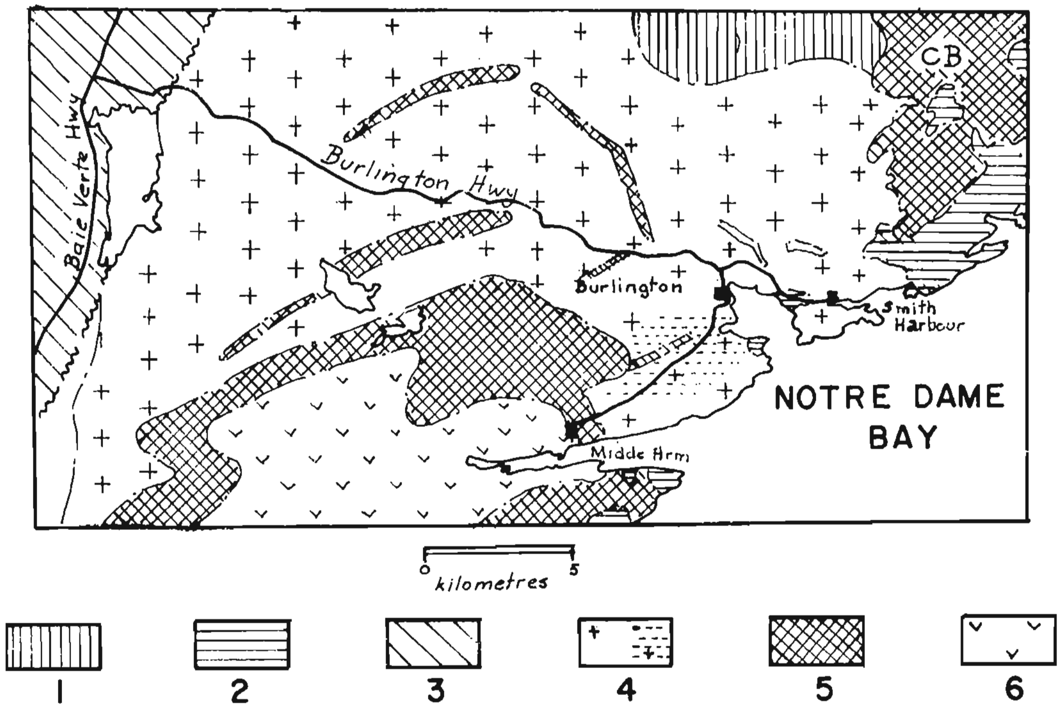

Figure 32. 3. Geological sketch of the Burlington area; (after Neale and Nash (1963), Williams (1964), DeGrace *et al.* (1975)). 1. Cape St. John Group and correlative rocks, mainly acid and mafic volcanics; 2. Snooks Arm Group and correlative rocks, pillowed mafic flows, sheeted dikes, ultramafics; 3. Baie Verte Group, mainly deformed mafic volcanics with ultramafic lenticles; 4. Burlington Pluton, cream to pink, massive to gneissic granodiorite. Dotted portion shows alkali metasomatism; 5. Feldspar and quartz-feldspar porphyries, minor rhyolite. CB = Cap Brule pluton, which also contains granitic phases; 6. Syenite, quartz syenite, with porphyritic and felsitic border phases.

foliation. The pluton is eventually truncated by a major fault which separates it from intensely deformed mafic volcanic rocks of the Baie Verte Group. To the east, near Smith Harbour, the rock acquires a southeast trending gneissosity, and commonly displays minor amounts of biotite. To the south, the rocks have been invariably metasomatized with development of reddish, quartz-poor, potash feldspar-rich rocks. South of Burlington village such rocks intrude and vein pillowed mafic lavas, apparently correlative to the Snooks Arm Group (Neale, 1958b).

The Burlington pluton contains a considerable number of small rounded inclusions. The most common are variously assimilated fine- to medium-grained biotite amphibolites, presumably derived from mafic volcanic rocks. Medium-grained, massive, brown diorite inclusions also occur, and no source for these is apparent. Diorites in the Baie Verte Group (Neale, 1958b) strikingly resemble the Burlington pluton. A sill outcropping on the road to Burlington a few tens of metres east of the Baie Verte highway is indistinguishable from the edges of the main pluton a few hundred metres farther east. Medium grained pink to brown aplitic dykes cut the pluton east of Burlington and are themselves truncated by younger porphyries. They may represent some late stage of the Burlington pluton.

Porphyries with a fine grained to aphanitic matrix and well defined euhedral phenocrysts intrude the Burlington pluton. In some cases a chill zone extends from the contact up to 30 cm into the porphyry, while a grey green granofels zone may extend up to 100 m into the host. In other cases the contact displays no chill, but development of chlorite and hematite stain is widespread, and inclusions in the porphyry are intensely assimilated. This latter case is spectacularly illustrated by the main mass of the Cap Brule pluton which has assimilated large amounts of mafic and ultramafic rocks of the Snooks Arm Group. The mass of porphyries centred southwest of Burlington display a metasomatic halo similar to fenitization. Metasomatism appears mainly along northeast trending fracture zones occupied by rhyolite dykes, but the bulk of the rocks have also been affected to a considerable extent.

The complex of porphyry dykes, sheets and stocks exhibits distinct temporal and compositional zoning. The oldest and most siliceous phases appear at the margins, passing inward through felsitic phases to a central syenitic core, locally quite coarse grained. By contrast the Cap Brule porphyry exhibits no marked concentric zoning, but gradational and patchy variation (DeGrace, pers. comm.). All phases of the pluton are fairly rich in quartz. Textures are extremely complex, with fracture and resorption of phenocrysts occurring several times.

The youngest intrusive rocks in this region comprise a northeast trending swarm of mafic dykes varying from 10 cm to 30 m in width. The dykes are commonly gabbroic in composition, with rounded feldspar megacrysts to 3 cm, and primary sulphide contents as high as 5 per cent. In the vicinity of Middle Arm the dykes form 3-6 per cent of the volume. The dykes are quite fresh in this area but become epidotized and weakly foliated to the northwest, and schistose to the northeast. Contacts with other rock units are sharp, although contacts with the youngest porphyries appear lobate and irregular, suggesting that the host may still have been plastic on emplacement.

The rocks show little "structure" in the usual sense, but there is a strong northeast trending fracture system overprinted on an earlier north-northeast trending cataclastic foliation. Strongly attenuated shear fractures suggest southeast-northwest compression.

The current investigation shows that the Burlington pluton intruded mafic volcanic rocks, reasonably correlated to the Snooks Arm Group, and a diorite, now found only as inclusions. Further the pluton itself is either composite, or displays a later, cross-cutting more potassic intrusion. Hence there seems to be some evidence that plutonic debris in the Snooks Arm Group could have come from rocks older than the Burlington pluton, or at least from older phases of the pluton. The porphyries intrusive into the pluton are all high level intrusions, but differ considerably among themselves. Although a reasonable hypothesis, it is not obvious that the Cap Brule pluton and the other porphyries share a similar origin. The unity (or lack of it) of the Cap Brule pluton therefore remains a purely mapping problem, now being attacked by John DeGrace. The emplacement of the porphyries was accompanied or preceded by intense metasomatic activity both outside the porphyries and within them (assimilation). The scale of metasomatism appears unusual in a high-level, presumably "dry" intrusion. This phenomenon is being further investigated.

References

Baird, D.M.
1951: The geology of the Burlington Peninsula, Newfoundland; Geol. Surv. Can., Paper 51-21, 70 p.

Bostock, H.H.
1975: Volcanic rocks of the Appalachian province: Roberts Arm Group, Newfoundland; *in* Report of Activities, Part A, Geol. Surv. Can., Paper 75-1A, p. 1-3.

Currie, K.L.
1971: The reaction 3 cordierite 2 garnet 4 sillimanite 5 quartz as a geological thermometer in the Opinicon Lake region, Ontario; Contrib. Mineral. Petrol., v. 53, p. 215-226.

1975: Studies of granitoid rocks in the Canadian Appalachians; *in* Report of Activities, Part A, Geol. Surv. Can., Paper 75-1A, p. 265-270.

DeGrace, J.R., Kean, B.F., Hsu, E., and Besaw, D.M.
1975: Geology of the Nippers Harbour map area (2E/13), Newfoundland; Govt. of Nfld., Dept. Mines, Energy File No. Nfld. 788, 59 p.

Hensen, B.J. and Green, D.H.
1971: Experimental study of cordierite and garnet in pelitic compositions at high pressures and temperatures; I. Compositions with excess aluminum silicate; Contrib. Mineral. Petrol., v. 3, p. 309-330.

Hutcheon, I., Froese, E., and Gordon, T.M.
1973: The assemblage quartz-sillimanite-garnet-cordierite as an indicator of metamorphic conditions in the Daly Bay complex, N.W.T., Contrib. Mineral. Petrol., v. 44, p. 25-34.

MacLaren, A.S.
1956: Cheticamp River, Inverness and Victoria Counties, Nova Scotia; Geol. Surv. Can., Paper 55-36.

Neale, E.R.W.
1956: Pleasant Bay; Geol. Surv. Can., Map 1119A.

1958a: Nippers Harbour, Nfld.; Geol. Surv. Can., Map 22-1958.

1958b: Baie Verte, Newfoundland; Geol. Surv. Can., Map 10-1958.

Neale, E.R.W. and Nash, W.A.
1963: Sandy Lake (east half), Newfoundland; Geol. Surv. Can., Paper 62-28, 40 p.

Snelgrove, A.K.
1931: Geology and ore deposits of Betts Cove-Tilt Cove area, Notre Dame Bay, Newfoundland; Can. Inst. Min. Met. Bull., v. 24, p. 477-519.

Taylor, F.C.
1967: Reconnaissance geology of Shelburne map-area, Queens, Shelburne and Yarmouth Counties, Nova Scotia; Geol. Surv. Can., Mem. 349, 83 p.

Taylor, F.C. (cont'd.)
1969: Geology of the Annapolis-St. Mary's Bay map-area, Nova Scotia; Geol. Surv. Can., Mem. 358, 65 p.

Taylor, F.C. and Schiller, E.A.
1966: Metamorphism of the Meguma Group of Nova Scotia; Can. J. Earth Sci., v. 3, p. 959-974.

Williams, H.
1964: Botwood, Newfoundland; Geol. Surv. Can., Map 60-1963.

Williams, H. and Payne, J.G.
1975: The Twillingate Granite and nearby volcanic groups: an island arc complex in northeast Newfoundland; Can. J. Earth Sci., v. 12, p. 982-995.

Yoder, H.S.
1973: Contemporaneous basaltic and rhyolitic magmas; Am. Mineral., v. 58, p. 153-171.

MEALY MOUNTAINS COMPLEX, GRENVILLE PROVINCE, SOUTHERN LABRADOR

Project 750011

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Regional and Economic Geology Division

The Mealy Mountains complex is the largest anorthositic complex in the northeastern part of the Grenville Province. The northwestern margin of the complex lies some 95 km south of the Grenville Front (Fig. 33. 1). It is the nearest neighbour to the similar major anorthositic complexes lying north of the Grenville Front in central and northern Labrador. Because of proximity to these unmetamorphosed plutons north of the Front, investigation of the rock types, mineralogy and structures of the Mealy Mountains complex, presumably involved in the Grenvillian Orogeny, is of special interest. Comparative studies should yield valuable information on the character of Grenvillian metamorphism and tectonism in this part of the Grenville Province.

The Mealy Mountains complex lies within the Battle Harbour-Cartwright area mapped on a reconnaissance scale by Eade (1962). Petrographic data on some rocks of the complex are given by Gillett (1956). Gittins (1972) reported whole-rock K-Ar ages for two very fine grained basalt dykes (average about 1100 m. y.) that postdate the complex. Fahrig *et al.* (1974) have reported on a paleomagnetic study of rocks of the complex.

Regional Setting

Although there has been some discussion about the exact position of the Grenville Front in this region, there can be no doubt that the Mealy Mountains complex lies well within the Grenville Structural Province. Rocks surrounding the complex consist dominantly of granitoid gneisses, in part migmatitic, containing pods and lenses of amphibolite, quartzite and paragneiss. Inclusions of these country rocks occur within the major rock units of the Mealy Mountains complex. The general regional metamorphic grade is amphibolite facies. K-Ar biotite ages in the region fall in the range of about 950 to 1000 m. y. presumably reflecting the closing stages of the Grenvillian Orogeny. The depositional ages of most rocks in the region are unknown.

The major basic rock units of the Mealy Mountains complex consist of leucotroctolite, leucogabbro and anorthosite. These rocks are intruded by large and small masses of pyroxene quartz monzonite which comprise the silicic rocks of the complex.

A swarm of olivine diabase dykes that range in width from less than 1 cm to 60 m, laces the Mealy Mountains complex. The dykes typically have chilled selvages and the smaller ones are fine grained but large dykes have central parts with grain sizes up to 2 or 3 mm and pegmatitic patches occur locally within them.

The Double Mer arkosic sandstone of presumed Hadrynian age outcrops on the north shore directly across Lake Melville from the Mealy Mountains complex. Two clastic dykes of brown arkosic sandstone (largest about 15 cm wide) fill sharp-walled fractures in anorthosite about 3 m above high tide level. If these dykes can be correlated with the Double Mer sandstone, it would imply that the present north and south shores of Lake Melville were at about the same relative elevation at the time of deposition of the Double Mer, assuming such clastic dykes are incapable of penetrating to great vertical depths. A corollary is that if the northwest-facing escarpment of the Mealy Mountains is fault-bounded and not an erosional feature, any significant vertical movement on the fault must have preceded deposition of the Double Mer sandstone.

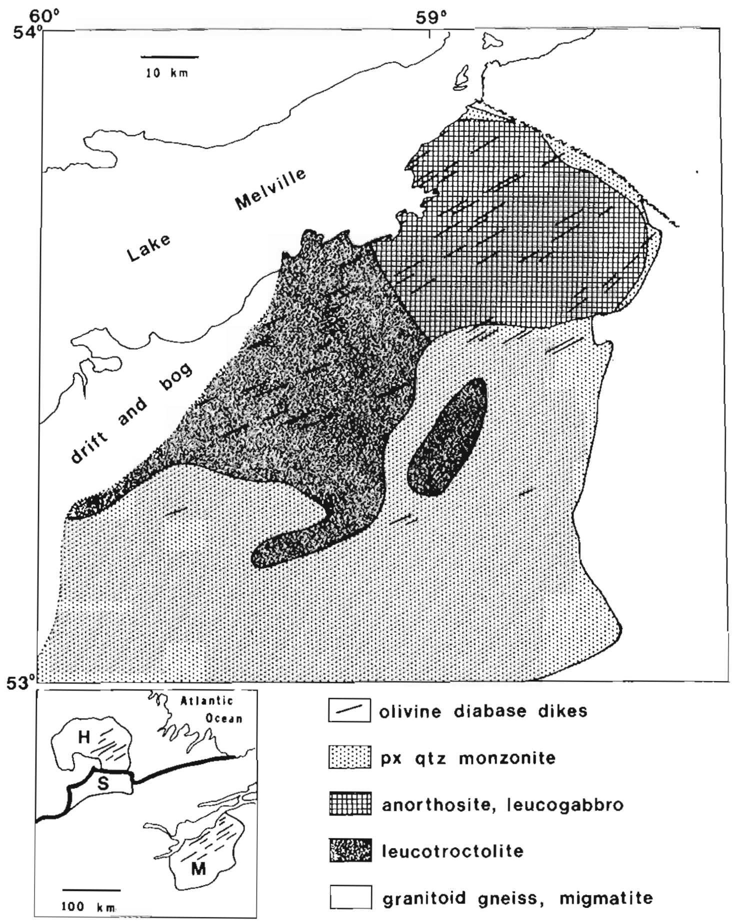


Figure 33. 1. Geological sketch map of the Mealy Mountains complex showing distribution of the major rock units. Small-scale map at lower left shows positions of the Harp Lake Complex (H), the Seal Lake synclinorium (S), and the Mealy Mountains complex (M); olivine diabase dyke swarms are indicated and the approximate position of the Grenville Front is a heavy black line.

Major units of the Complex

Like many other large anorthositic plutons, the Mealy Mountains complex can be readily subdivided into two major rock groups: 1) older basic rocks consisting of leucotroctolite, leucogabbro and anorthosite; 2) a younger group of variable composition centred around pyroxene quartz monzonite. Further subdivision of the basic group is possible because the areal distribution of leucotroctolite is restricted to the southwestern part of the complex whereas anorthosite and leucogabbro are dominant in the topographically higher northeast part. This is in accord with a greater resistance to erosion of the more feldspathic compositions and has been observed in other complexes. The basic rocks are medium to very coarse grained, with leucotroctolites, in general, being less coarse than anorthosite and leucogabbro. A number of pegmatoid patches 1 to 5 m across with red graphic granite cores were found in anorthosite and leucogabbro (Fig. 33. 8).

The pyroxene quartz monzonite group is variable in composition and texture. Much of the rock is medium grained equigranular with a characteristic greenish lustre on fresh surfaces. Other varieties are pinkish grey to pink and one type, with 2 to 5 cm ovoidal K-feldspar crystals, is a deep reddish colour (Fig. 33. 9). Dykes of pyroxene quartz monzonite cut anorthositic rocks and inclusions of anorthosite occur in pyroxene quartz monzonite. Locally, xenoliths of country rocks are abundant within the pyroxene quartz monzonite.

Internal Structures

Within leucotroctolite, leucogabbro and anorthosite, igneous mineral layering on a scale of 1 cm to 10 m, or more, is common (Fig. 33. 2). Where rock exposure is good, layers or layer contacts can be found on most clean subvertical rock surfaces more than a few metres high. In leucotroctolite the mineral layering involves differences in proportions of olivine and plagioclase, whereas in leucogabbro and anorthosite, variations in pyroxene and plagioclase proportions are the chief cause. In addition to mineral layering, manifestation of layers is due to differing plagioclase grain sizes and shapes and in some cases, plagioclase colour particularly in leucogabbro and anorthosite. Dark layers in some anorthositic rocks were found to consist of darker coloured plagioclase than that in the enclosing host rock. Such "dark plagioclase" layers were evident in some places even where the enclosing rock contained a noticeably higher proportion of ferromagnesian minerals. Large areas of the complex have relatively consistent layering attitudes. For example, the northwestern part of the complex has dips to the northwest and west at angles less than 30 degrees. Other areas have remarkably discordant layer attitudes over short distances and sometimes even within the same outcrop. Such discordance seems to be a primary igneous feature of the crystal accumulation process and is not due to superimposed deformation.

Rare, thin, streaky dark layers occur in places in the pyroxene quartz monzonite but in the field much of the rock appears relatively massive.

Giant Pyroxenes

Orthopyroxene megacrysts with exsolved plagioclase lamellae parallel to (100) occur in many large anorthositic complexes (Emslie, 1975). A sizable collection of such pyroxenes (Fig. 33. 3-33. 7) made from the Mealy Mountains complex displays macroscopic features similar to those found elsewhere. A great range in size is present from 2 cm to more than 1 m. Many but not all of the crystals have rounded or subrounded outlines. The crystals occur singly, in groups of a few or a great many and sometimes are packed closely together in large pods (Fig. 33. 5), the pods themselves having smooth, commonly rounded or ellipsoidal outlines. Examples were found of rounded, plagioclase lamellae-bearing orthopyroxene megacrysts with overgrowths of orthopyroxene free of plagioclase lamellae and intergrown with adjoining plagioclase crystals (Fig. 33. 6). The field evidence is consistent with the megacrysts having been brought up from depth after initially crystallizing under higher pressures.

Olivine diabase dyke swarm

The heavy concentration of olivine diabase dykes intruding the Mealy Mountains complex is of more than passing interest. Their mean strike direction closely parallels that of a swarm of similar dykes in the Harp Lake Complex (Emslie *et al.*, 1972) about 150 km to the northwest, on the other side of the Grenville Front (Fig. 33. 1). Macroscopically the dykes are similar in all respects to those at Harp Lake, but of course correlation on the basis of appearance is hazardous. Further petrographic and chemical investigation will be required to test the correlation. If both swarms are similar in age they may be related to the proposed Seal Lake Rift Zone (Baer *et al.*, 1974) and would then imply that large rotational motion of the northeastern Grenville Province (Fahrig *et al.*, 1974) did not take place following dyke emplacement.

Metamorphism and deformation of the Complex

Macroscopic effects of a regional metamorphic overprint on the Mealy Mountains complex are remarkably small. Leucotroctolites do not show evidence of reaction between olivine and plagioclase nor is significant recrystallization visible in most rocks of the basic group. In fact, the bulk of the basic group rocks are indistinguishable in the field from similar rocks of the Michikamau, Harp Lake and Nain complexes north of the Grenville Front.

The pryoxene quartz monzonites, however, do show some evidence of metamorphic reaction and recrystallization. Small amounts of garnet occur in places, usually associated with clusters of mafic minerals. The textures of the rocks as seen on clean weathered surfaces are slightly blurred, rather than pristine igneous textures, and presumably result from incipient recrystallization.

Deformation of the complex also appears to be relatively minor. A strong east-northeast joint set, in part occupied by the olivine diabase dyke swarm,

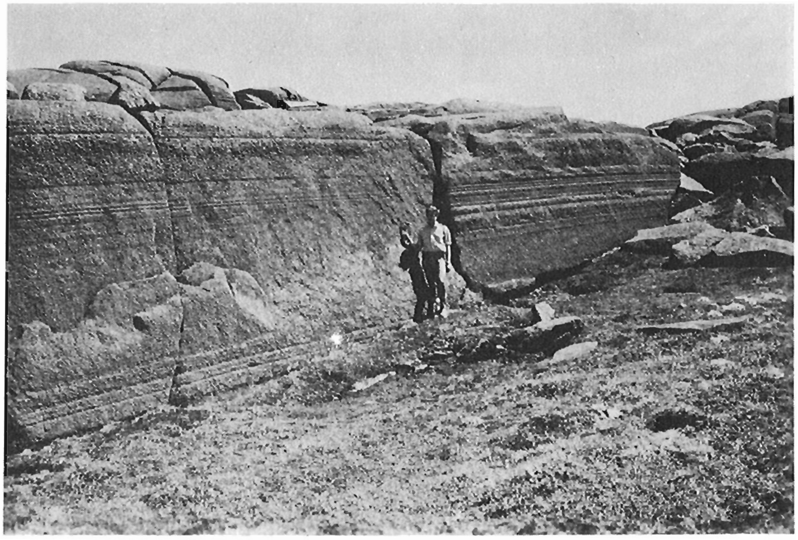


Figure 33. 2.

Unusually well- developed mineral and textural layers in anorthosite. GSC 202928.

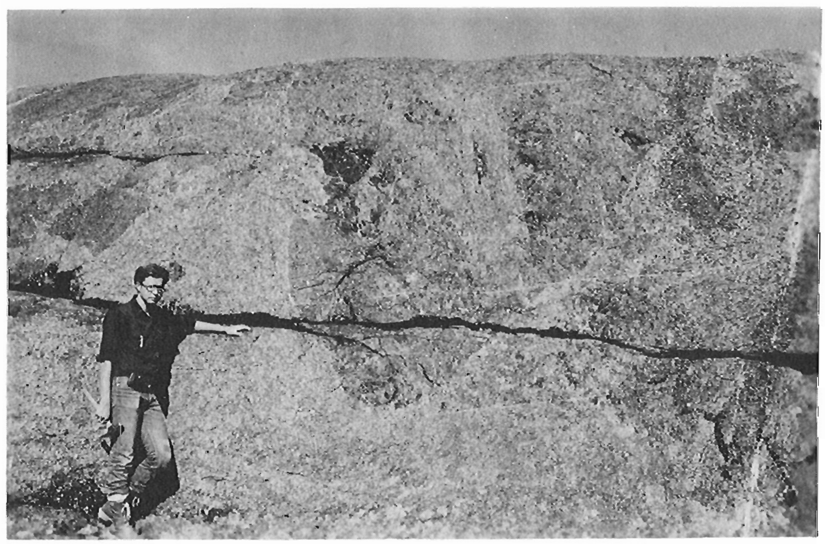


Figure 33. 3

Older, very coarse grained anorthosite (medium grey) invaded by younger, medium grained anorthosite (light grey). Dark mass in upper middle is an orthopyroxene megacryst within the very coarse grained rock. GSC 202928-A.

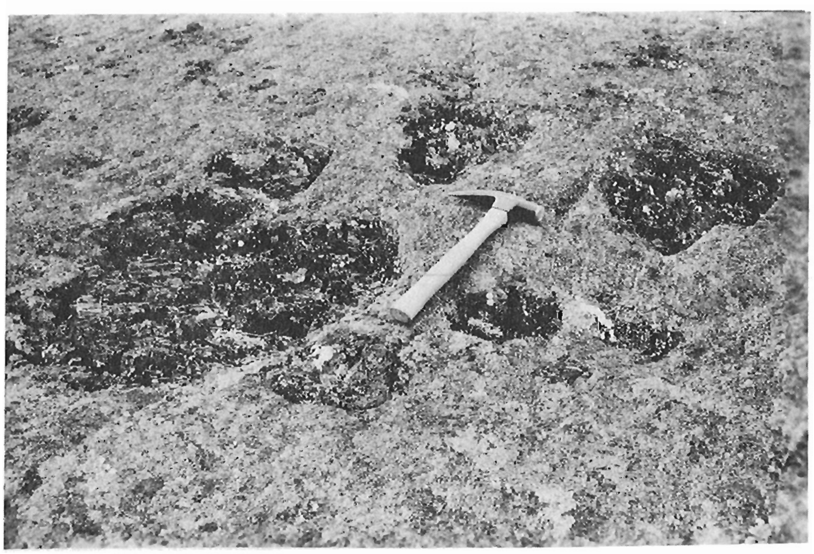


Figure 33. 4.

Group of subrounded orthopyroxene megacrysts in anorthosite. GSC 202928-B.

Figure 33. 5.

Closely packed, subrounded orthopyroxene megacrysts within a large pod. GSC 202928-E.

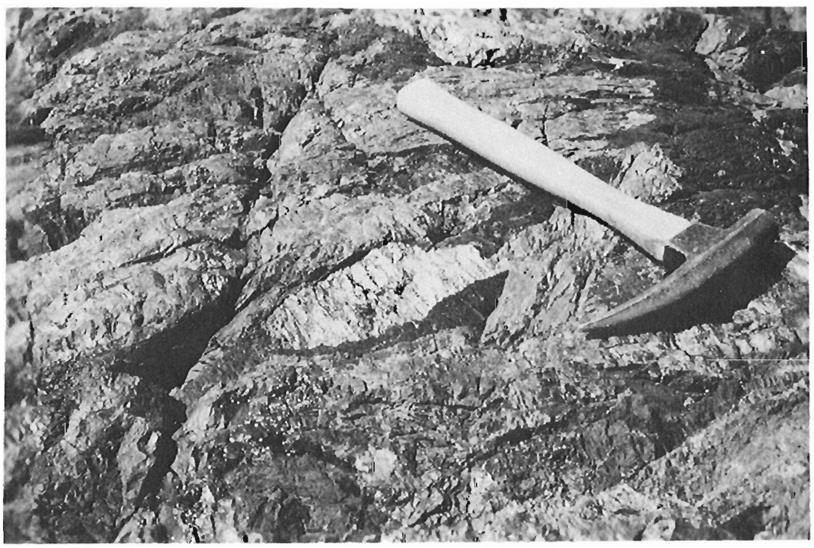
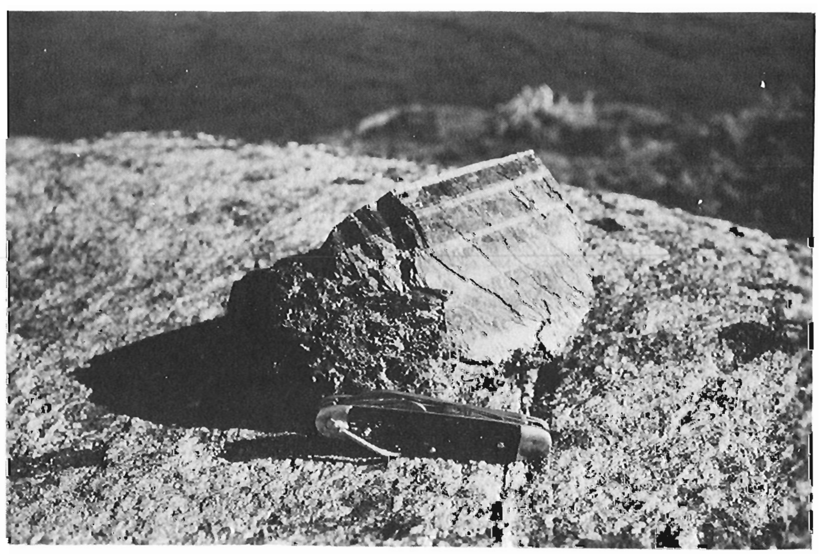


Figure 33. 6

Subrounded orthopyroxene megacrysts (light) surrounded by overgrowth of secondary ortho- pyroxene (dark) intergrown with adjacent plagioclase crystals. In very coarse leucogabbro. GSC 202928-C.

Figure 33. 7.

Orthopyroxene megacryst showing kink- banded cleavage. Kink-banded is a common feature in these megacrysts and the enclosir rocks are not deformed. GSC 202928-D.



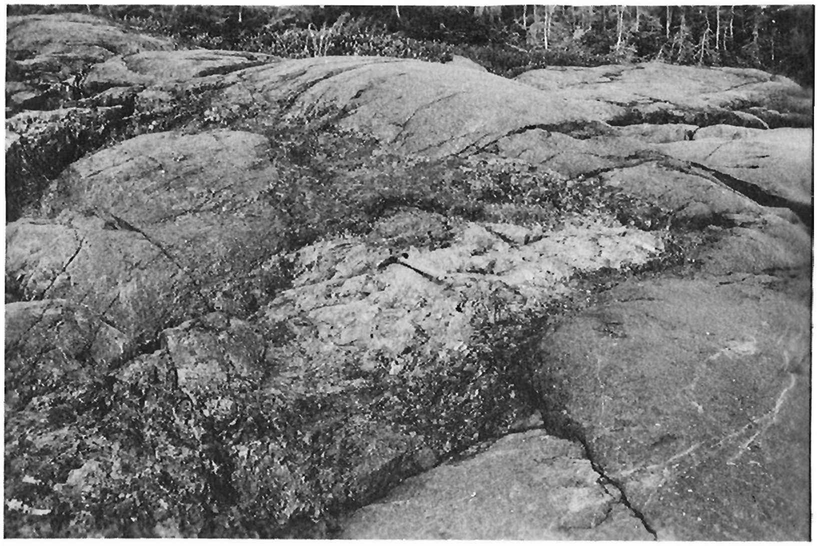
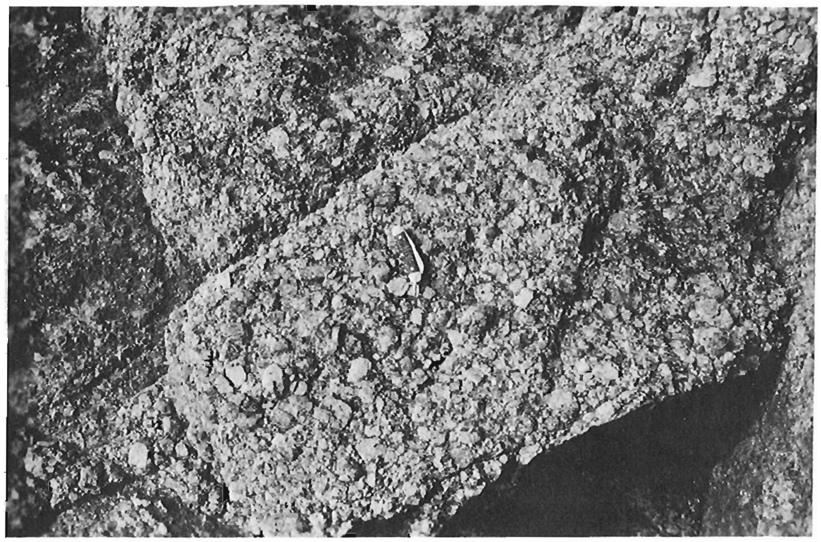


Figure 33.8.

Pegmatoidal mass in leucogabbro. The central area (light) is brick-red graphic granite which is surrounded by coarse pyroxene and oxide-rich material. GSC 202928-F.

Figure 33.9

Reddish, coarse, ovoidal K-feldspar rock of the pyroxene quartz monzonite group. GSC 202928-G.



dominates topographic features within the complex. Some shearing is present parallel with this direction and, at least in part, it postdates emplacement of the diabase because some of the larger dykes have locally sheared margins.

A major fault has been assumed by previous workers to bound the northeastern margin of the complex. A strong topographic lineament is certainly present but rock exposure is poor along it and unequivocal evidence of a major dislocation was not established.

Economic Geology

Some concentrations of Fe-Ti oxide minerals are known to occur in the basic rocks of the Mealy Mountains complex through exploration work chiefly by the Newfoundland and Labrador Corporation during the middle 1950's. Significant tonnages have not been proven and titaniferous magnetite seems to be more common than ilmenite. A few small pods of massive magnetic oxide in leucogabbro and anorthosite were discovered during the field season. Several gossan areas, some nearly 100 m across, contain small amounts of disseminated pyrrhotite and chalcopyrite. The largest gossan observed is on a ridge about 7.5 km due east of the mouth of Etagaulet River.

References

Baer, A.J., Emslie, R.F., Irving, E., and Tanner, J.G.
1974: Grenville geology and plate tectonics; Geosc., v. 1, p. 54-61.

Eade, K.E.
1962: Geology of Battle Harbour-Cartwright area, Coast of Labrador, Newfoundland; Geol. Surv. Can., Map 22-1962.

Emslie, R. F.
1975: Pyroxene megacrysts from anorthositic rocks: new clues to the sources and evolution of the parent magmas; Can. Mineral., v. 13, p. 138-145.

Emslie, R. F., Morse, S. A., and Wheeler, E. P., 2nd.
1972: Igneous rocks of central Labrador, with emphasis on anorthositic and related intrusions; XXIV Int. Geol. Cong. field excursion guidebook A54, 72 p.

Fahrig, W. F., Christie, K. W., and Schwarz, E. J.
1974: Paleomagnetism of the Mealy Mountain anorthosite suite and of the Shabogamo gabbro, Labrador, Canada; Can. J. Earth Sci., v. 11, p. 18-29.

Gillett, L. B.
1956: Anorthosites and syenites of the Mealy Mountains area, Labrador; Unpubl. M.Sc. thesis, McGill Univ.

Gittins, J.
1972: A note on the age of basalt dikes in the Mealy Mountains, Labrador, Canada; Can. J. Earth Sci., v. 9, p. 1337-1338.

PETROLOGICAL STUDIES IN THE SHERRIDON AREA, MANITOBA

Project 740024

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The geology of the Sherridon area combines several features which make detailed studies worthwhile. A lithologic diversity of rocks permits stratigraphic mapping and the presence of marker horizons facilitates structural investigations. In spite of the high grade of metamorphism, conformable copper-zinc orebodies have been preserved. Petrological studies in the area will be directed towards an evaluation of metamorphic conditions, an understanding of the stability of sulphide-silicate assemblages, and a recognition, as far as possible, of the primary rocks and associated ores. As a basis for petrological studies, an area 6 by 8 miles between Kississing Lake and Star Lake will be mapped on a scale of 1: 25 000.

The high-grade gneisses of the area comprise a great variety of rock types, derived from a wide range of sedimentary and volcanic rocks. Conformable layers of sulphides, extending over remarkably long distances, are characteristic and give rise to several copper-zinc orebodies. The geological setting is conveniently seen on a compilation map by Bailes (1971). The area has been mapped on a scale of 1 inch to 1 mile (Bateman and Harrison, 1946; Robertson, 1953). Following the work of Robertson (op. cit.), a two-fold division of the gneisses became accepted. The older Nokomis Group consists of gneisses and megmatites derived from greywackes; the younger Sherridon Group consists of metamorphosed arkoses and calcareous rocks with some interlayered volcanic rocks. A review of the regional geology by Harrison (1951) includes all references to published studies with the exception of reports by Bruce (1929, 1933) and a later structural study reported in the Northern Miner (1966). Figure 34.1 shows a sketch map based on previous work and mapping during 1975.

Sulphides in varying concentration occur along three horizons within the Sherridon Group:

1. A zone in quartzitic gneisses adjacent to the diorite forming the boundary between the Sherridon Group and Nokomis Group, extending from Camp Lake through Sherlett Lake to Molly Lake.
2. A zone including the Sherritt-Gordon orebodies and probably the Jungle Lake deposit. A pit south of Singsing Lake carries disseminated chalcopyrite in anthophyllite rock and probably also belongs to this horizon.
3. A zone including the Bob Lake deposit, extending from Bob Lake to the east end of Cree Lake and Found Lake and north again to Creek Lake. Drilling between Bob Lake and Jonah Lake has indicated some interesting mineralization, probably a continuation of the Bob Lake deposit.

Mapping during 1975 supported, in general, the validity of the previously established stratigraphy and structure. Lithologic units define an approximately concentric structure. A prominent foliation is subparallel to lithologic contacts. At least one earlier deformation is reflected by small isoclinal folds with axial planes parallel to the main foliation. Semipelitic rocks near the centre of the Sherridon structure, previously interpreted as domed-up Nokomis rocks, are actually interlayered with quartzitic gneisses typical of the Sherridon Group. Several sporadic occurrences of anthophyllite rocks were noted. In some instances, e.g. at the east orebody of the Sherritt-Gordon, they appear to be associated with mineralization.

References

Bailes, A.H.
1971: Preliminary compilation of the geology of the Snow Lake – Flin Flon – Sherridon area; Man. Mines Br., Geol. Paper No. 1/71.

Bateman, J.D. and Harrison, J.M.
1946: Sherridon, Manitoba; Geol. Surv. Can., Map 862A.

Bruce, E.L.
1929: The Sherritt Gordon copper-zinc deposit, northern Manitoba; Econ. Geol, v. 24, p. 457-469.

1933: Mineral deposits of the Canadian Shield; The MacMillan Company, Toronto.

Harrison, J.M.
1951: Precambrian correlation and nomenclature, and problems of the Kisseynew gneisses, in Manitoba; Geol. Surv. Can., Bull. 20.

Northern Miner, 3 and 24 February 1966.

Robertson, D.S.
1953: Batty Lake map-area, Manitoba; Geol. Surv. Can., Mem. 271.

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Figure 34.1. Geological sketch map of the Sherridon area, Manitoba.

35. VOLCANIC ROCKS OF THE APPALACHIAN PROVINCE: ROBERTS ARM GROUP, NEWFOUNDLAND (2 E and 12 H)

Project 730043

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Mapping of the Roberts Arm Group, Notre Dame, Newfoundland, at a scale of 1: 25 000 was continued to include Pilley's Island and an area of predominantly volcanic rocks south and east of Crescent Lake. Samples of the volcanic rocks were collected for preliminary petrochemical studies, and a section through the Roberts Arm Group consisting of 26 sites was drilled in collaboration with E. J. Schwarz of the Paleomagnetic Section for paleomagnetic studies.

A brief description of the lithologies examined in the Roberts Arm area has been given in Bostock (1974). The following notes amplify that report.

Roberts Arm Group

The Roberts Arm Group consists mostly of green to grey, massive to porphyritic, in large part abundantly amygdular, basaltic to andesitic lavas that are commonly pillowed. Volcanic breccias are widely present as lenses, and sedimentary beds including chert, siltstone and greywacke, mostly from a few centimetres to a few metres thick, are intercalated locally. Felsite, either massive or porphyritic, is concentrated within the northern and western parts of the group.

A columnar basalt sill showing 22 or more faintly defined layers, each a few centimetres thick is exposed on the shore at Burnt Head. Other layered bodies of basalt, up to about 10 metres thick, are also probably sills. Several small gabbro bodies, some with an elongate lenticular form that parallel the regional strike, are also probably shallow intrusives.

Greywacke, possibly of waterlain pyroclastic origin, is composed of grains of mafic volcanic material, glass, chert and felsite in a quartz-bearing matrix and occurs in a poorly exposed body of mappable size southwest of Boot Harbour Second Pond. Commonly the rock is massive but in places it occurs in beds 10 cm or more thick that are characterized by differing grain size. Layers of similar greywacke interfinger laterally with the neighbouring basalts, and locally show a decrease in grain size away from the main body.

Small bodies of polymict breccia a few metres thick composed of fragments of felsite commonly 5 to 6 cm in diameter, red chert, and more mafic volcanic rocks in a tough, greenish volcanic matrix, are interleaved with, and locally intruded into basalt on the prominent hill 1½ km northwest of Boot Harbour Pond. Some fragments in this breccia contain sparsely disseminated pyrite.

Basalts southeast of Crescent Lake that may form part of the Roberts Arm Group are commonly slightly foliated and in places more altered than those to the northwest. Near Kippins Pond green siltstone and breccia become increasingly abundant within the southeastern belt of basalt, and it is possible that the basalt passes by facies change southwestward into tuffs, breccia and siltstone.

Small bodies of brownish felsite and felsite-like rock, present southeast of Crescent Lake, are mostly aphanitic. The largest of these, a lens of quartz-plagioclase porphyry about 1 km long, is probably at least 100 metres thick. Green aphanitic to quartz-phyric felsite layers are present between siltstone and basalt on the south shore of Tommys Arm. These rocks project into the hills to the southwest where they terminate abruptly, apparently along a fault. Small pyrite gossans are present along the northwest margin of these felsites.

Roberts Arm – Exploits Group contact

Southeast of Crescent Lake basalts similar to those of the Roberts Arm Group are interbanded on a large scale with siltstone, slate and chert that include a few lenses of greywacke. The sediments in contrast to the volcanic rocks, are mostly severely deformed, but in the extreme east and southeast part of the area mapped, deformation is again less severe. There, where a few lenses of conglomerate and minor carbonate beds are intercalated and greywacke is perhaps more common, the rocks have previously been included in the Exploits Group (Williams, 1964).

Locally pillows and minor scour and fill structures within and near the volcanic belts suggest that the basic volcanic rocks overlie the sediments, and in places interbedding of sediments and basalt is evident suggesting that some contacts are conformable. It is possible therefore that a predominantly sedimentary rock sequence (the Exploits Group) in the southeast part of the area mapped, conformably underlies a predominantly volcanic sequence (the Roberts Arm Group) to the north and west, and that the contact zone has been folded. For the sake of simplicity of presentation this interpretation is accepted here. However at least one mappable band of siltstone and chert, like those in the area southeast of Crescent Lake, is present well within the Roberts Arm Group to the west of the lake. Furthermore, several faults penetrate the boundary zone between the two groups and parts of the sedimentary section may be more closely related to the Roberts Arm Group than to the Exploits Group. Conversely, the basalts south and east of Crescent Lake may not belong to the Roberts Arm Group.

Springdale Formation

The Springdale Formation within the area about Roberts Arm occurs on Sunday Cove and Pilley's islands where it lies along the north margin of the Roberts Arm

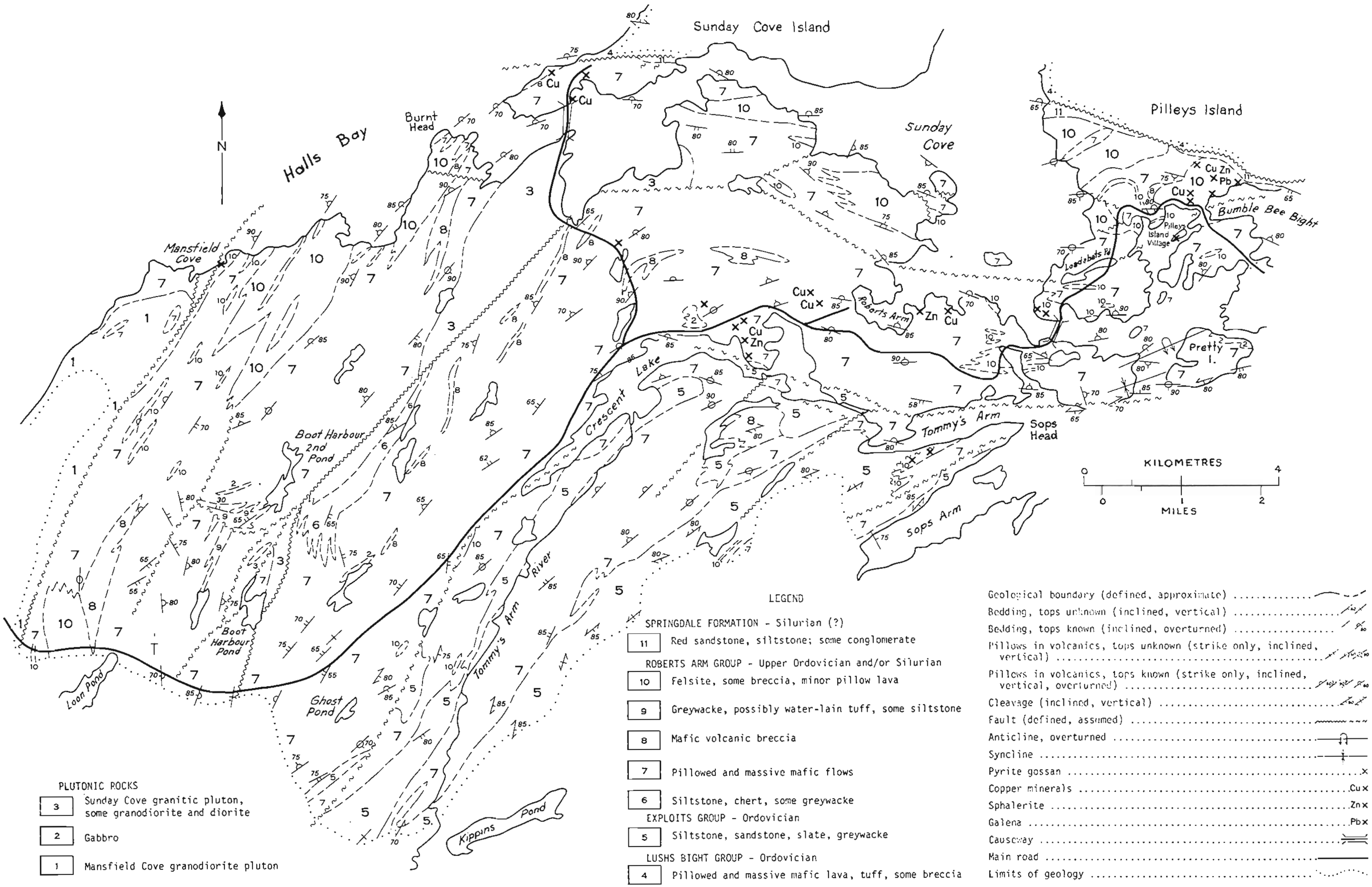
Sunday Cove Island
Pilleys Island
Holls Bay
Burnt Head
Mansfield Cove
Sunday Cove
Bumble Bee Bight
Pilleys Island Village
Roberts Arm
Pretty I.
Crescent Lake
Tommy's Arm
Sops Head
Sops Arm
Boot Harbour 2nd Pond
Boot Harbour Pond
Ghost Pond
Tommy's Arm River
Kippins Pond
Loon Pond
N
KILOMETRES
MILES
LEGEND
SPRINGDALE FORMATION - Silurian (?)
11 Red sandstone, siltstone; some conglomerate
ROBERTS ARM GROUP - Upper Ordovician and/or Silurian
10 Felsite, some breccia, minor pillow lava
9 Greywacke, possibly water-lain tuff, some siltstone
8 Mafic volcanic breccia
7 Pillowed and massive mafic flows
6 Siltstone, chert, some greywacke
EXPLOITS GROUP - Ordovician
5 Siltstone, sandstone, slate, greywacke
LUSHS BIGHT GROUP - Ordovician
4 Pillowed and massive mafic lava, tuff, some breccia
PLUTONIC ROCKS
3 Sunday Cove granitic pluton, some granodiorite and diorite
2 Gabbro
1 Mansfield Cove granodiorite pluton
Geological boundary (defined, approximate)
Bedding, tops unknown (inclined, vertical)
Bedding, tops known (inclined, overturned)
Pillows in volcanics, tops unknown (strike only, inclined, vertical)
Pillows in volcanics, tops known (strike only, inclined, vertical, overturned)
Cleavage (inclined, vertical)
Fault (defined, assumed)
Anticline, overturned
Syncline
Pyrite gossan
Copper minerals
Sphalerite
Galena
Causeway
Main road
Limits of geology

Figure 35. 1

Group and is in fault contact with volcanic rocks of the Lushs Bight Group to the north. Recent exposures along a road north of Pilley's Island village support the contention of Espenshade (1937) that the Roberts Arm – Springdale contact is an unconformity. Basalt and weathered basalt of the Roberts Arm Group on the south pass northwards into several feet of talus breccia composed of angular fragments of weathered basalt up to about 10 cm in diameter in a sandy, iron-bearing, carbonate-rich matrix. This breccia is succeeded by an interval of some 3 m or more in which beds of smaller basalt fragments are intercalated in red siltstone. These beds in turn are overlain by thin bedded, and locally cross-bedded, red siltstone typical of the Springdale Formation. Elsewhere, on the shore east of Bumble Bee Bight, basal conglomerate of the Springdale Formation containing boulders in places reaching 1 m in diameter of felsite, chert and sandstone overlies altered Roberts Arm basalt without evidence of faulting at the contact.

Structure

The Roberts Arm Group west and northwest of Crescent Lake is considered to form two major, steeply dipping, northwest-facing fault blocks (Bostock, 1974). A third major, steeply dipping, southwest-facing block is inserted between the former two to the southwest of a prominent flexure in the dividing fault. Southwestward splaying of this fault has produced a series of subsidiary fault wedges within which granitic to dioritic rocks are exposed and some of their accompanying basic volcanic rocks metamorphosed.

The Roberts Arm Group to the north of Crescent Lake is likely divided into two major fault blocks, comparable to those farther west, along a fault passing through Loadabats Pond and possibly along the lineament west of Bear Cove. The northern fault block thus defined pinches out at the east edge of the area mapped where its southern fault boundary merges with the Lobster Cove fault.

Southeast of Crescent Lake but northwest of Sops Arm and Kippins Pond the rocks lie in a zone of more intense deformation which includes the boundary between the Exploits and Roberts Arm groups. Most but not all observed indicators of stratigraphic sequence in this zone face northwest and it is inferred that the rocks have been both folded and faulted. Minor folds in sediments within this zone commonly appear to fold an earlier foliation and are steeply plunging possibly reflecting transcurrent movement on faults that penetrate the zone.

East of Crescent Lake the deformed zone narrows and is restricted approximately between Sops Head and the north shore of Tommys Arm. Farther east along the north shore of Tommys Arm fractured to schistose basic volcanic rocks are plastered against the coastal hillside locally truncating at high angles the trend of pillow lavas immediately inland to the north. Both deformed and less deformed rocks are cut by joints and fractures showing gently south-plunging slickensides.

To the northeast of Sops Head, and north of the zone of strongest deformation, pillow lavas of the Roberts Arm Group are folded in an east-northeasterly trending belt that is truncated on the west by the zone of more severe deformation. Major folds in this belt are closed and upright to northerly overturned. The belt of folding probably passes through Pretty Island but does not appear to affect the southern coast of Pilley's Island where pillow lavas are all north-facing.

North of the fault through Loadabats Pond structural interpretation is difficult because of the predominance of massive felsite. Pillow lavas in the bluffs above Pilley's Island village and farther west face north but those farther north face to the south suggesting that the felsite in the vicinity of the old Pilley's Island mine occupies the core of a syncline, or is separated from the basalts along one contact by a fault.

Economic Geology

Although numerous minor occurrences of pyrite were observed during the current work no new occurrences of particular interest were found. The largest showing consists of a series of small gossans as large as 3 m^2 comprising disseminated pyrite in felsite south of Tommys Arm. Several minor bodies of felsite and of felsite-like rocks with associated minor pyrite were discovered in the region south and east of Crescent Lake but further effort is required to define the nature of these rocks and to evaluate their economic significance. Mineral prospects in the vicinity of Ghost Pond, said to contain copper and zinc mineralization, have not yet been examined.

References

Bostock, H.H.
1974: Volcanic rocks of the Appalachian Province: Roberts Arm Group, Newfoundland; *in* Report of Activities, Part A, Geol. Surv. Can., Paper 75-1A, p. 1-3.

Espenshade, G.H.
1937: Geology and mineral deposits of the Pilley's Island area; Nfld. Dep. Nat. Resour., Bull. 6.

Williams, H.
1964: Botwood; Geol. Surv. Can., Map 60-1963.

36. METALLIC MINERALIZATION OF CARBONIFEROUS TO TRIASSIC STRATA IN THE APPALACHIAN REGION

Project 740110

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Carboniferous to Triassic strata underlie a northeasterly trending belt in the Atlantic Provinces that is roughly 800 km long and which widens to approximately 300 km near the southwestern end. The composite thickness has been estimated to exceed 10 000 m. The lithostratigraphy has been subdivided into seven groups: Horton, Windsor, Canso, Riversdale, Cumberland, Pictou and Fundy groups. Marine conditions existed during Windsor time, and the lithology of all other groups reflects fluviatile or lacustrine conditions.

Metallic mineralization occurs at various stratigraphic horizons as stratiform and stratabound deposits (Wright, 1972). Such deposits are known to be often closely associated with regional unconformities. The documentation of such unconformities is being given a high priority of the initial phase of the project. Widespread copper mineralization at the base of the Windsor (Binney, 1975), important lead-zinc occurrences near the top of the Lower Windsor (MacEachern and Hannon, 1974) and copper occurrences in the basal Riversdale (Wright, 1950) are some of the known occurrences which fall within the group of unconformity-related mineralization.

Detailed investigations of the sedimentology, related tectonic movements and the diagenesis of mineralized strata are to be undertaken starting in 1976 in order to evaluate the mineral potential of these strata on a regional scale. A survey of industry, provincial mines departments and universities in 1975 has shown an increasing interest in the Carboniferous-Permian of the Atlantic Provinces.

References

Binney, W. P.
1975: Copper occurrences in Lower Carboniferous sedimentary rocks of the Maritime Provinces; Geol. Surv. Can., Open File 281, 156 p.

MacEachern, S. B. and Hannon, P.
1974: The Gays River discovery — a Mississippi Valley type lead-zinc deposit in Nova Scotia; Can. Inst. Min. Met. Bull., v. 67, no. 750, p. 61-66.

Wright, J. D.
1972: Economic geology, *in* Stratigraphy and economic geology of Carboniferous basins in the Maritime Provinces, H. W. Van de Poll; 24th Int. Geol. Congr., Canada, Field Excursion A60 guidebook, p. 33-48.

Wright, W. J.
1950: Dorchester copper deposits, Westmorland County, N. B.; Mines Br., Paper 50-9.

Project 740023

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Geological mapping of the Trepassey map-area commenced in 1974 (Williams and King, 1975) and is now complete. The results are summarized in Figure 37. 1.

The Avalon Peninsula is part of the much larger Avalon Zone in Newfoundland (Williams *et al.*, 1974) that consists of late Precambrian volcanic and sedimentary rocks locally overlain by Cambrian and Ordovician strata. Correlatives occur southwestward along strike at Cape Breton Island of Nova Scotia, southeast New Brunswick, and eastern Massachusetts. Still more southerly correlatives occur in Virginia and the Carolinas, making the Avalon Zone one of the most continuous and best defined in the Appalachian Orogen. The tectonic significance of this zone in any plate tectonic model for the Appalachian System is at present poorly understood.

All rocks in the map-area are of late Precambrian (Hadrynian) age. The oldest comprise a bimodal volcanic assemblage known as the Harbour Main Group (Rose, 1952). These are overlain by three assemblages of sedimentary rocks, which in order of decreasing age are the Conception, St. John's, and Signal Hill groups (Rose, 1952), revised from the 1974 work (Williams and King, 1975).

The Conception Group is dominated by green siliceous rocks, except toward the southwest where it is composed chiefly of thick-bedded (1- to 6-foot) coarse sandstones. It is dividied into five formations. From bottom to top these are the Mall Bay, Gaskiers, Drook, Briscal, and Mistaken Point formations. The Gaskiers Formation is a thick Precambrian tillite and the Mistaken Point is profusely fossiliferous toward its top. In the east, the Drook Formation of the Conception Group overlies the Harbour Main volcanics with structural conformity and a local basal conglomerate. In the west, it is underlain by the Gaskiers and Mall Bay formations with the base of the sequence unexposed. The situation suggests either eastward overstep of the Drook Formation onto the Harbour Main Formation, or else that the Gaskiers and Mall Bay formations in the west are time equivalents to part of the Harbour Main volcanic sequence in the east. The Briscal Formation and the Cape English and Peter's River members of the Drook Formation all appear to be southwesterly derived coarse clastic facies of the siliceous Drook beds exposed farther north. The easily recognizable red sandstones and slates of the Mistaken Point Formation are chosen to define the top of the Conception Group. Its fossils also define a biostratigraphic zone, and a concentration of thin tuff horizons in the formation indicates that it is a chronologic unit as well. The fossiliferous Mistaken Point Formation has been traced 10 miles north of the map-area to Cape Broyle. The Hibbs Hole Formation (Hutchinson, 1953) at Carbonear of Conception Bay appears to be a direct correlative.

The St. John's Group, elevated from St. John's Formation (Rose, 1952), is a dominantly grey to dark grey sandstone-shale sequence that is conformable and gradational above the Conception Group. It is divided into the Trepassey, Fermeuse and Renews Head formations that are all separated by conformable gradational boundaries. These lithic divisions are all recognizable at the type area of the previous St. John's Formation 40 miles northeast of the map-area at St. John's city. Elevation of the St. John's Formation to group status therefore seems valid throughout the eastern Avalon Peninsula.

The Signal Hill Group, elevated from Signal Hill Formation (Rose, 1952), is a dominantly grey to red sandstone sequence that overlies the St. John's Group. It is divided into four new formations, namely the Cappahayden, Gibbett Hill, Ferryland Head, and Cape Ballard formations. Elevation of the former formation to group rank is in accord with imminent nomenclature changes for similar rocks in the type area near St. John's, where the Signal Hill Formation (Rose, 1952) is now known to embrace three distinct mappable units, all of formational status. Because of these proposed changes in nomenclature, the name Cabot Group (Rose, 1952) that was earlier introduced to include the St. John's and Signal Hill formations, is now dropped. The base of the Signal Hill Group is drawn at the Renews Head-Cappahayden contact. At most places there is local erosional disconformity so that the contact is abrupt and easily defined in the field. All other formations within the Signal Hill Group are characterized by conformable gradational boundaries.

A small fault-bounded body of granite in the northeast part of the map-area is similar to granite at Holyrood of Conception Bay (McCartney, 1967). Dykes and stocks of Whalesback Gabbro are all localized in the Drook Formation of the Conception Group.

The late Precambrian rocks of the map-area are folded about southwest-trending axes that plunge gently southwest. The folds are open with steep axial surfaces and with regional axial planar cleavage that is vertical to steeply southeast- or northwest-dipping. Major faults parallel fold axes. Regional cleavage is co-planar with the axial surfaces of major folds, except north of Biscay Bay where prominent folds in the Drook Formation appear to be earlier than the cleavage. Possibly these folds were precursors to later structures and formed by early warping and décollement detachment above a resistant substrate, as might be provided by the Harbour Main volcanics.

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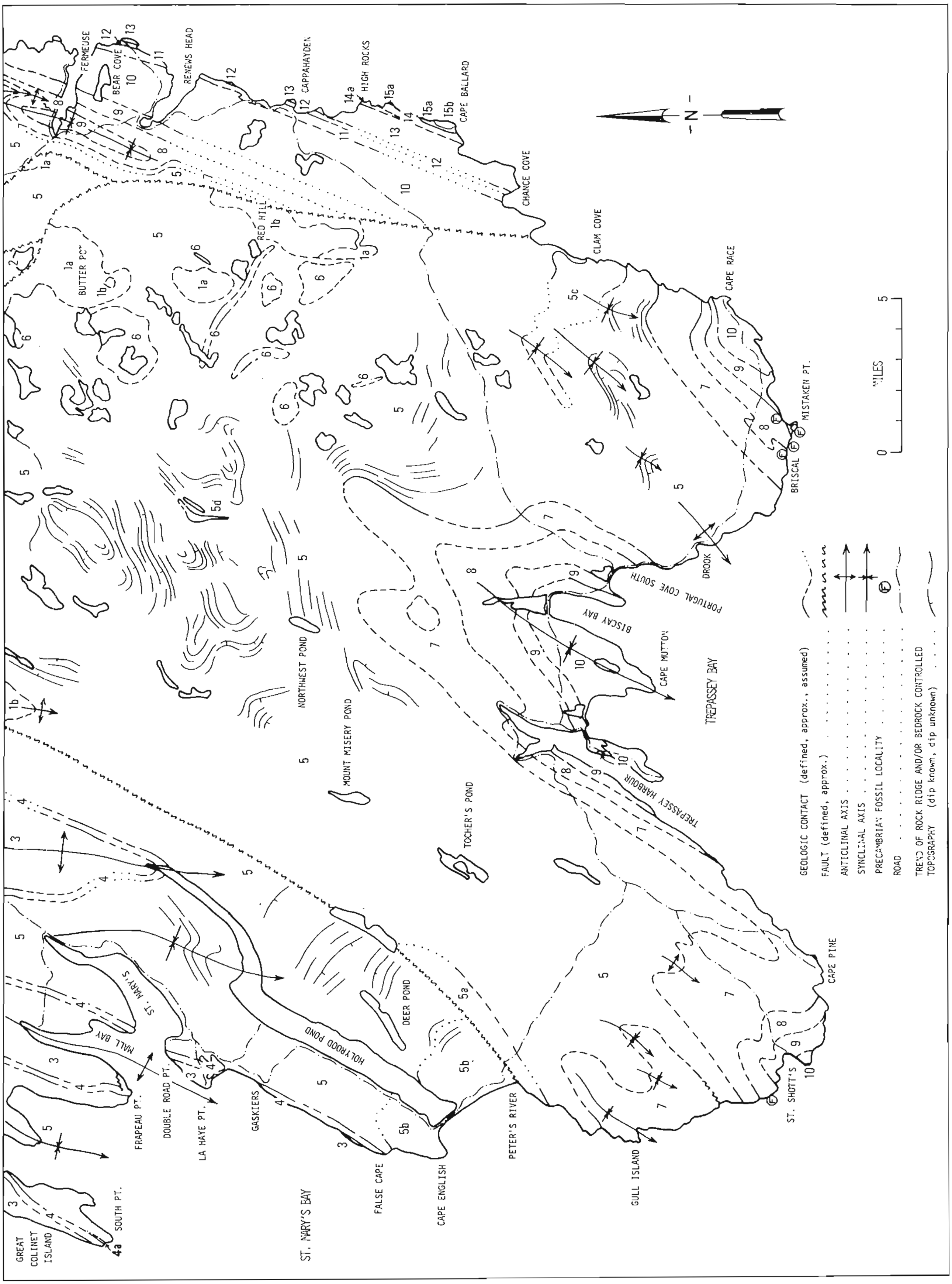
FERMEUSE
BEAR COVE
RENEWS HEAD
CAPPAHAYDEN
HIGH ROCKS
CAPE BALLARD
CHANCE COVE
CLAM COVE
CAPE RACE
MISTAKEN PT.
BRISCAL
DROOK
PORTUGAL COVE SOUTH
BISCAY BAY
CAPE MUTTON
TREPASSEY BAY
TREPASSEY HARBOUR
CAPE PINE
ST. SHOTT'S
GULL ISLAND
PETER'S RIVER
CAPE ENGLISH
FALSE CAPE
ST. MARY'S BAY
GREAT COLINET ISLAND
SOUTH PT.
FRAPEAU PT.
DOUBLE ROAD PT.
LA HAYE PT.
GASKIERS
MALL BAY
ST. MARY'S
HOLYROOD POND
DEER POND
TOCHER'S POND
MOUNT MISERY POND
NORTHWEST POND
RED HILL
BUTTER POT
N
MILES
0
5
GEOLOGIC CONTACT (defined, approx., assumed)
FAULT (defined, approx.)
ANTICLINAL AXIS
SYNCLINAL AXIS
PRECAMBRIAN FOSSIL LOCALITY
ROAD
TREND OF ROCK RIDGE AND/OR BEDROCK CONTROLLED TOPOGRAPHY (dip known, dip unknown)

Figure 37. 1

PRECAMBRIAN (HADRYNIAN)

SIGNAL HILL GROUP (12–15)

Cape Ballard Formation

15 15a, grey shale and siltstone with minor purple shale; 15b, thick-bedded grey sandstone and quartz granule conglomerate

Ferryland Head Formation

14 Thin- to medium-bedded grey sandstone; 14a, High Rocks Member, red wavy-bedded sandstone and shale

Gibbett Hill Formation

13 Thick-bedded light grey sandstone with alternating units of thin bedded dark grey sandstone and siltstone

Cappahayden Formation

12 Fissile parallel-laminated light grey siltstone

ST. JOHN'S GROUP (9–11)

Renews Head Formation

11 Irregular thin-bedded dark grey sandstone with minor shale

Fermeuse Formation

10 Grey to dark grey shale with thin lenses of buff-weathering sandstone and siltstone. Mainly light grey thin-bedded shale and sandstone toward base. Contains trace fossil *Aspidella terranovica*

Trepassey Formation

9 Medium- to thin-bedded grey sandstone and shale, graded sandstone beds with shale tops

CONCEPTION GROUP (3–8)

Mistaken Point Formation

8 Medium-bedded grey to pink sandstone and green to purple and red shale, minor thin tuff horizons. Contains Precambrian fossils toward the top

Briscal Formation

7 Thick-bedded coarse grey sandstone, thick-bedded olive to grey argillite, coarse pale red sandstone and arkose, local units of thin-bedded grey siltstone and shale

Drook Formation

5 Parallel-bedded olive green to grey and buff argillaceous chert, siliceous siltstone and sandstone, silicified tuff, locally includes thick sandstone beds alternating with shale and siltstone, minor purple argillite; 5a, Peter's River Member, grey thick-bedded coarse sandstone; 5b, Cape English Member, medium- to thick-bedded grey sandstone with thick units of pale red sandstone; 5c, Clam Cove Member, grey pebbly sandstone and shale with lozenge-shaped transposed beds; 5d, Biscay Member, mafic pillow lava

Gaskiers Formation

4 Grey tillite with thin alternating units of graded laminites that contain 'dropstones'. Red tillite overlain by red mudstone at top; 4a, South Point Member, red agglomerate

Mall Bay Formation

3 Green siliceous siltstone and argillite, grey sandstone, chert, thick-bedded tuffaceous sandstone and green siliceous tuff and agglomerate, white quartzose sandstone and minor limestone

HARBOUR MAIN GROUP (1)

1 Poorly bedded mixed volcanic rocks; 1a, pink to grey silicic tuff and agglomerate, pink to red rhyolite and welded tuff; 1b, altered green to purplish massive basalt

INTRUSIVE ROCKS

Whalesback Gabbro

6 Fine- to medium- and coarse-grained massive gabbro

Holyrood Granite

2 Medium-grained massive pink granite

NOTE: New stratigraphic names herein used are provisional and will be described formally in a later report.

North of the map-area, the distribution of Gaskiers tillite indicates major northeast-plunging structures, so that the anticlines near St. Mary's close both to the northeast and southwest.

The geologic history of the map-area suggests mainly terrestrial silicic volcanism during initial stages of development (Harbour Main) that was locally accompanied, but mainly followed by marine deposition of the Conception Group. A proximal volcanic terrane during deposition of the lower Conception Group is indicated by the presence of agglomerate and tuff associated with the Gaskiers tillite and numerous volcanic clasts within the tillite. Pillow basalts, mafic dykes and Whalesback Gabbro stocks within the Drook Formation also indicate continuing igneous activity. The St. John's Group records a shoaling of the Conception marine basin, and then deposition of deltaic terrigenous sandstones of the Signal Hill Group. Local tuff beds in the Mistaken Point and high formations diminish upward in the stratigraphic section and record waning volcanic activity during encroachment of the earlier volcanic centres by thick Precambrian sedimentary deposits. Arkosic beds in the Conception toward the south and southwest (Cape English and Peter's River members of the Drook Formation) imply unroofing of a granitic terrane to the southwest that was possibly associated with Harbour Main volcanism.

Provenance directions and the distribution of coarse clastic facies in the Conception Group indicate transport from the south and southwest in the southern part of the map-area. This is contrasted with transport from the northeast in the northern parts of the map-area and farther north in eastern Avalon Peninsula. The total picture therefore appears to be one of northeast and southwest convergence toward the presently exposed Harbour Main Group. This indicates that the bulk of the Conception must overlie the Harbour Main and that the middle and upper Conception could not have been derived from the presently exposed volcanic core of Avalon Peninsula.

A major aeromagnetic anomaly that crosses the entire Avalon Peninsula and extends offshore to the east (Papezik *et al.*, 1975) has no expression in the surface geology of the map-area. Possibly it relates to a basement feature in rocks beneath the Conception Group.

No economic mineral deposits have ever been mined in the map-area and there are very few, if any, promising prospects. Gold has been reported in late Precambrian sedimentary rocks of the eastern Avalon Peninsula, but there are no known occurrences in the map-area. Gravel deposits suitable for road metal abound throughout the map-area (Henderson, 1972).

References

Henderson, E. P.
1972: Surficial geology of Avalon Peninsula, Newfoundland; Geol. Surv. Can., Mem. 368, 121 p.

Hutchinson, R. D.
1953: Geology of Harbour Grace map-area, Newfoundland; Geol. Surv. Can., Mem. 275, 43 p.

McCartney, W. D.
1967: Whitbourne map-area, Newfoundland; Geol. Surv. Can., Mem. 341, 135 p.

Papezik, V. S., Hodych, J. P., and Goodacre, A. K.
1975: The Avalon magnetic lineament: a possible continuation of the Triassic dyke system of New Brunswick and Nova Scotia; Can. J. Earth Sci., v. 12, p. 332-335.

Rose, E. R.
1952: Torbay map-area, Newfoundland; Geol. Surv. Can., Mem. 265, 64 p.

Williams, Harold, Kennedy, M. J., and Neale, E. R. W.
1974: The northeastward termination of the Appalachian Orogen; *in* Nairn, A. E. M. and Stehli, F. G. (editors), The Ocean Basins and Margins, v. 2, p. 79-123, Plenum Pub. Corp., New York.

Williams, Harold and King, A. F.
1975: Southern Avalon, Newfoundland; *in* Report of Activities, Part A, Trepassey map-area (1 K), Geol. Surv. Can., Paper 75-1A, p. 11-15.

THE DUNNAGE MELANGE, NEWFOUNDLAND (2 E)

E. M. R. Research Agreement No. 1135-D13-4-94/75

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Regional and Economic Geology Division

The Dunnage Mélange (Kay and Eldredge, 1968; Horne, 1969; Kay, 1972) has been delineated in its entirety and its largest volcanic blocks and a variety of intrusions localized in the chaotic terrane are outlined (Fig. 38.1). This recent investigation has brought out several interesting relationships, both within the Dunnage and between the Dunnage and nearby rock groups.

A variety of small intrusions that are localized within the mélange terrane are rare or absent in surrounding country rocks. The most common intrusive rock is quartz-feldspar porphyry (Coaker Porphyry of Kay, 1972) that makes up about one-third of the exposed rocks in the northeast. It clearly cuts black matrix shales of the mélange and it is itself cut by the Loon Bay batholith along the east side of the mélange. Other small intrusions include the Causeway Diorite and Puncheon Syenite (Marshall Kay, pers. comm. 1972), the Grapnel Gabbro (Hibbard, in prep.), and a variety of silicic to mafic dykes.

Biotite from the Causeway Diorite has yielded K/Ar ages of 428 ± 13 and 435 ± 13 m.y. (Marshall Kay, pers. comm. 1972) and a Rb/Sr age of 454 or 480 m.y., dependent upon the choice of Rb/Sr decay constants used in calculating the age (R. F. Cormier, pers. comm., 1972). This rock is distinctive in outcrop for it contains numerous inclusions of gabbro and ultramafic rock in various stages of alteration. Like the Causeway Diorite, local exposures of the widespread Coaker Porphyry have the same mafic inclusions, suggesting a similar source and age for both. The isotopic ages, combined with the absence of intrusions like the Coaker and Causeway in nearby Middle Ordovician and Silurian rocks, suggest that the intrusions are essentially coeval with mélange formation. Intricate contacts between silicic dykes and black shales, with patches of igneous rock isolated in shale or with shale deeply penetrating chilled igneous rock, also suggest intrusion into unconsolidated muds. Local splashy mixtures of mafic volcanic rock and shale indicate contemporaneous mafic volcanism as well.

Toward the west at Burnt Bay, the Dunnage has dykes and large blocks of gabbro, and Coaker Porphyry is absent. Farther west at Thwart Island, similar gabbro occurs as numerous sills in the New Bay Formation (Helwig, 1969) and locally the gabbro constitutes up to 90 per cent of thick stratigraphic sections.

The Dunnage Mélange is bounded by the Reach Fault toward the east and it is bordered by a thin belt of north-facing Caradocian black shales of the Dark Hole Formation (Horne, 1970) along its northwest side on New World Island and southwestward. The Dunnage-Dark Hole contact has been variously interpreted as tectonic (Kay, 1972) and depositional (Horne, 1969). Certainly the contact is fault-modified in some places, but in others it is difficult to define a sharp boundary. Toward the northeast at Cheneyville and nearby, the Dark Hole Formation has a pebble to boulder conglomerate at its base which contains fragments of porphyry that resemble nearby Coaker Porphyry in the Dunnage Mélange. This relationship, combined with the continuity of the narrow Dark Hole horizon along the northwestern side of the mélange, suggests a depositional contact and a pre-Middle Ordovician age for intrusions throughout the Dunnage terrane.

The Dildo Fault (Kay, 1972) that occurs at or near the Dunnage-Dark Hole contact, is not thought to have the same structural significance as the Lukes Arm, Cobbs Arm and Toogood faults to the north at New World Island. The latter faults repeat the Ordovician-Silurian stratigraphic sections and they have north-facing Silurian sediments on their southern sides with Ordovician volcanic rocks toward the north. In contrast, the Dildo Fault has the Dunnage Mélange on its southern side, which is interpreted as older than Dark Hole shales on its northern side. Sansom greywackes and Silurian Goldson conglomerates that overlie the Dark Hole represent gradual sedimentary infilling of a marine trough, so that the stratigraphic section northwest of the Dunnage can be viewed as an upward shoaling sequence above a marine mélange basement.

The Dunnage Mélange overlies and apparently interdigitates with the New Bay Formation of the Exploits Group (Helwig, 1969) at its southwest extremity near Thwart Island and Stanhope. Farther west, the New Bay Formation is overlain by the Lawrence Head Volcanics and then Caradocian black argillites (Lawrence Harbour Shale) like those of the Dark Hole Formation. The Lawrence Head Volcanics and equivalents at New World Island are probably represented by a wide zone along the northwest part of the Dunnage that contains all of the largest mafic volcanic blocks. The relationships suggest that the mélange is, in part, a chaotic facies of the pre-Caradocian New Bay Formation and Lawrence Head Volcanics of the Exploits Group.

East of the Dunnage terrane, greywackes and volcani-clastic rocks like those of the New Bay Formation are sparse or absent, and volcanic rocks, so widely represented toward the west, are equally rare. A thick section of manganiferous cherts at Loon Bay and east of the Dunnage Mélange is interpreted as oceanic crust (Kay, 1975), although this is not supported by the presence of an ophiolite suite beneath the cherts.

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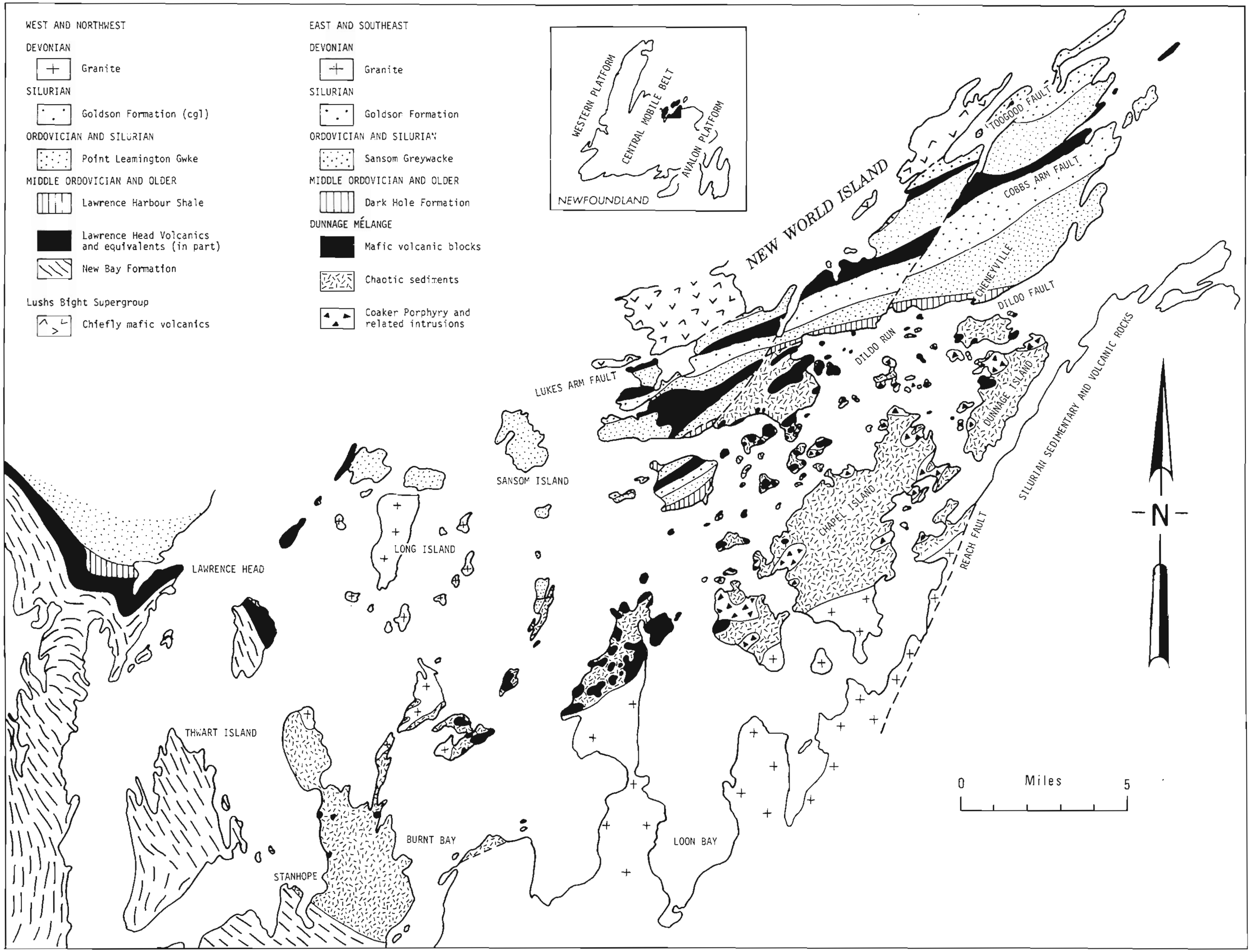

Figure 38.1. Generalized geology of the Bay of Exploits region, northeastern Newfoundland (2 E).

At Carmanville, 20 miles farther east mélange like the Dunnage, but without associated intrusions, reappears at the base of the Middle Ordovician Davidsville Group (Kennedy and McGonigal, 1972). Thus in a general way, the Dunnage and possible eastward equivalents constitute a broad, soft zone that lies to the east and is partly coeval with the well-known volcanic terrane in northeast Newfoundland.

The thickness and age of the Dunnage Mélange are not precisely known. The mélange is from 5 to 8 miles wide across its structural trend and everywhere it is bounded by steeply dipping formations. If the mélange is itself steeply dipping, and not imbricated by thrusts, its thickness is in the order of at least 30 000 feet. The upper parts of the mélange are probably no younger than Caradocian and porphyry fragments at the base of the Dark Hole Formation suggest a pre-Caradocian age for large parts of the mélange terrane that are cut by similar porphyrys. Poorly preserved denroid graptolites in matrix shales near Stanhope suggest a Lower Ordovician age for the rocks there, and it may not be entirely coincidental that the Middle Cambrian limestones and associated volcanics at Dunnage Island (Kay and Eldredge, 1968) are found at the southeast margin of the mélange and toward its presumed base. Conceivably, the mélange developed continuously from Middle Cambrian to Middle Ordovician.

The exact significance of the Dunnage Mélange remains unknown. In many ways the chaotic rocks resemble mélanges of western Newfoundland and black and green matrix shales are almost everywhere the same. However, the west Newfoundland mélanges are thin and they are clearly related to the assembly and transport of allochthonous masses. The Dunnage is much thicker and it is not apparently controlled by a similar tectonic style. Intrusions that are localized throughout the Dunnage and nearby Exploits Group suggest direct linkages and siting above a deep crustal feature, rather than simple surficial control.

Lower formations of the Exploits Group and the dominant mafic volcanic sequences of Notre Dame Bay are interpreted now as typical volcanic arc assemblages (e.g. Kean and Strong, 1975). The Dunnage is therefore sited immediately east of an island arc complex. There, it may represent the vestige of a convergent plate boundary (Dewey, 1969; Horne, 1970; Kay, 1972), either representing a trench deposit, or more likely a mélange ridge (Hibbard and Williams, 1975) situated east of the magmatic arc and west of the actual subduction zone.

References

Dewey, J.F.
1969: Evolution of the Appalachian/Caledonian Orogen; Nature, v. 222, p. 124-129.

Helwig, James
1969: Redefinition of Exploits Group, Lower Paleozoic, northeast Newfoundland; *in* Kay, Marshall (ed.) North Atlantic – Geology and Continental Drift, Am. Assoc. Pet. Geol., Mem. 12, p. 408-413.

Hibbard, J.P.
The southwestern extension of the Dunnage Mélange and its relationships to nearby groups; M.Sc. thesis, Memorial University of Newfoundland. (in prep.)

Hibbard, J.P. and Williams, Harold
1975: The Dunnage Mélange, northeast Newfoundland (abstr.); Geol. Assoc. Can., *in* Geol. Soc. Am. abstracts with programs, v. 7, no. 6, p. 781-782.

Horne, G.S.
1969: Early Ordovician chaotic deposits in the central volcanic belt of northeastern Newfoundland; Geol. Soc. Am. Bull., v. 80, p. 2451-2464.

1970: Complex volcanic-sedimentary patterns in the Magog belt of northeastern Newfoundland; Geol. Soc. Am. Bull., v. 81, p. 1767-1788.

Kay, Marshall
1972: Dunnage Mélange and Lower Paleozoic deformation in northeastern Newfoundland; 24th Intern. Geol. Cong., Sect. 3, p. 122-133.

1975: Campbellton sequence, manganiferous beds adjoining the Dunnage Mélange, northeastern Newfoundland; Geol. Soc. Am. Bull., v. 86, p. 105-108.

Kay, Marshall and Eldredge, N.
1968: Cambrian trilobites in central Newfoundland volcanic belt; Geol. Mag., v. 105, p. 372-377.

Kean, B.F. and Strong, D.F.
1975: Geochemical evolution of an Ordovician Island Arc of the central Newfoundland Appalachians; Am. J. Sci., v. 275, p. 97-118.

Kennedy, M.J. and McGonigal, M.H.
1972: The Gander Lake and Davidsville Groups of northeastern Newfoundland: new data and geotectonic implications; Can. J. Earth Sci., v. 9, p. 452-459.

39. OPERATION SAINT ELLAS, YUKON TERRITORY: PRE-CENOZOIC VOLCANIC ASSEMBLAGES IN THE KLUANE RANGES

Project 730035

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Two and one half months of field work were devoted to a study of the stratigraphy, structure, and economic potential of the volcanic rocks in the Kluane Ranges. Remapping yields a similar distibution of volcanic units to that shown in the Dezadeash (115A) (Kindle, 1953), eastern Mount Saint Elias (Kaskawulsh, 115B) (Wheeler, 1963), and Kluane Lake (115G) (Muller, 1967) map-areas. Stratigraphic, paleontologic and structural data indicate that the "Mush Lake Group" (Kindle, 1953) of presumed Triassic and Jurassic (?) age and volcanic rocks correlated with it by Muller (1967) and Wheeler (1963) comprise three or more volcanic assemblages of different ages juxtaposed by movement along major faults (Fig. 39.1). New data necessitate two important revisions to conclusions reached after the first full field season in 1974 of Operation Saint Elias (Read and Monger, 1975; Campbell and Dodds, 1975). These are: (1) The Duke River Fault can be traced far to the south of Jarvis River, where it was previously thought to terminate against the Dalton Fault, and (2) immediately southwest of the Dalton Fault, the volcanic and sedimentary rocks, which were formerly thought to be of possible Devonian age, are now considered to be probably Permian.

Pre-Devonian Volcanic Rocks

Pre-Devonian volcanic rocks lie west of the newly discovered southerly extension of Duke River Fault (Fig. 39.1), in the vicinity of Field Creek (Read and Monger, 1975), and along Alsek River between Marble and Raft creeks. Intrusions and Tertiary sedimentary and volcanic rocks limit the northern and southern extensions of the pre-Devonian volcanics exposed along Alsek River (Fig. 39.1).

Stratigraphy

Along Alsek River several thousand feet of grey-green massive and commonly porphyritic (plagioclase) meta-andesite outlines a major anticline (Figs. 39.2 and 39.3). Scattered sections of pillow lava up to one hundred feet thick give facing directions. Bedded tuffs intercalated with the lowest few hundred feet of volcanics mark the transition into an underlying greywacke-grey phyllite sequence with sparse pebble conglomerate layers. Within 500 feet of the top of the volcanics, thin layers of grey phyllite are intercalated with non-porphyritic meta-andesite. The overlying sedimentary unit changes from tuffaceous greywacke and grey phyllite along Alsek River south of Raft Creek to a massive grey to green siltstone on Alsek River opposite Marble Creek. At Raft Creek, a distinctive white-weathering limestone has fossiliferous argillaceous lenses of probable Devonian age (B.E.B. Cameron, pers. comm.), and is correlated with assurance to the limestone at Marble Creek and with less certainty to the one near Field Creek (Read and Monger, 1975). At Marble Creek, a thin, dark green metavolcanic with white limestone lenses overlies the limestone and underlies a sequence of interbedded grey phyllite, limy grey phyllite and phyllitic or argillaceous limestone of possible Mississippian age (B.E.B. Cameron, pers. comm.).

Structure

Along Alsek River, the volcanic and sedimentary sequence outlines a series of northwest trending folds with the highly variable plunges produced by polyphase deformation (Fig. 39.3). Between Raft and Marble creeks, the repetition of stratigraphy and changes of bedding attitudes, asymmetry of mesoscopic folds, and facing directions define a first phase anticline which trends northwesterly. Although Tertiary sedimentary and volcanic rocks cover much of the northwesterly extension of the fold to and beyond Dusty River, absence of volcanic rocks and presence of limestone and phyllite in that direction infer a regional northwesterly plunge for the anticline. South of Marble Creek, steeply plunging second phase folds overprint the first phase. Underlying Goatherd Mountain and vicinity, a broad belt of phyllite, limy phyllite and grey limestone of possible Mississippian or younger age forms the core of a first and/or second phase syncline. Southwards in Field Creek, a second phase antiform outlined by white limestone is cored by volcanic rocks of Late Cambrian or Early Ordovician age (G.A. Cooper, pers. comm.). These volcanic rocks are probably equivalent to those exposed between Raft and Marble creeks.

Paleozoic Volcanic Rocks

Volcanic rocks southeast of Bates Lake, in an area bounded by Mush, Silver and Squaw (Dollis) creeks on the northeast and Wolverine Creek and Bridge River on the southwest, are of unknown age and uncertain correlation. The Duke River Fault forms the northeastern limit of the volcanics, Tertiary sediments extensively cover them, and they terminate to the south against Paleozoic sediments along the British Columbia-Yukon boundary.

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Stratigraphy

The volcanic rocks are massive, dark grey-green flows with local pillow lava and thin intercalations of limestone and siliceous phyllite. A few miles north of Tatshenshini and Bridge rivers, the volcanic assemblage apparently overlies a thick succession of limestone and interbedded grey phyllite. Southeasterly, in Squaw Creek-Rainy Hollow, British Columbia (Watson, 1948), and Porcupine District, Alaska (Eakin, 1919), fossils from these sediments yield Carboniferous to Permain and Late Pennsylvanian to Early Permian ages, and infer a post-Devonian age for the volcanic sequence. Southeastward extension of the northwesterly plunging structures in the Alsek River implies that the volcanics should be pre-Devonian. At present this contradiction is unresolved.

Permo-Triassic and (?) Older Rocks

Permo-Triassic and (?) older rocks, from the Alaska boundary in the northwest to beyond the British Columbia boundary in the southeast, are bounded by the Dalton-Shakwak faults on the northeast and the Duke River Fault on the southwest. The belt includes Kindle's "Mush Lake Group" in the type area and all other volcanic rocks correlated to the "Mush Lake Group" and "Cache Creek Group" by other workers.

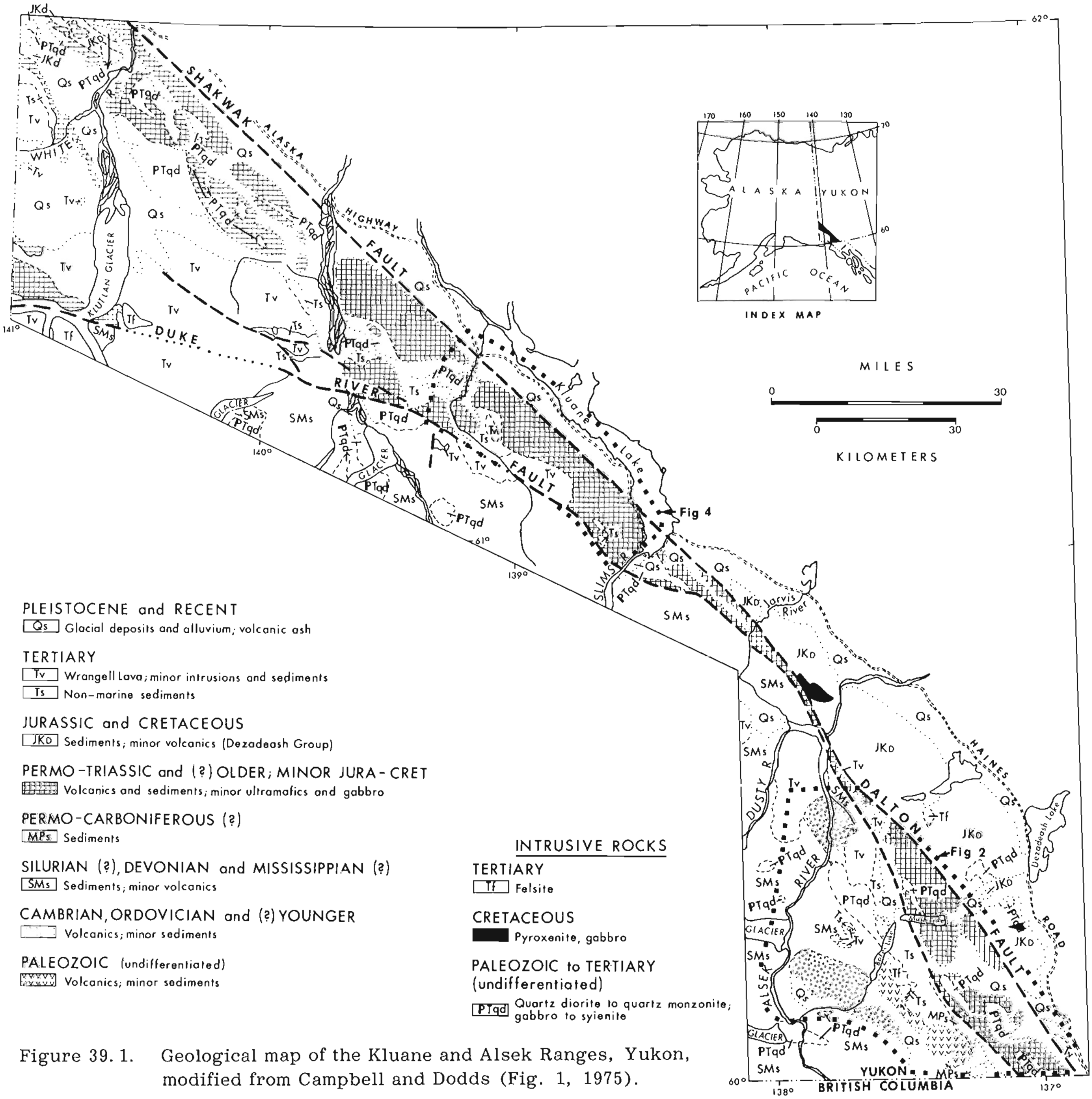

Figure 39. 1. Geological map of the Kluane and Alsek Ranges, Yukon, modified from Campbell and Dodds (Fig. 1, 1975).

Detailed stratigraphic and structural studies begun in the vicinity of Mush Lake and in the area between Duke and Slims River (Read and Monger, 1975), were continued and extended (*see* Figs. 39.2 and 39.4).

Stratigraphy

The stratigraphy described by Read and Monger (1975, p. 57-58) from north of the Duke River, applies with minor changes, southeastward to Jarvis River. From north of Duke River, Permian sediments and underlying volcaniclastics grade southeastward into volcanic flows and volcaniclastics with some phyllite and limestone intercalations. The main stratified units in order of decreasing age, are as follows:

1. Permian and (?) older volcanic rocks: Northwest of Nines Creek, light to medium grey-green

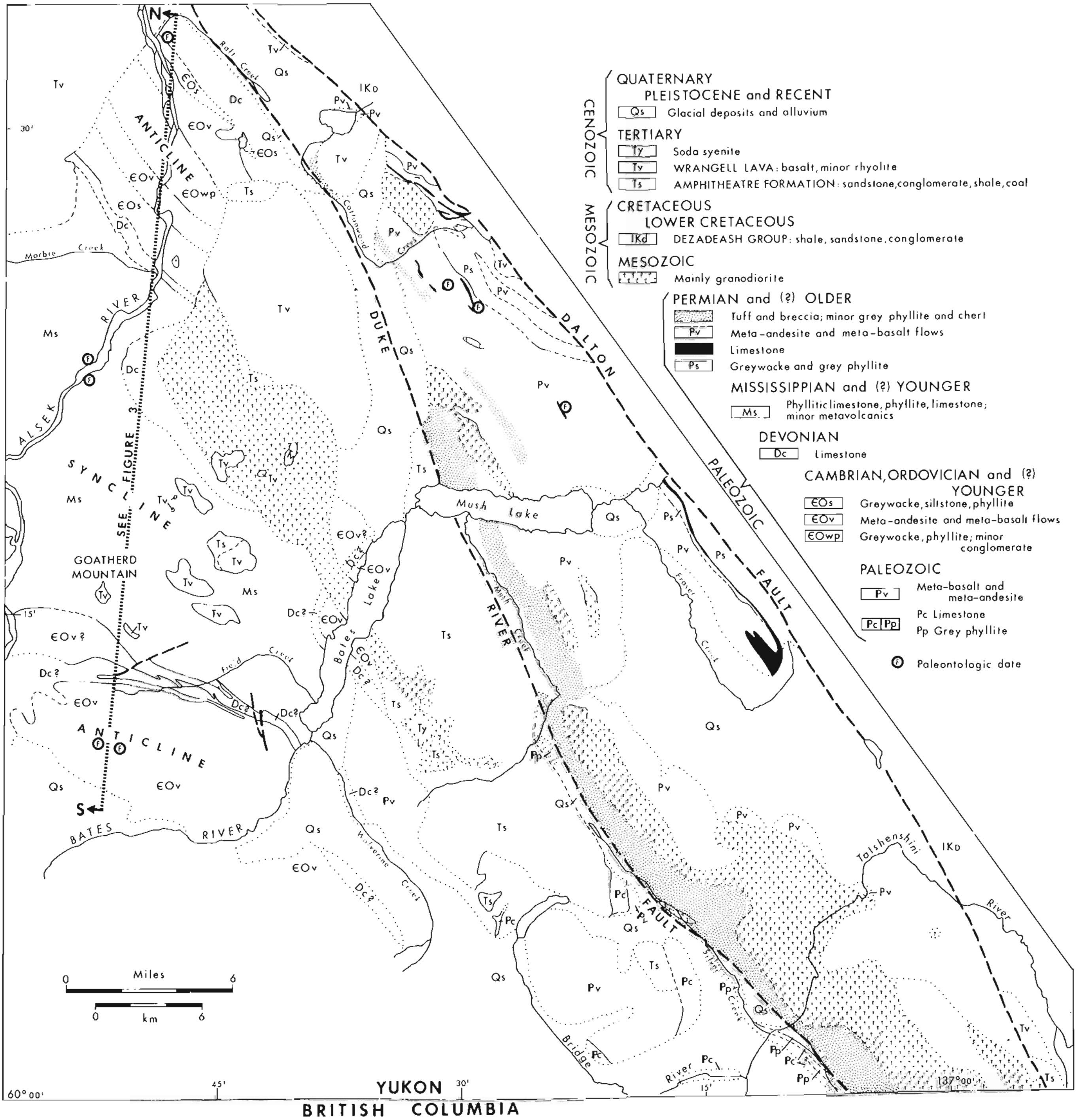

Figure 39.2. Preliminary geological map of the southwest corner of Dezadeash map-area.

volcanic breccia and tuff contain scattered lenses of well-bedded tuff and tuffaceous argillite less than 200 feet thick. Southeast of Nines Creek, tuff and breccia are less abundant and massive green flows, locally amygdaloidal and pillowed, are prominent. Southeast of Williscroft Creek to Jarvis River, basic flows dominate, volcaniclastic rocks are rare, and sedimentary lenses are absent. Southeast of Jarvis River, a Permian section, approximately 20 000 feet thick, consists mainly of flows and volcaniclastics which were previously thought to be probably Devonian (Read and Monger, 1975, p. 56).

2. Permian sedimentary rocks change lithofacies from siliceous argillite, greywacke and chert pebble conglomerate with minor chert and bioclastic limestone in the area between Duke River and the head of Congdon Creek to grey phyllite and conglomerate near Sheep and Fisher creeks. The unfossiliferous conglomerate, composed of angular grey phyllite and Permian volcanic clasts directly underlies Triassic volcanics. Generally the Permian sedimentary sequence overlies the Permian volcanics, but in the north fork of Nines Creek, Permian volcanics are interbedded with and overlie Permian sediments.

3. Triassic volcanic rocks are maroon and dark green metabasalt with prominent chlorite and calcite amygdules. Volcanic breccia, agglomerate and pillow lava are common only within a few hundred feet of the base. The remaining few thousand feet of amygdaloidal volcanics locally display sequences of flows, each 5 to 20 feet thick, with boundaries marked by amygdule-rich zones or thin, layered tuff. The prominent amygdules of Triassic volcanics distinguish them from Permian volcanics.

4. Upper Triassic limestone: Adjacent to Duke River Fault, near Sheep Creek and in Fisher Creek, massive limestone and limestone breccia up to several hundred feet thick, overlie Triassic volcanics and

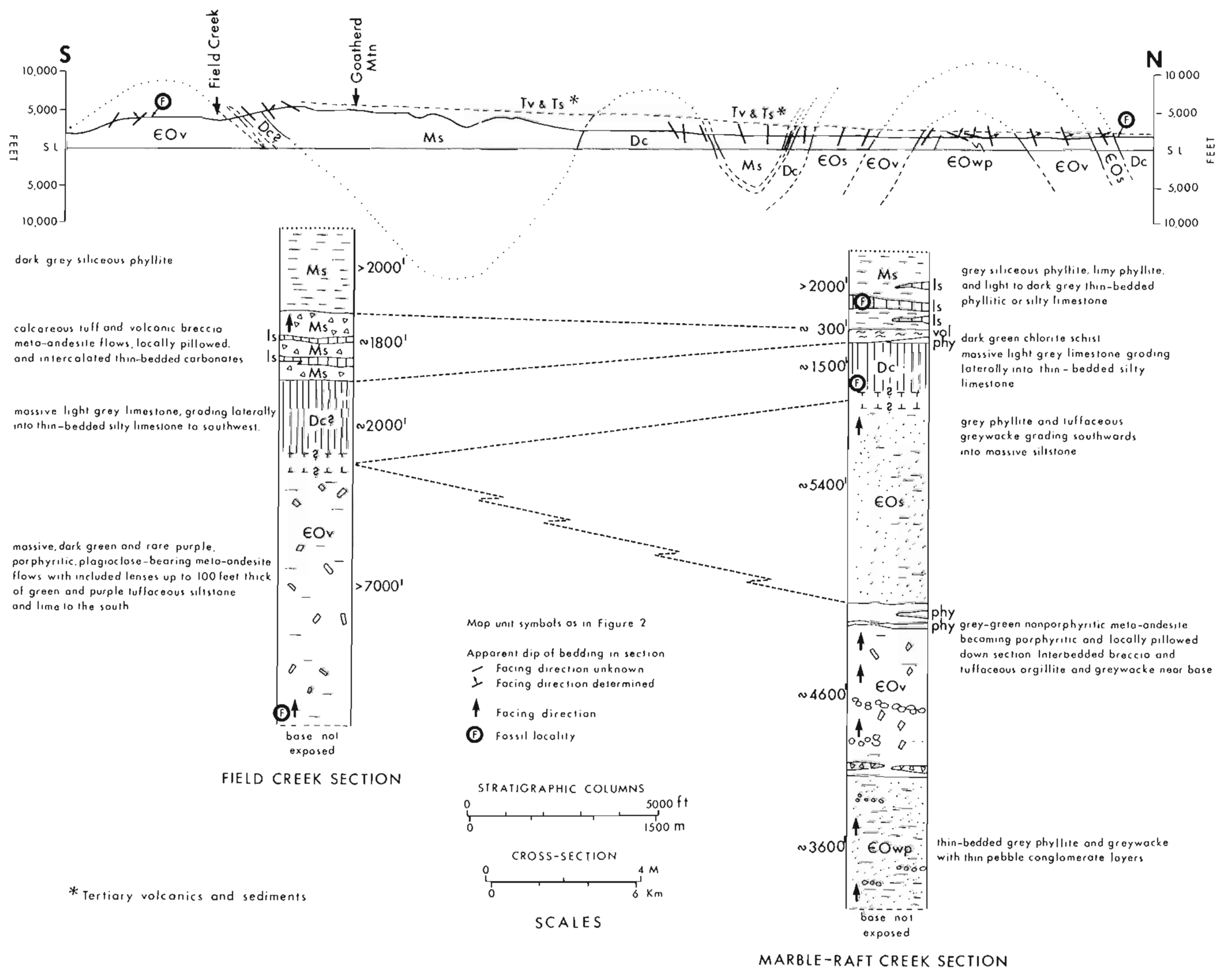

Figure 39. 3. Structure and stratigraphy of Paleozoic rocks exposed along Alsek River, Dezadeash map-area (115A). Pre-Tertiary structures removed by Tertiary erosion are shown dotted above the dashed Tertiary unconformity.

underlie thin-bedded grey limestone. North and northeastwards, the massive limestone thins to several tens of feet or less, forms lenses, contains bioclastic zones, and lies above, or within, Triassic volcanics. These limestones, devoid of macrofauna, locally contain abundant microfauna indicating a Late Triassic age (B. E. B. Cameron, pers. comm.).

5. Upper Triassic and (?) Jura-Cretaceous clastic rocks are mainly grey phyllite, some greywacke and rare, round-pebble conglomerate with sparse granitic detritus.

6. Pyroxene gabbro, clinopyroxenite and peridotite sills extensively intrude Permian rocks and in Williscroft Creek they intrude rocks of Late Triassic age.

Tertiary Rocks

Generally only the location of the boundary between Tertiary and older rocks was determined; the Tertiary rocks were not examined except in the north fork of Bocks Brook. Here the contact between Permo-Triassic and Tertiary rocks dips under the Tertiary at angles ranging from 25 degrees to nearly vertical. Tertiary sediments of the Amphitheatre Formation (Eisbacher, 1975) commonly underlie the Tertiary Wrangell Lava

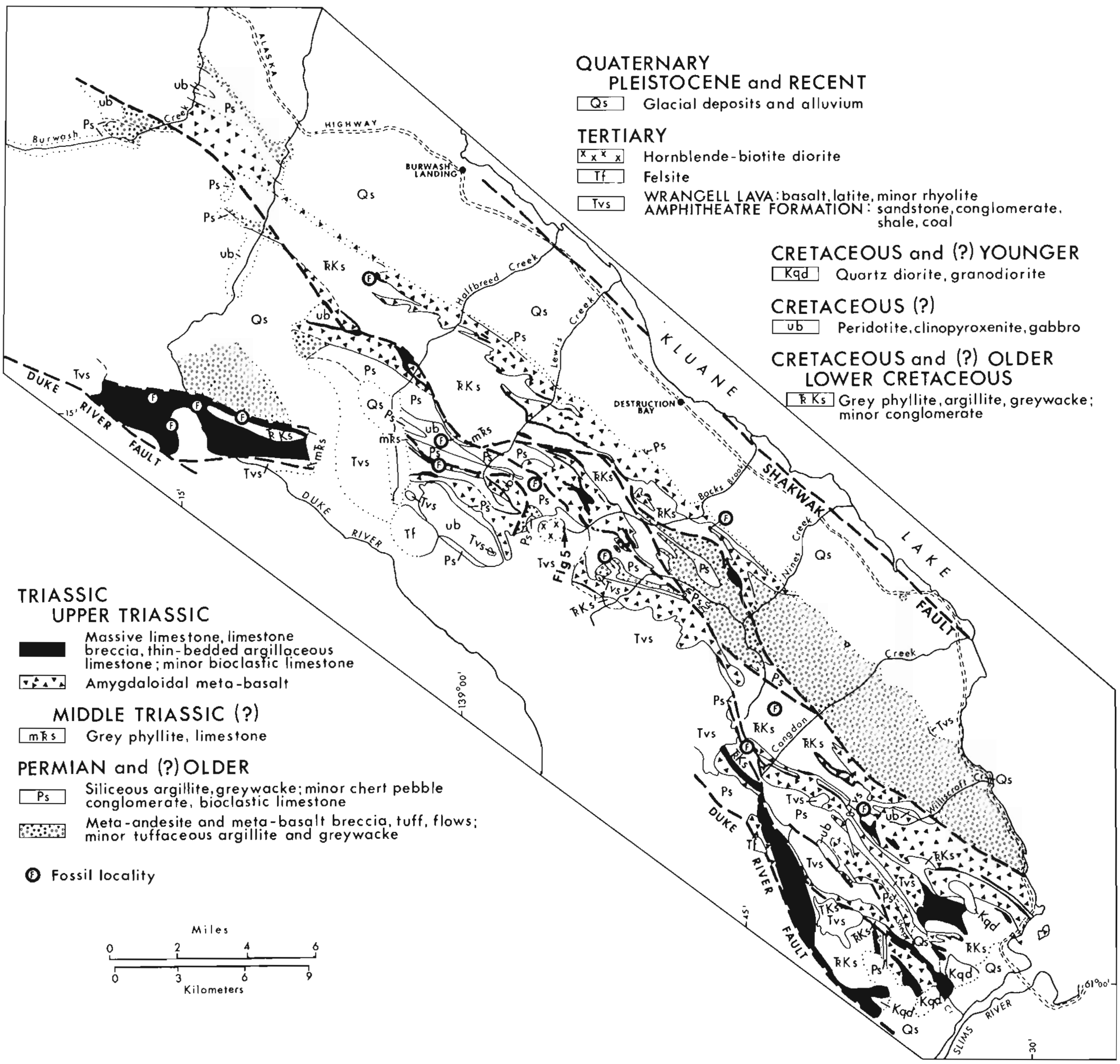

Figure 39.4. Geological map of the Kluane Ranges between Burwash Creek and Slims River.

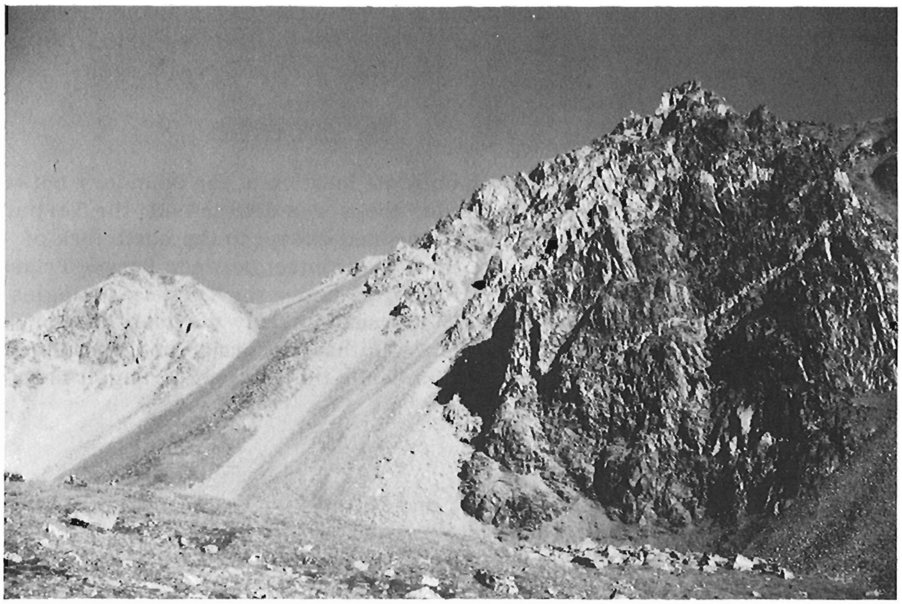

Figure 39.5

At the head of Bocks Brook, stock and dykes of Tertiary diorite (light) intrude Wrangell Lava (dark).

(Souther, 1975), but in this area of steep dips, the sediments are absent. A medium grained, hornblende-biotite diorite stock, about one mile in diameter, and related dykes intrude the Wrangell Lava (Fig. 39.5). These preliminary data suggest a Tertiary vent two miles in diameter, which is cored by a late intrusion. The vent is complete on all but the southern side where it originally may have been breached to allow escape of Wrangell Lava which dips gently southwesterly. In McCarthy quadrangle, eastern Alaska, MacKevett (1970, 1972), reported similar intrusions in Wrangell Lava.

Structure

The Shakwak-Dalton and Duke River faults bound the belt of Permo-Triassic and (?) older rocks. Eisbacher (1975) suggested about 300 km of dextral strike-slip movement on the Dalton-Shakwak-Denali fault system, and slickensides on the Duke River Fault indicate sub-horizontal movement. The Dalton Fault cuts Tertiary rocks but the Duke River Fault, which offsets the base of the Wrangell Lava at least as far southeast as Duke River (Campbell and Dodds, 1975), produces negligible offset of the contact southeast of Slims River.

Permo-Triassic and (?) older rocks are folded along northwesterly trending axes and slivered by subparallel faults. From northwest of Duke River, where folds are broad and faults few, the intensity of deformation increases southeasterly to Bocks Brook where folds are isoclinal, faults numerous and the stratigraphy commonly dismembered. Northwest of a structural culmination in Congdon Creek, folds gently plunge northwesterly with minor reversals. Folds plunge moderately southeast between the culmination and a depression centred between Slims and Jarvis rivers. New data support the earlier structural interpretation (Read and Monger, 1975) that southeast of Jarvis River, Permian rocks form the northeast limb of a northwesterly plunging syncline. Macroscopic parasitic folds on this syncline are exposed just south of Jarvis River and in the heads of Cottonwood and Fraser creeks.

Mineralization

In the Permo-Triassic and (?) older rocks of the Kluane Lake area, copper mineralization is restricted to the Triassic volcanics. Chalcocite and bornite fill fractures and these minerals and native copper form sporadic disseminations in Triassic volcanics. Nickel-copper-platinum mineralization is concentrated along contacts between Permian stratified rocks and basic to ultrabasic intrusions.. Copper deposits at Canyon City and nickel-copper deposits near Quill Creek (Wellgreen Property) and White River (Canalask Nickel Mines) are being investigated by students from the Department of Geology, University of British Columbia. In the Dezadeash map-area, copper mineralization in the Permian and (?) older volcanics forms fracture-fillings of chalcopyrite and bornite in shear zones close to major faults or to the contacts of granitic intrusions.

References

Campbell, R.B. and Dodds, C.J.
1975: Operation Saint Elias, Yukon Territory; *in* Report of Activities, Part A, Geol. Surv. Can., Paper 75-1A, p. 51-53.

Eakin, H.M.
1919: The Porcupine gold placer district, Alaska; U.S. Geol. Surv., Bull. 699.

Eisbacher, G.H.
1975: Operation Saint Elias, Yukon Territory: Dezadeash Group and Amphitheatre Formation; *in* Report of Activities, Part A, Geol. Surv. Can., Paper 75-1A, p. 61-62.

Kindle, E.D.
1953: Dezadeash map-area, Yukon Territory; Geol. Surv. Can., Mem. 268.

MacKevett, E.M.
1970: Geologic map of the McCarthy C-5 quadrangle, Alaska; U.S. Geol. Surv., Map GQ 899.

1972: Geologic map of the McCarthy C-6 quadrangle, Alaska; U.S. Geol. Surv., Map GQ 979.

Muller, J.E.
1967: Kluane Lake map-area, Yukon Territory; Geol. Surv. Can., Mem. 340.

Read, P.B. and Monger, J.W.H.
1975: Operation Saint Elias, Yukon Territory: The Mush Lake Group and Permo-Triassic rocks in the Kluane Ranges; *in* Report of Activities, Part A, Geol. Surv. Can., Paper 75-1A, p. 55-59.

Souther, J.G. and Stanciu, C.
1975: Operation Saint Elias, Yukon Territory: Tertiary volcanic rocks; *in* Report of Activities, Geol. Surv. Can., Paper 75-1A, p. 63-70.

Watson, K. DeP.
1948: The Squaw Creek-Rainy Hollow area, northern British Columbia; Dep. Mines, Bull. 25.

Wheeler, J.O.
1963: Kaskawulsh map-area, Yukon Territory; Geol. Surv. Can., Map 1134A.

40. AN INVESTIGATION OF RIVERBANK AND COASTAL EROSION, BANKS ISLAND, DISTRICT OF FRANKLIN

Project 750079

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Terrain Sciences Division

Introduction

This project was designed to investigate the erosional processes occurring along the riverbanks and coastal areas of Banks Island. It was conducted as part of a larger program examining the hydrology, surficial geology and sedimentary environments (Anderson and Durrant, 1976; Stephen, 1976; Vincent and Gauthier, 1976). The magnitude and sensitivity of many of the erosional processes found along the coasts and riverbanks of the High Arctic require that future development in these areas be undertaken only after a careful evaluation of the conditions at specific sites has been conducted. This study is intended to provide an overview upon which to base more detailed work in the future.

Field work began on May 29, approximately a week before breakup occurred, and continued until August 21, a few weeks prior to freezeup. Specific objectives of the program were as follows:

1) to observe the seasonal regime of riverbank and coastal eorsion and to compare this with the timing of the observed sediment production;

2) to determine the dominant processes affecting the stability of the riverbanks and coasts and to estimate the relative magnitude of the resulting erosion;

3) to examine the relationship between the various erosional processes, the observed rates of erosion, and the resulting morphology. In this context environmental factors such as ice content, surface water supply, and material texture of the bank were investigated; and

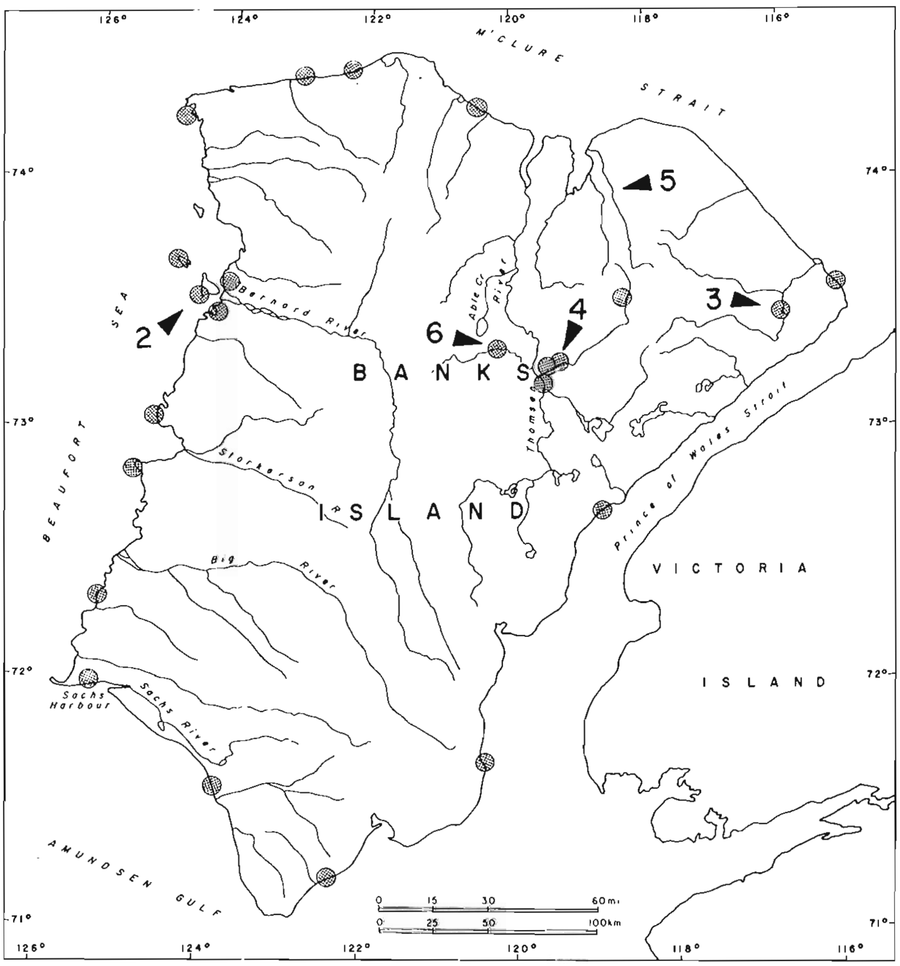

Figure 40. 1.

Map of Banks Island showing the location of field sites and of Figures 40. 2 to 40. 6.

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4) to evaluate the feasibility of using distinctive riverbank and coastal features, which are identifiable on aerial photographs, as indicators of riverbank and coastal stability.

Field Methods

Six river sites representing a range in morphological characteristics were established prior to or during breakup. The relative magnitude of the summer's erosion was monitored through the use of reference stakes and repeated photography. Similarly, sixteen coastal sites were selected in conjunction with a study of sedimentary environments undertaken by Stephen (1976). Areal surveys of most of the major river systems were conducted in conjunction with reconnaissance flights by T.J. Day, project leader. Photographs were taken of representative sites such that additional erosion rates could be estimated through comparison with existing aerial photography. Samples of both the frozen bank and the sediment actually being introduced into the rivers were taken. Data from the second set of samples will be incorporated with the results of studies on fluvial sediment transport undertaken by T.J. Day. The ice content of the frozen samples was estimated in the field from a comparison of wet and dry weights. Grain size analysis and a detailed examination of relevant aerial photographs will be conducted this winter.

Field Results

The seasonal regime of riverbank erosion and sediment production was found to be complex and to vary with locality and process. Prior to breakup, snow melt and sloughing of the active layer contribute significant amounts of fine sediment to the river channels. Eolian material incorporated into snowbanks is locally important, particularly in regions of unvegetated sand deposits. During breakup the low water temperature (near 0°C), the frozen bank, and protection by snow (or snow metamorphosed into ice) inhibit lateral erosion. This condition seems to persist throughout much of the immediate postbreakup period, but high-level strandlines cut into gravel banks indicate that some erosion does occur.

By June 25 (2 to 3 weeks after breakup) the water temperature on the braided channels of Bernard River and westerly draining rivers north of the Bernard ranged from 10.5 to 13°C. (Temperatures were measured with two unshielded thermometers at a depth

Figure 40.2. A series of nivation hollows in low ice-content sands and gravels on the southwest corner of Bernard Island. Despite significant basal erosion and headwall steepening, no headward erosion was observed. An estimated rate of retreat would be less than 20 cm per year over the face of the indicated hollow. Photo taken on June 14, 1975.

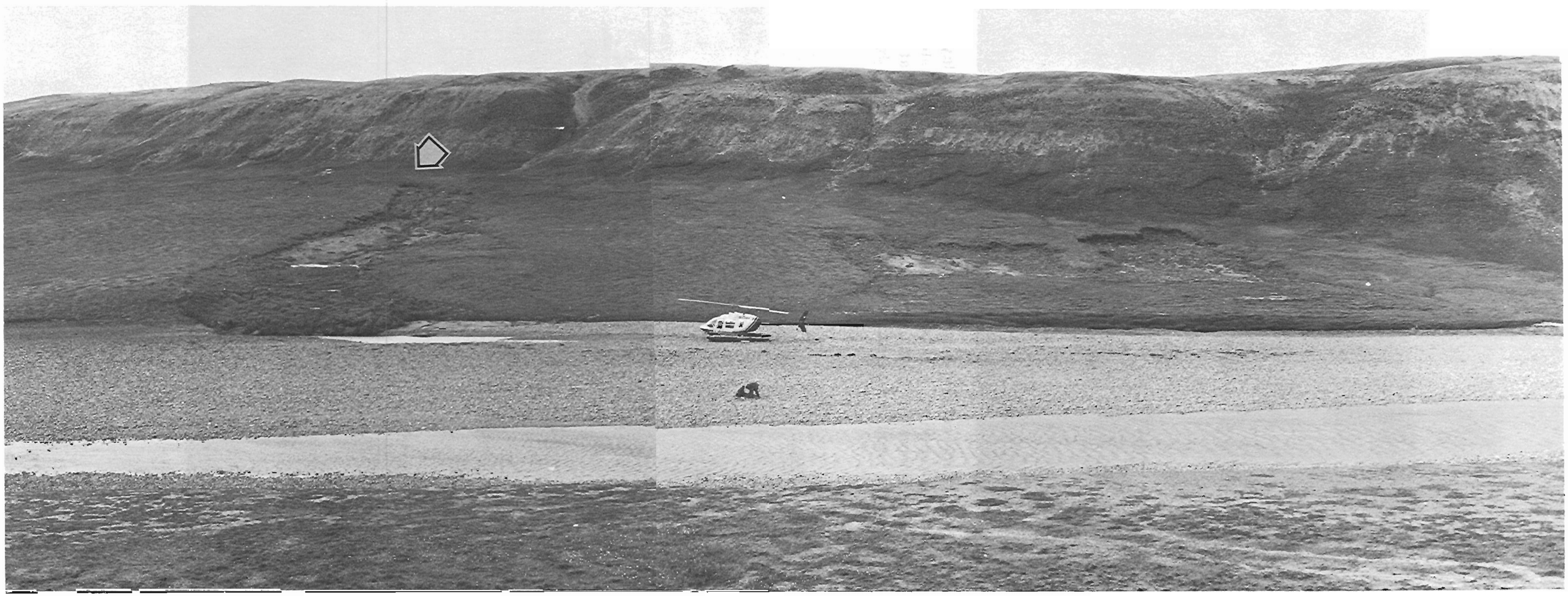

Figure 40. 3. A "skinflow in relatively high ice-content colluviated clay on the Parker River. Retreat of the indicated headwall was greater than 3. 3 m over the summer; however no movement was observed in the terminal deposit. Note the position of the skinflows below gullies which provide a locally increased supply of meltwater. Photo taken on July 29, 1975.

Figure 40. 4. A relatively high ice-content bank consisting of interbedded clay, sand, gravel, and peat through which ice-wedge growth has occurred. The measured rate of retreat was greater than 2. 3 m over the summer. Note the storage of eroded material at the base of the slope. Vertical height of the bank is approximately 16 m. Photo taken on July 25, 1975 on a tributary of Thomsen River.

Figure 40. 6. A "thermal cirque" typical of high ice-content clays. Headward retreat was measured at greater than 3. 5 m over the summer. These features are a source of significant quantities of fine sediment and are a reliable indicator of high ice-content materials. Similar features at coastal sites showed varying rates of retreat depending on the thickness of the high ice-content material. In the northern part of the island lower rates of retreat probably are associated with colder temperatures. Note the two obvious ice wedges, marked with arrows, in the photograph. The width of the face is approximately 50 m. Photo taken July 9, 1975 on a tributary of Thomsen River.

Figure 40. 5 (opposite)

An area of relatively high ice content in a bank composed of sand with gravel inclusions on a tributary to the east of Mercy Bay. Significant thermal niching and the resultant collapsed blocks can only occur when the overlying material has a sufficiently high ice content that it is capable of being self supporting. The collapsed block in this photograph is approximately 10 m wide. Note the mudflow in the foreground. Photo taken on July 27, 1975.

of 30 cm, near the mouths of the rivers.) This corresponds to observations of very active thermal undercutting in the headwaters of the rivers north of the Bernard.

Water levels decreased throughout July and August, and for much of this period most rivers were flowing within their channel zone rather than being confined by their banks. The exposure of the bank and increased air temperatures resulted in active thermal erosion in areas with high ice contents. As a result of the low water levels, however, much of the material was deposited before it reached the active channel and therefore would not be entrained until the high water associated with the following year's breakup. Storm events appeared to have only minor effects on these deposits as the observed rainfall was insufficient to significantly elevate stream levels.

Nivation results from local accumulations of drifting snow and seems to be the dominant process affecting banks composed of material with low ice content. Erosion is greatest at the lower edge of the snowbank due to the increased concentration of meltwater and can result in considerable undercutting and over-steepening of the bank at such sites. Most snowmelt occurs in early June and most sites were free of snow by late July. Downslope movement resulting from undercutting of the base, however, was still occurring in late August. Visual observations seem to indicate that this process may be responsible for much of the lateral displacement of steep gravel banks.

The coastal areas exhibited morphological features similar to those observed on the rivers. Open water existed along the east and south coasts for much of the summer and the resulting wave action was responsible for both thermal erosion and sediment transport. Ice-push features were observed on all the coasts; the north-western corner and the northern coast of the island were areas where ice push was of significant magnitude. For example a ramp of pack ice was observed 150 m inland and 15 m above sea level in the vicinity of the westernmost beach site on McClure Strait (Fig. 40.1). Similar examples of ice action were observed at Cape Wrottesley on the northern tip of the island. It is interesting to note the relative lack of colluvial material and the sharpness of the break in slope between the foreshore and the bank, that occurs in areas which appear to be commonly icefast. As the banks in these areas are actively eroding, the resulting colluvium must be periodically removed through either ice scour or infrequent wave action. At present the relative importance of these two processes in unknown.

Typical relationships between process, morphology, terrain composition, and rates of erosion can be observed in Figures 40.2 to 40.6. The locations of the photographs are indicated on Figure 40.1.

References

Anderson, J.C. and Durrant, R.L.
1976: Hydrologic reconnaissance, Thomsen River basin, Banks Island, District of Franklin; *in* Report of Activities, Part A, Geol. Surv. Can., Paper 76-1A, report 45.

Stephen, W.J.
1976: A reconnaissance study of the coastal processes on Banks Island, District of Franklin; *in* Report of Activities, Part A, Geol. Surv. Can., Paper 76-1A, report 56.

Vincent, J.S. et Gauthier, R.C.
1976: Inventaire des dépôts de surface de l'Ile de Banks, District de Franklin; *in* Report of Activites, Part A, Geol. Surv. Can., Paper 76-1A, report 6.

TERRAIN INVENTORY: BATHURST, CORNWALLIS, AND ADJACENT ISLANDS, NORTHWEST TERRITORIES

Project 740067

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Terrain Sciences Division

Introduction

The purpose of the terrain inventory in this part of the Arctic Islands is twofold: (1) it constitutes part of the ongoing national program to describe surficial geology and glacial history; and (2) it is also part of the Terrestrial Environment Program (TEP-1) for the Environmental-Social Program, Northern Pipelines, which requires evaluation of the terrain, including botanical and pedological components.

Bathurst, Byam Martin, Cameron, Little Cornwallis, Cornwallis, and several adjacent islands were studied (Fig. 41.1) and data were gathered to produce maps at a scale of 1: 125 000 for an area of 11 000 square miles (28 000 km^2).

Detailed assessment of terrain performance was carried out simultaneously by P.J. Kurfurst (Project 740046) at selected sites already disturbed and at others showing some indication for potential disturbance.

Transport was primarily by Bell 206B helicopter with additional ground mobility provided by Honda ATC 90 motor tricycles. A modified CRREL coring barrel with a Haynes power unit was used to obtain subsurface information at selected sites. The stony nature of most surface materials made this drill unsuitable for much of the terrain. Near-surface data were collected using vane test apparatus for shear strength of silts and clays, penetrometers for evaluating penetration resistance of uniform fine grained materials, and hydrochloric acid for testing for carbonates. Surface samples were collected for textural and behavioural analysis. Per cent cover of selected plant communities were measured by pin frame.

The TEP-1 portion of the project was the first opportunity to test the pilot-integrated mapping system

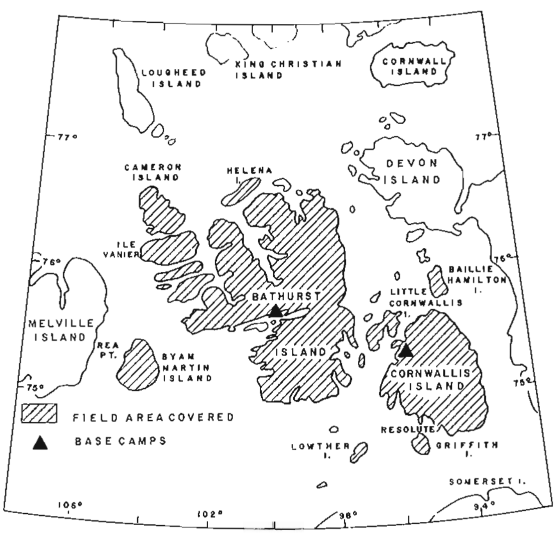

Figure 41.1

developed on eastern Melville Island (Barnett *et al.*, 1975a; Barnett *et al.*, 1975b) in an adjacent area which had not been visited by any of the participants previously. For TEP-1 a soils (pedological) component is planned for the expanded legend, the data for which were gathered by C. Tarnocai of the Soils Research Institute.

Procedures

Aerial photographs at a scale of 1: 60 000 were used for a preliminary interpretation of vegetation cover and surface materials. Following the method used on Melville Island, formation boundaries were plotted onto the 1: 125 000 airphoto mosaics, which were to be used as mapping bases in the field because major structural and material differences associated with the different bedrock types are generally quite distinct. In places where the boundaries were not clear they were plotted with the aid of the published bedrock maps (Kerr, 1974; Thorsteinsson, 1972; Thorsteinsson and Kerr, 1968).

Obvious terrain-geobotanical boundaries also were delineated and possible sampling sites were selected. These sites represented typical surface materials and vegetation, areas that had potential for unravelling the Quaternary history, or areas that were difficult to interpret.

About 200 square miles (500 km^2) per day were mapped in order to cover the area assigned. On this basis a maximum of ten detailed stops were possible for any one day.

In the field at and around the selected sites, detailed regional and local field descriptions of terrain, botanical, and ground ice conditions were recorded, and at least one pit was dug to the frost table. Pit materials were noted and at least one channel sample was taken. Generally one sample per site was sufficient because any original stratification commonly had been disturbed by frost churning. Where appropriate, strength measurements were made along pit walls and river bank exposures. A total of 276 surface samples were collected from Bathurst Island and 184 were collected from Cornwallis Island. Two hundred CRREL core samples also were taken from 15 clusters of sites.

Locations of sites visited were plotted on 1: 125 000 mosaics and later were transferred to 1: 250 000 topographic maps at base camp. In this way areas covered and remaining gaps could be appreciated; on the basis of helicopter reconnaissance and information obtained from stops selected by preliminary interpretation, key areas and additional sites of potential interest were determined.

The mapping technique was modified from that used on Melville Island. At that time daily mapping was confined to one Landscape Type unit so that its uniformity and range of variability could be appreciated. This method was not used on Bathurst and Cornwallis islands, partly for convenience and partly due to weather restrictions. Flying conditions were marginal for much of the time so that it was necessary to find an area which was not fog-bound and map that portion of the island as a block. Because of the large area, the flying time involved, and the distribution of Landscape Types, it was more convenient to map in blocks. In this way only one or two mosaics were required in the helicopter, rather than six or seven which would have been necessary to map by Landscape Type.

Detailed navigation at low levels was a problem because of the lack of distinctive terrain detail over large areas, especially over the central part of Cornwallis Island.

Major difficulties were encountered with the drilling system, resulting in inadequate information about ground-ice characteristics in coarse materials. Coring was confined generally to those areas having fine grained frozen materials or organic deposits. The maximum length of core extracted at any site was 215 cm, and in most cases only about 75 cm depth was achieved. Many holes, therefore, did not penetrate into perennially frozen ground although core was obtained from materials which were below the active layer during the summer of 1975.

Another problem affecting site selection was encountered. The double objectives of determining Quaternary history and establishing the nature of the surface materials are not logistically compatible in this part of the Arctic Islands, since the Quaternary materials *per se* are very limited in area. Weathered bedrock rather than Quaternary deposits constitute most of the surface materials. Since 500 km^2 a day had to be mapped to cover the required terrain, little time could be spent in determining Quaternary stratigraphy and materials.

The Mapping System

The system of integrated terrain mapping that was developed on Melville Island was successfully extended to adjacent areas. Similar repetitive land patterns were observed across the sedimentary basins of the central Arctic Islands. As the system was extended to Bathurst and Cornwallis islands, new Landscape Types (such as Disappointment Bay Limestone) were added to those previously recognized, and a few additional geobotanical and terrain descriptors were dovetailed into the preexisting system where required. On the Griper Bay Sandstone, for instance, an additional nonslumping, poorly vegetated facies was delineated on Bathurst Island; similar Griper materials were not present on Melville Island. A rubble facies, significant for trafficability, was added to the Hecla Bay Sandstone Landscape Type.

The application of the system to Cornwallis Island is proving more difficult than for Bathurst Island. Although there are a number of different bedrock formations, they are mainly carbonates of lower Paleozoic age outcropping as a uniform plateau. Although these different bedrock types have locally recognizable differences in texture, vegetation, and geomorphic detail, they are not marked by significant regional differences that have led to development of visually distinctive terrain types. In addition faulting and other structural disruption associated with the Cornwallis fold belt has produced numerous small outliers and inliers. These units were too numerous to allow adequate ground checking and have surface characteristics sufficiently similar to those of the adjacent rock types to prevent accurate mapping by airphoto interpretation.

On Melville Island and on Bathurst and Cornwallis islands the basic mapping unit that has been used is Landscape Type. It is important to reiterate here that terrain and geobotanical characteristics are directly related to the nature of the bedrock and that their patterns change abruptly at bedrock boundaries. Although the entire area has been glaciated, neither conspicuous erosional nor depositional forms associated with glaciation are present here. Most of the surface is frost-shattered bedrock which breaks down to its characteristic forms according to its primary sedimentary characteristics and the later structural stresses that have been applied to it. On Melville Island the marine limit was used as a Geobotanical Facies boundary. The marine limit is a real and useful boundary on Bathurst and Cornwallis islands as well. Marine influences can be readily identified in many areas, exhibited as changes in surface materials, geobotanical patterns, or ground ice conditions or combinations of these.

The extended legend format of data presentation can be adequately applied to both Bathurst and Cornwallis islands to provide a rapid system of information retrieval, but in the case of Cornwallis Island, some of the Landscape Type will have similar descriptions.

Vegetation

A general correlation of surface materials with plant community composition and cover was apparent on Bathurst and Cornwallis islands and adjacent smaller islands, a trend noted earlier on Cornwallis Island by Thorsteinsson (1958, p. 13). These relationships often assisted with the delineation of Landscape Types and their subdivisions.

Soil moisture availability has a marked influence on the per cent vegetation cover and in some cases the composition of plant communities. The development of a dense cryptogamic lower stratum was directly related to the presence of a sufficient amount of moisture throughout the growing season.

Bathurst Island showed more dense and diverse plant communities than did Cornwallis Island. The latter was extremely barren as noted by Arkay (1972) and Schofield and Cody (1955), except for the lowlands to the northwest and west and along several major river valleys.

The extensive barren areas of both major islands are coincident with the distribution of fragmented limestone and dolomite. Elsewhere saxifrage-based communities were most common, with willow, poppy, or grass in places co-dominant. *Salix-Dryas* dwarf shrub communities were present but were limited in distribution, most commonly found on recent gravel alluvial terraces. *Luzula*-based tussock communities were restricted almost entirely to Griper and Hecla Bay sandstones on Bathurst Island and adjacent islands. On coastal lowlands willow was in places co-dominant. Graminoid wetlands and sedge meadows associated with abundant moisture (i. e. seepage slopes and around ponds), were present on both major islands but the existence of substantial sedge meadows on Cornwallis Island was very local.

On each traverse the types and numbers of birds and mammals were recorded. Preliminary analysis shows a relationship in summer between presence of caribou and muskoxen and certain plant communities.

Engineering Aspects

The surface materials throughout the study area do not present major engineering problems. Most are lithified, competent sandstones and calcarenites. There is little evidence of regional slope stability problems except for the Eids Shale. Very few thaw-flow scars were observed; landslides were limited to talus slopes along the cliffed portions of the coast. Local minor stability problems may arise from the melting of ground ice in a small percentage of the total area. These areas can be easily avoided, except for the Eids Shale which repeats as narrow east-west trending bands on Bathurst Island. Low bearing strengths could prove to be a problem on the long, low-angle seepage slopes which prevail on some of the upland areas of Bathurst Island (generally on the Bathurst Island and Bird Fiord Landscape Type) and on the extensive plateau area of central Cornwallis Island. Stability problems would arise for both travel and trenching activities in these areas during the period following snowmelt and heavy rains.

Over most terrain, trafficability limitations are determined by relief and degree of dissection rather than by behavioural problems associated with the surface materials themselves. Northern Bathurst Island and the adjacent islands, and the eastern coastal area of Cornwallis Island are rugged, with local relief up to 250 m.

Potential Bedrock Boundary Revisions

Some minor additional occurrences of bedrock formations have been tentatively identified. In particular, the K22a unit (Tozer and Thorsteinsson, 1964, p. 158), a Cretaceous black silt which mantles Paleozoics on central and southern Melville Island, has been extended to Byam Martin Island. A small patch also may be present on the Allen Bay Formation on southwestern Cornwallis Island. Other tentative additions are Tertiary sands and silts, found along a low coastal zone on eastern Bathurst Island and in small graben on Cornwallis Island. Beaufort Gravels also may be present on eastern Bathurst Island. These additions are not yet confirmed. Some minor alterations of the boundaries of Hecla Bay Sandstone and Bird Fiord Formation are proposed for Byam Martin and Bathurst islands.

Glacial History

Earlier studies which have commented on the glacial history of Bathurst Island (Blake, 1964, 1970, and 1974) and Cornwallis Island (Thorsteinsson, 1958) have been based either on field work which did not cover the entire area or in the case of Cornwallis Island, on work primarily aimed at nonglacial objectives.

The entire area examined has been glaciated but the terrain is not a typical glacial landscape, and many of the classical glacial indicators are either missing or poorly represented.

Shield provenance igneous erratics have a widespread distribution on all the major islands at a variety of elevations and with a size range from granules to boulders. This distribution suggests that Laurentide ice once reached well north of the limit indicated on the Glacial Map of Canada (Prest *et al.*, 1968). The glacial event which distributed these must have been preclassical Wisconsin in age and possibly coincident in time with the event on eastern Melville Island which left patches of weathered till on upland surfaces (Barnett *et al.*, 1975b). A substantial upland zone of southeast Cornwallis Island carries a thin till cover with abundant igneous erratics, but patches of similar provenance and similar age were not located on Bathurst Island. Granule size igneous erratics on Bathurst Island were common on raised alluvial surfaces; this distribution also suggests an extensive till cover but one of considerable antiquity.

The presence of the former Innuitian ice sheet proposed by Blake (1970) has been ably demonstrated by reference to marine uplift, the magnitude of which would demand a substantial ice mass over Bathurst and Cornwallis islands. Additional evidence for a postglacial marine episode includes abundant shells, whalebone, and minor amounts of driftwood. Morphological and depositional evidence for land emergence includes abundant raised beaches in morphologically suitable localities, raised deltas and marine silts admixed with weathered mantle which support an enriched vegetation.

Direct evidence of the Innuitian ice sheet is more difficult to substantiate. Till from the Innuitian episode is difficult to identify conclusively because: (a) erratics are "local" sedimentary rocks (as opposed to Shield rocks) and their provenance is uncertain due to commonly recurring lithologies; (b) active cryoturbation is mixing the surface materials from the top 50 to 75 cm. These materials therefore may include erratics from an earlier episode mixed with weathered local bedrock materials; (c) till may not have been very abundant as the Innuitian ice sheet would have a thermal regime more 'polar' than that of the Laurentide ice sheet, and therefore the basal layers of the ice may have been relatively inactive; (d) strong surface material boundaries commonly prevail in the weathered mantle coincident with mapped bedrock boundaries despite cryoturbation and the effects of glaciation.

Glacial Landforms

Glacial drainage channels are the most abundant and well developed landforms. The ice marginal forms tend to conform to a pattern suggesting that ice persisted on the higher terrain of Bathurst Island. Cirques are present on northern Bathurst Island and eastern Cornwallis Island, but they are poorly developed.

A few landforms which have the appearance of eskers occur on both Bathurst and Cornwallis islands, but cursory inspection of the constituent materials offers no support for a fluvioglacial origin. Materials observed consisted of very angular, platy, single-lithology bedrock without apparent bedding. Topographic position and morphology, however, support the identification as eskers. No other obvious alternative explanation occurs.

Fiords were identified by Thorsteinsson (1958, p. 15) on the east coast of Cornwallis Island; some of the steep-sided drowned valleys of northern Bathurst Island also may be glacially modified. Moraines are poorly developed, isolated, and few in number; one of the largest seen was on the northwest coast of Cornwallis Island (Thorsteinsson, 1958, p. 15). Only one deposit identified as a kame was observed at Dundee Bight on Bathurst Island.

Ice-movement indicators such as striae were not common but where they were observed on central Lowther Island and east central Cornwallis Island, they were well developed. On Lowther Island a northward ice flow was indicated at N10°E. This direction, plus the distribution of Laurentide erratics in 'fresh' till, suggest that the late Wisconsin (?) Laurentide ice margin reached north of Lowther Island but perhaps not to Cornwallis Island. On east central Cornwallis Island striae were oriented perpendicular to the coast, possibly reflecting late ice movement from a central remnant ice mass.

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References

Arkay, K.E.
1972: Species distribution and biomass characteristics of the terrestrial vascular flora, Resolute Bay, N.W.T.; unpubl. M.Sc. thesis, Dept. Geography, McGill Univ., Montreal.

Barnett, D.M., Edlund, S.A., and Dredge, L.A.
1975a: Interdisciplinary environmental data presentation for eastern Melville Island: an approach; *in* Report of Activities, Part B, Geol. Surv. Can., Paper 75-1B, p. 105-107.

Barnett, D.M., Edlund, S.A., Dredge, L.A., Thomas, D.C., and Prevett, L.S.
1975b: Terrain classification and evaluation, eastern Melville Island, N.W.T., vols. I and II; Geol. Surv. Can., Open File 252, 747 and 571 p., 3 maps, 3 legends.

Blake, W., Jr.
1964: Preliminary account of the glacial history of Bathurst Island, Arctic Archipelago; Geol. Surv. Can., Paper 64-30, 8 p.

1970: Studies of glacial history in Arctic Canada. I. Pumice, radiocarbon dates and differential postglacial uplift in the eastern Queen Elizabeth Islands; Can. J. Earth Sci., v. 7, p. 634-664.

1974: Studies of glacial history in Arctic Canada. II. Interglacial peat deposits on Bathurst Island; Can. J. Earth Sci., v. 11, no. 8, p. 1025-1042.

Kerr, J.W.
1974: Geology of Bathurst Island group and Byam Martin Island, Arctic Canada; Geol. Surv. Can., Mem. 378, 152 p.

Prest, V.K., Grant, D.R., and Rampton, V.N.
1968: Glacial Map of Canada; Geol. Surv. Can., Map 1253A.

Schofield, W.B. and Cody, W.J.
1955: Botanical investigations on coastal southern Cornwallis Island, Franklin District, N.W.T.; Can. Field Nat., v. 60, p. 116-1128.

Thorsteinsson, R.
1958: Cornwallis and Little Cornwallis islands, District of Franklin, Northwest Territories; Geol. Surv. Can., Mem. 294, 134 p.

1972: Five unedited geological maps of the Canadian Arctic Archipelago, legend, notes; Geol. Surv. Can., Open File 139.

Thorsteinsson, R. and Kerr, J.W.
1968: Cornwallis Island and adjacent smaller islands, Canadian Arctic Archipelago; Geol. Surv. Can., Paper 67-64.

Tozer, E.T. and Thorsteinsson, R.
1964: Western Queen Elizabeth Islands, Arctic Archipelago; Geol. Surv. Can., Mem. 332.

SURFICIAL GEOLOGY MAPPING: VALLEYFIELD-RIGAUD AREA, QUEBEC (31 G/1, 8, 9)

Project 740068

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Field investigation and mapping of the surficial geology of the Vaudreuil (31 G/8), the northwest quarter of the Huntingdon (31 G/1), and the southwest quarter of the Lachute (31 G/9) map-areas has been completed. A study of the groundwater resources and the bedrock surface of the Vaudreuil map-area was carried out for the Geological Survey of Canada in 1960 and 1961 by Tremblay and Hobson (1962). The unconsolidated Quaternary deposits are well developed in this area in surface extent and in thickness over both the Paleozoic limestone and sandstone platform and the Precambrian basement uplands of the Rigaud and Oka mountains. The bedrock outcrops mainly on Rigaud Mountain immediately south of the town of Rigaud, on Oka Mountain north of the town of Oka, and on both sides of the valley of the St. Lawrence River at Valleyfield.

Glacial Deposits

Glacial deposits make up about one quarter of the land surface of the Vaudreuil map-area. They occur mainly as an undulating to rolling mantle of grey, calcareous, silty or sandy, compact ground moraine — mainly derived locally — covering flat-lying or hilly and hummocky bedrock. Till plains are all small in extent; the largest is in the southeastern part of the map-area around Valleyfield and St-Timothée and a smaller one occurs on the southern lower slopes of Rigaud Mountain north of Ste-Marthe. Most of the glacial sediments occur as a rolling, hummocky, or ridged moraine running in a belt across the map-area from Glen Nevis in the southwest to La Trappe d'Oka in the northeast.

This morainic belt is not continuous but consists of areas of low relief, hummocky topography, and parallel till ridges separated by areas of shallow basins filled with marine sediments. The morainic ridges are all oriented northeast-southwest and are well developed at and west of St-Telesphore, west of Ste-Justine-de-Newton, south and east of Très-St-Rédempteur, and at Rigaud where they wrap themselves around the lower part of the northwestern and northern sides of Rigaud Mountain. They are not present east of Rivière-Beaudette and Ste-Justine-de-Newton. Their northeast-southwest orientation is roughly perpendicular to the orientation of the numerous drumlins that are developed to the west and southwest in the southern part of the Alexandria map-area and throughout the Cornwall map-area. Both types of glacial features gradually overlap and interfinger into one another, suggesting formation by a single ice lobe advancing from a source area that lay to the northwest somewhere in the Laurentide Mountains north of the Ottawa River.

Fossil marine shells in a matrix of glacial till have been found in fresh exposures in hummocky or ridged moraine at many localities over a distance of more than thirty miles between the north shore of the St. Lawrence River at Rivière-Beaudette in the south and La Trappe d'Oka on the north shore of the Ottawa River in the north. Only a few of these localities are mentioned here.

About one mile northwest of Ste-Justine-de-Newton in an area of well developed hummocky and ridged moraine, a roadside cut 5 m (16 feet) deep through a morainic ridge exposes fossiliferous till about 3 m (10 feet) below ground surface. Till is the only sediment present throughout the section. Fossil shells of the following marine molluscs were found in a sample of this till that came from an elevation of 375 feet a. s. l. and were identified by the author: *Mya arenaria, Hiatella arctica, Mya truncata,* and *Macoma balthica.* At a site 2 miles east of Très-St-Rédempteur on the lower slopes of the western flank of Rigaud Mountain, fossiliferous glacial till is exposed in a small creek bed in an area of well developed hummocky and ridged moraine where the moraine is attached to the west side of the mountain. The creek bed is cut to a depth of 2m (6 feet); the elevation of the collection site was 335 feet a. s. l. Fossil shells of the following marine molluscs were recovered: *Balanus hameri, Hiatella arctica, Mytilus edulis,* and *Macoma balthica* (Fig. 42. 1).

At several localities the fossiliferous 'till' underlies undisturbed fossiliferous beach materials. One such site is two miles southwest of Ste-Justine-Station in Vaudreuil County where a freshly cut ditch opened for field drainage purposes at an elevation of 250 feet a. s. l. exposes a fossiliferous till overlain by fossiliferous beach gravels and sands 2 m (6 feet) below the ground surface. The beach materials wedge out into a water-washed bouldery hummocky moraine surface towards the west. The till is very clayey and silty, grey in colour, with a large number of blue limestone pebbles and cobbles derived from the local Ordovician Ottawa limestone; it includes fossil shells of *Hiatella arctica, Mya arenaria,* and *Macoma balthica.* Another locality of this type is one-half mile southwest of Ste-Justine-de-Newton, where fossiliferous till is exposed on the floor of a borrow pit underlying a fossiliferous marine beach gravel ridge at an elevation of 370 feet a. s. l.

Fossiliferous diamicton, associated with ice-contact materials and interpreted as till, occurs at a large number of localities. A typical example is one mile south of the town of Rivière-Beaudette where a fresh gravel pit 8 m (26 feet) deep is cut through a well developed ridge of gravel and sand that rises above the surrounding marine clay plain. Exposures in this pit reveal numerous clayey or silty fossiliferous till lenses which vary from less than 1 m (3 feet) to more than 3 m (10 feet) in thickness. They lie above an unfossiliferous deposit of well sorted and bedded ice-contact gravels, which forms the core of the ridge,

Figure 42. 1

Fossiliferous till exposed on the floor of a small creek cutting into the Ste-Justine-de-Newton ridged moraine where it is anchored against the west side of Rigaud Mountain. Photo taken two miles east of Très-St-Rédempteur, Vaudreuil County, Quebec.

Figure 42. 2

Fossiliferous till lens exposed inside the core of the Rivière Beaudette ice frontal moraine. Photo taken one mile south of the town of Rivière-Beaudette in Soulanges County, Quebec. Note sharp irregular contact with unfossiliferous fluvioglacial outwash gravels below the till unit.

and below fossiliferous marine beach gravels and sands (Fig. 42. 2). At this site fossil shells of *Hiatella arctica, Mya arenaria, Macoma balthica,* and *Balanus hameri* were recovered from the till samples collected at elevations of from 175 to 185 feet a. s. l.

Another such occurrence is at a locality 2 miles south of Très-St-Rédempteur where the northeast-southwest trending morainic belt is barely a mile in width. Here at an elevation of 260 feet a. s. l. in an active gravel pit, lenses of fossiliferous till are exposed adjacent to deformed fossiliferous marine beach gravels 2 m (6 feet) from the ground surface. They are overlain by 1. 5 to 2 m (5 to 6 feet) of undisturbed fossiliferous beach gravels and are underlain by and interfinger with unfossiliferous ice-contact or ice-frontal fluvioglacial outwash sands and gravels. Several fossiliferous beds of former marine shore or nearshore deposits are severely deformed, possibly due to overriding by advancing glacial ice, so that the shells are now pressed together and broken (Fig. 42. 3)

Another example of this type lies north of Ottawa River 3 miles north of Oka, where lenses of fossiliferous till are exposed in the walls and on the floor of an active gravel pit adjacent to deformed and overridden fossiliferous marine beach gravels. *Hiatella arctica, Mya arenaria, Balanus hameri,* and *Macoma balthica* were identified in a sample collected at a depth of 7 m (23 feet) and at an elevation of 450 feet a. s. l.

These occurrences of marine shells in till are thought to be further evidence of a readvance of ice into the Champlain Sea basin about 11 200 radiocarbon years B. P. (Richard, 1975). Dates on shell samples collected this past summer should confirm the age of this event.

Figure 42. 3

Fossiliferous till lens adjacent to deformed fossiliferous marine beach gravels exposed in the wall of a gravel pit opened into the Ste-Justine-de-Newton ice-frontal moraine. Photo taken two miles south of Très-St-Rédempteur. Vaudreuil County, Quebec.

Marine Sediments

Postglacial marine sediments deposited in the Champlain Sea, following the withdrawal of Wisconsin ice, are more widespread than the glacial deposits and make up approximately three-fifths of the land surface of the areas surveyed. The most important of these in area and size consists of grey, massive, blocky, unctuous, calcareous, fossiliferous marine clays and silts which were deposited in depressions and basins in flat-lying Paleozoic limestones and sandstones and in long, narrow depressions found in areas of hummocky and ridged moraine. They occur up to an elevation of 220 to 245 feet a. s. l. in the northern and western parts of the map-area, to an elevation of 175 to 200 feet a. s. l. in the central part, and only to an elevation of 140 to 165 feet a. s. l. in the southern and eastern parts of the map-area in St. Lawrence River valley. These deposits form gently sloping clay plains whose surfaces are primary surfaces of marine sedimentation. The most extensive deposits are in four major districts: 1) in the central part of the map-area between St-Télesphore, Ste-Justine-Station, Ste-Marthe, and St-Lazare-Station where they form a continuous marine clay plain interrupted in few places by low glacial till knolls; 2) in the northwest between Glen Robertson, Ste-Anne-de-Prescott, and Très-St-Rédempteur where the clay plain is much smaller in extent and is interrupted much more by small tabular bedrock uplands, mounds, or hummocks of glacial till; 3) in the St. Lawrence River valley upstream and downstream from Valleyfield where the clay plain surface also is interrupted by small undulating till plains or tabular bedrock outcrops; and 4) in the Ottawa River valley on both sides of the river where the marine clay has filled all the depressions and low areas between and around the two small uplands of Rigaud and Oka mountains.

On both sides of the Ottawa River, the middle and upper parts of the marine clay exhibit horizontal, burgundy-red beds alternating with grey beds. This phenomenon is limited to the Ottawa Valley and does not exist south of Rigaud Mountain and the St-Lazare marine deltaic sand plain. At one locality 2 miles southwest of Hudson, Vaudreuil County, fossil shells of the marine molluscs *Mya arenaria, Macoma balthica,* and *Mytilus edulis* were collected from an exposure (elev. 215 feet a. s. l.) of interbedded grey and red clay. This finding indicates that the interbedded grey and red clay was deposited in a marine environment.

North of Rivière-Beaudette and west and south of Hudson and St-Lazare, two large and in places thick bodies of sand were constructed during submergence of the area by the Champlain Sea. The sand is well sorted, medium to fine grained, and commonly shows well developed foreset and topset deltaic bedding. In the St-Lazare delta the pockets of massive sand are commonly highly fossiliferous and yield large quantities of shells of *Macoma balthica* and *Hiatella arctica.* One mile northwest of Rivière-Beaudette, fossil shells of *Mya truncata, Balanus hameri, Hiatella arctica,* and *Macoma balthica* were recovered from these sands at an elevation of 190 feet a. s. l. , the lowest elevation at which marine shells have been recovered from shoreline sands of the Champlain Sea in the Vaudreuil map-area. Two miles southwest of St-Lazare, fossil shells of *Hiatella arctica* and *Macoma balthica* were recovered from massive nearshore marine sands in the St-Lazare delta at an elevation of 230 feet a. s. l. ; this is the lowest elevation at which shoreline fossils of Champlain Sea have been recovered near the Ottawa River valley. Radiocarbon analysis of these shells will date the last stand of Champlain Sea in the Vaudreuil and Valleyfield area.

During submergence by Champlain Sea, several fossiliferous sand and gravel beach ridges were constructed around the higher parts of the emerging glacial topography. The highest fossiliferous beach was found at an elevation of 580 to 590 feet a. s. l. on

Rigaud Mountain two miles south of Rigaud; the submergence limit was determined to lie at approximately 630 to 640 feet a. s. l. on this same mountain. The highest part of the mountain appears to have escaped marine inundation and hence must have stood as an island in Champlain Sea. The lowest fossiliferous beaches were found at elevations of 230 feet a. s. l. two miles west of Hudson, south of Ottawa River, and one-half mile east of La Trappe on the north side of the river. The Hudson beach lies to the west of a marine sand that overlies clay; the La Trappe beach directly overlies bluish grey marine clay. Shells from the two sites will be submitted for radiocarbon dating.

Post-Champlain Sea Fluvial Sediments

Following the last stand of the sea a large amount of marine sediments, especially marine clay, was removed from a narrow belt on either side of Ottawa River below an elevation of 225 feet a. s. l. by postglacial fluvial erosion. The limits of this eroded belt in the northern part of the map-area are well marked by abandoned river bluffs cut in marine clay, the oldest and highest of which occur at an elevation of 210 to 220 feet. Other river-cut escarpments are present between those elevations and present river level, and remnants of the abandoned channel floors occur as terraces between these bluffs. In addition to cutting erosional terraces in Champlain Sea deposits, the degrading Ottawa River exposed the underlying glacial drift or bedrock (i. e., at Ile-Cadieux and west of Como). Downstream from glacial drift or bedrock exposures, the surfaces of erosional clay terraces commonly are mantled by a thin cover of alluvial sand derived from the exhumed material. The largest unfossiliferous alluvial sand terraces occur south of St-Eugène, at Hudson, and north, west, and east of Oka.

References

Richard, S. H.
1975: Surficial geology mapping: Ottawa Valley Lowlands (Parts of 31 G, B, F); *in* Report of Activities, Part B, Geol. Surv. Can., Paper 75-1B, p. 113-117.

Tremblay, J. J. L. and Hobson, G. D.
1962: Vaudreuil map-area, Quebec; Part I. Groundwater resources of the east half: Part II. The seismic method applied to a bedrock channel problem; Geol. Surv. Can., Paper 61-20, Map 30-1961, 18 p.

QUATERNARY GEOLOGY AND TERRAIN INVENTORY, KANANASKIS LAKES (82 J) EAST OF THE CONTINENTAL DIVIDE, ALBERTA

Project 740095

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Terrain Sciences Division

Field mapping and sample collection in the study area (Fig. 43.1) was completed during summer 1975. A total of 170 samples have now been collected during the combined field seasons of 1974 and 1975. These samples are being analyzed for such properties as specific gravity, texture, Atterberg limits, and clay mineralogy. The data obtained from these analyses, augmented with highway and foundation borehole data, will be combined with field mapping to produce a terrain inventory map. The units depicted on the map will be delineated according to the engineering and physical properties of the surficial deposits, the morphology of these deposits, and the physical processes acting upon them. This will allow the rating of each terrain unit according to its suitability for various types of land use.

In addition to the terrain mapping and attendant sample analysis, stratigraphic, geomorphic, tephrochronologic, radiocarbon, mineralogic, and botanical data and samples were gathered in order to clarify and interpret the Quaternary history of the area beyond that inferred from field data alone. This portion of the study will be synthesized in a map of the Quaternary deposits of the area that will depict the deposits as stratigraphic units as opposed to the terrain units previously discussed.

Preliminary interpretation of field data indicates the following chronology for the deposition of Quaternary deposits:

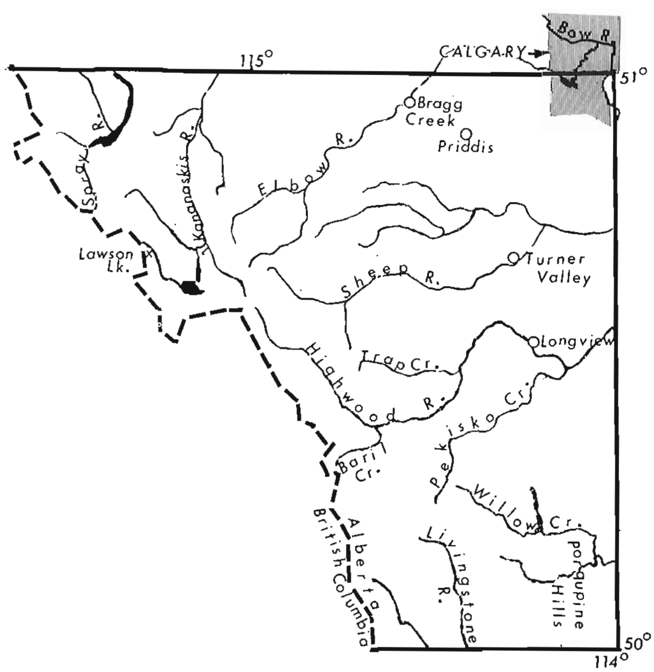

Figure 43.1. Kananaskis Lakes study area.

Fragmentary evidence for two or more glacial periods, which predate the draft sheet of the Erratics Train glaciation (discussed below), is found at several points within the study area. Granitic erratics derived from the Canadian Shield and scattered patches of till containing these erratics have been observed in this study and by past authors (Stalker, 1958; Day, 1971) to elevations in excess of 1650 m (5500 ft.) in the Porcupine Hills (Fig. 43.1), to 1455 m (4480 ft.) along the eastern margin of the Livingstone Range (Douglas, 1950, p. 6), and to 1410 m (4700 ft.) in the Sheep River basin (Jackson, 1975, p. 411). These erratics indicate deep penetration of Laurentide ice into the Foothills. Alley (1973) suggests, on the basis of his work in the Oldman and Crowsnest valleys south of the study area, that these erratics and isolated till occurrences represent two or more Laurentide advances. Evidence for a Cordilleran advance which predates the Erratics Train glaciation is found in a granite-free till underlying the Erratics Train drift sheet in the south Calgary area and in a granite-free till locally observed in the upper reaches of Willow Creek (Fig. 43.1).

The last extensive glaciation of the study area that involved coalescence of Cordilleran and Laurentide ice is here referred to as the Erratics Train glaciation after the Foothills erratics train described by Stalker (1956). This remarkable narrow band of thousands of quartzite erratics, which skirts the Foothills from west of Edmonton to south of the International Border, rests upon the last till sheet in the eastern one third of the study area. This till sheet is nearly devoid of granitic erratics, except in the eastern margin of the area and along Willow Creek valley where it cuts across the Porcupine Hills (Figs. 43.1 and 43.2). Drumlins, crag and tail features, and drumlinized bedrock ridges in the Bragg Creek and Pekisko Creek areas and grooves on ground moraine five miles east of Longview indicate southward flow of ice in the eastern one third (Foothills part) of the area during the Erratics Train glaciation. This apparent southward flow, the presence of the Foothills erratics (with their origin in the Jasper area [Mountjoy, 1958]), and the paucity of granitic erratics in the Foothills in the study area are best explained by progressive southward diversion of piedmont lobes of Cordilleran ice north of the area as well as those from the Bow, Elbow, Sheep, and Highwood valleys by Laurentide ice impinging on the Rocky Mountain Foothills from the north-northeast. Flow from the north-northeast by Laurentide ice synchronous with the transport of the Foothills erratics train has been documented by Stalker's (1957, 1973) mapping of flow-direction features east and northeast of the study area.

Following the maximum of the Erratics Train glaciation, the valley glaciers and the piedmont lobes

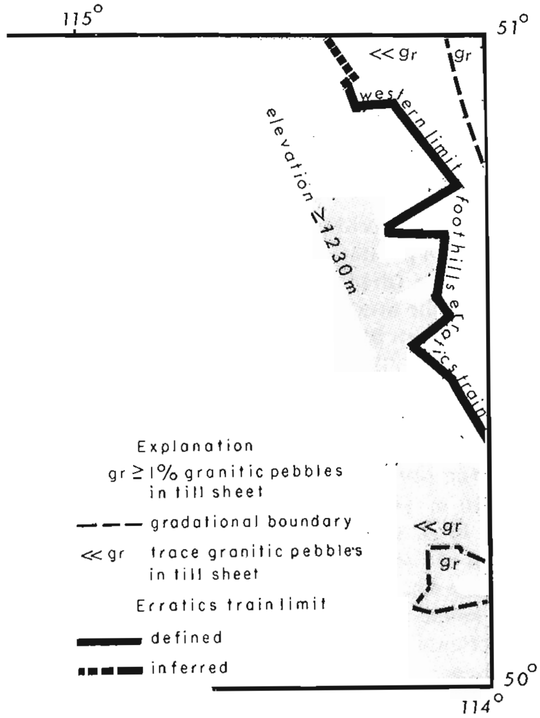

Figure 43. 2. Some features of part of the Erratics Train till sheet.

separated and a chain of ice-dammed lakes formed along the Foothills. Lake elevations up to and possibly in excess of 1350 m (4500 ft.) have been noted. The piedmont ice at this time was probably completely under the influence of the Laurentide Ice Sheet with which it had coalesced, and behaved as part of it. This supposition is indicated by the following observations:

(1) The ice-ponded lakes along the Foothills appear to have migrated progressively eastward and decreased in elevation. For example, a lake in the Bragg Creek area at about 1350 m (4500 ft.) was followed by a lower lake in the Priddis area to the east at about 1230 m (4100 ft.), which in turn was succeeded by a still lower lake in the Turner Valley area to the south and east (Fig. 43. 1).

(2) The Foothills erratics, although resting on Cordilleran ground moraine throughout most of the study area, are not found above 1230 m (4100 ft.) and extend up valleys in the easternmost Foothills (Fig. 43. 2). This pattern is best explained by their deposition along an easterly and northerly receding ice margin. In this case, the marginal ice was of western provenance but acted under the influence of Laurentide ice.

After the Erratics Train glaciation, Laurentide ice never again entered the study area. Two Cordilleran readvances have been recognized however on geomorphological evidence. The older is best shown in the

Table 43. 1

Tentative Correlations of the Quaternary Stratigraphy of the Kananaskis Lakes Study Area with Stratigraphic Units Recognized in Selected Studies of Adjacent Areas

This Study	Alley, 1973 (Crowsnest and Oldman valleys)	Rutter, 1972 (Bow Valley)	Tharin, 1960 (Calgary area)
Younger Cordilleran till (as recognized in the Baril Creek area)	Cache Creek till	Eisenhower Junction till	———
Older Cordilleran till (as recognized in the Trap Creek area)	not recognized	Canmore till	Morley till
Erratics Train glaciation till sheet (coalescence of Laurentide and Cordilleran ice)	Gap deposits and Beaver Creek till	Bow Valley till	Spy Hill till Balzac till Crossfield till
Pre-Erratics Train glaciation tills and erratics	Two Cordilleran and two Laurentide tills recognized	One older till recognized	Older till and stratified drift recognized

Trap Creek drainage basin (Fig. 43.1) where its moraines are well exposed. Evidence for this readvance is also seen in Kananaskis, Spray, Livingstone, Highwood, Sheep, and Elbow valleys. It reached the Foothills in several places. A later but smaller readvance, confined to valleys adjacent to the Continental Divide, is best seen in the Baril Creek area where its moraines are well exposed (Fig. 43.1).

At least two Neoglacial advances are recognized along the Continental Divide. Terminal moraines from these advances are found between 1980 m (6600 ft.) to over 2250 m (7500 ft.). The younger advance is the more common, with evidence of the older advance only observed at two locations in the study area. It is best seen in the Lawson Lake area (Fig. 43.1) of upper Kananaskis River. Dendrochronological data indicate this advance took place more than 500 years ago. Lichenometric data (G.D. Osborn, University of Calgary, pers. comm.) and dendrochronological data gathered on the younger advance in the Lawson Lake area indicate it to be between 100 and 150 years old.

Table 43.1 tentatively correlates the Erratics Train till sheet and the subsequent Cordilleran tills with glacial units recognized in other studies of adjacent areas. These correlations are based upon field-demonstrated continuity, in the case of the till sheet of the Erratics Train glaciation, and both continuity and/or similarity in magnitude in the case of later Cordilleran readvances.

No dates are available for any of the glaciations. A date of 10 600 ± 100 radiocarbon years B.P. (GSC-2162), obtained from shells collected 7 miles west of Longview, supplies a minimum age for retreat of ice from this part of the Highwood River valley.

References

Alley, N.F.
1973: Glacial stratigraphy and the limits of the Rocky Mountain and Laurentide Ice Sheets in southwestern Alberta, Canada; Bull. Can. Petrol. Geol., v. 21, no. 2, p. 153-177.

Day, D.L.
1971: The glacial geomorphology of the Trout Creek area, Porcupine Hills, Alberta; unpubl. M.Sc. thesis, Univ. of Calgary, Calgary, 158 p.

Douglas, R.J.W.
1950: Callum Creek, Langford Creek, and Gap map-areas, Alberta; Geol. Surv. Can., Mem. 255, 124 p.

Jackson, L.E.
1975: Quaternary geology and terrain inventory, Kananaskis Lakes, 82 J, east of the Continental Divide, Alberta; *in* Report of Activities, Part A, Geol. Surv. Can., Paper 75-1A, p. 112.

Mountjoy, E.M.
1958: The Jasper area – source of foothills erratics; J. Alta. Soc. Petrol. Geol., v. 6, p. 226-229.

Rutter, N.W.
1972: Geomorphology and multiple glaciation in the area of Banff, Alberta; Geol. Surv. Can., Bull. 206, 54 p.

Stalker, A.M.
1956: The Erratics Train, Foothills of Alberta; Geol. Surv. Can., Bull. 37, 31 p.

1957: Surficial geology, High River, Alberta; Geol. Surv. Can., Map 14-1957.

1958: Surficial geology, Fort Macleod, Alberta; Geol. Surv. Can., Map 21-1958.

1973: Surficial geology of the Drumheller area, Alberta; Geol. Surv. Can., Mem. 370, 122 p.

Tharin, J.C.
1960: Glacial geology of the Calgary, Alberta area; unpubl. Ph.D. thesis, Univ. Illinois, 131 p.

FRASER DELTA SEDIMENTATION, VANCOUVER, BRITISH COLUMBIA

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A total of 642 samples collected over the tidal flats and slope of the Fraser Delta during the first series of sediment sampling operations (Luternauer, 1975a, b) have now been grain size analyzed. During the 1975 field season 46 one-metre cores were collected from the tidal flats. These will be used to study changes in the lithologic and geochemical character of the delta front (the latter investigation is the subject of a research contract awarded the Department of Geological Sciences at the University of British Columbia on September 3, 1975). During the summer, work has also proceeded on research agreements granted the Department of Civil Engineering at the University of British Columbia to assess Fraser Delta slope stability and to the British Columbia Institute of Technology to photogrammetrically examine morphologic changes on the tidal flats. As the Canadian Coast Guard Hovercraft, which had served as a field vehicle during 1974 sampling periods, was not available this past summer, seasonal patterns of sedimentation that were observed during the first time series study of the flats could not be verified as originally had been planned.

Figure 44.1, which describes the broad morphologic character of the delta front, contains information derived from a tidal flat airphoto survey performed by Lockwood Survey Corp. Ltd. (now Pacific Survey Corp.) in 1967 and from a 1974 Canadian Hydrographic Service bathymetric survey of the delta slope. A preliminary compilation of sedimentologic data acquired to date is displayed in Figures 44.2 to 44.6.

Echo-sounding records (Fig. 44.7) indicate that in general the slope of the delta is free of relief where it is not cut by major canyons. Two morphologically complex sites, however, are apparent off the mouth of the main channel and along southernmost Roberts Bank. The first is characterized by what appears to be a dense network of fairly regular, shallow canyons (Fig. 44.7, profiles D and E). It is important to establish by what mechanism(s) they were formed, to what extent they influence lateral and downslope sediment movement, and ultimately how they may affect local slope stability. The second region, off the superport, has a relief which becomes increasingly irregular from west to east (Fig. 44.7, profiles H, I, and J). As echo-sounding tracklines here are oriented along the dip of the slope (unlike the obliquely oriented profiles D and E), it is more reasonable to attribute this relief to mass wasting and not to canyon erosion.

The rate at which the distributary channels replenish tidal flat sand (Fig. 44.2) that is gradually swept into deeper water is not known. The seaward inflection of the 40 per cent sand contour however is a measure of the relative contribution of sand from Main Channel, North Arm, and Middle Arm to the slope.

The implications of the anomalously high sand concentrations on the southern Roberts Bank slope must be carefully examined as this part of the delta front is adjacent to the segment of tidal flat for which major development has been proposed. Is the anomalous sedimentologic character of this portion of the slope attributable to a different Pleistocene history, the influence of the local physical oceanographic climate, man's disruption of the natural flow patterns of the river, or a combination of these factors? A preliminary comparison of a 1974 prefreshet and postfreshet bathymetric survey (Luternauer, 1975b) suggests that portions of this southern slope are retreating. This is supported by the suspected presence of the aforementioned slump features (Fig. 44.7) and by indications in bottom photographs (Fig. 44.8) that the sea floor in this area is eroding.

The postfreshet sand distribution (Fig. 44.3) reveals (compare the location of the 40 per cent contour in Figs. 44.2 and 44.3) that freshet-related sedimentation tends to decrease the grain size of upper slope sediments off Main Channel but appears not to alter the sedimentary character of the southernmost slope.

The distinct difference in the sedimentologic regimes off Sturgeon and Roberts banks is clearly represented by the variation in clay distribution across the delta front (Fig. 44.4). The paucity of clay on the southern slope sharply contrasts with the high clay content of sediments just beyond the edge of the Sturgeon sand banks. The newly acquired information on the dispersal direction and accumulation sites of the finer Fraser River suspended matter can be directly applied to the interpretation of metallic ion pollutant levels in local sediments (Grieve and Fletcher, 1975). It also can be used in site selection studies for any future local waste discharge and disposal facilities.

Information compiled in Figure 44.5 summarizes the changes in modal size in what is primarily the bedload component of the tidal flat sediments. The degree of variation in modal size over the year probably is related to fluctuations in local depositional/erosional regimes. Along the edge of Sturgeon Bank the modal size shifts only slightly, if at all, through the year. This suggests that fine sediment supply is sufficiently low and/or wave and current energy is high enough that a dynamic equilibrium is maintained. Of particular concern here is the effect that further man-induced alteration of sand flow will have on the stability of the leading edge of the tidal flats. Variation in modal size along the channels crossing the centre of the bank suggests that sediment-charged freshet flows are capable of significantly altering the sedimentary character in this zone. The modal variability noted just north of the seaward end of the north jetty of Main

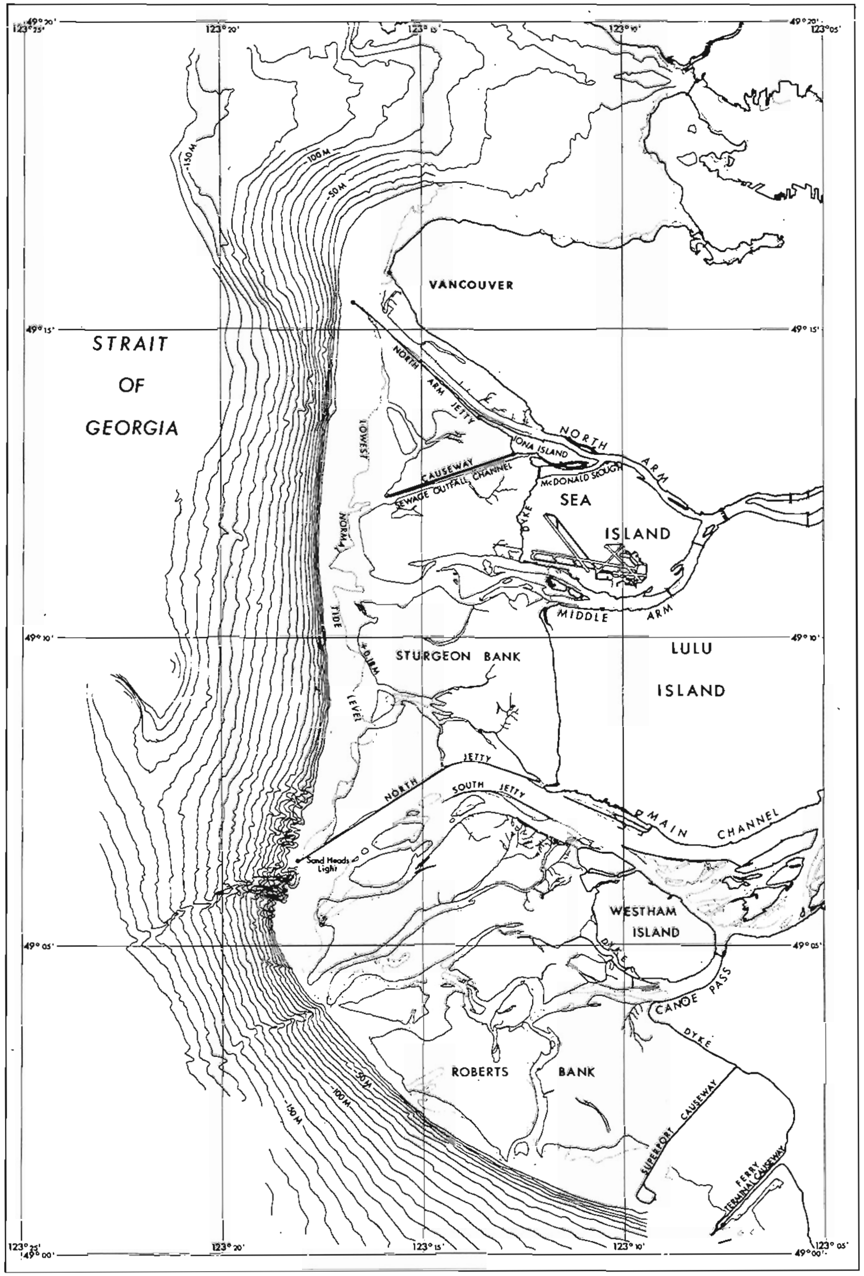


Figure 44. 1. Morphology of and geographic locations at the Fraser Delta front. Compiled from a 1967 photogrammetric survey of the tidal flats (Lockwood Survey Corp. Ltd.) and a 1974 bathymetric survey of the slope (Canadian Hydrographic Service). Contour interval is 10 m (except for 0. 18 m contour on flats).

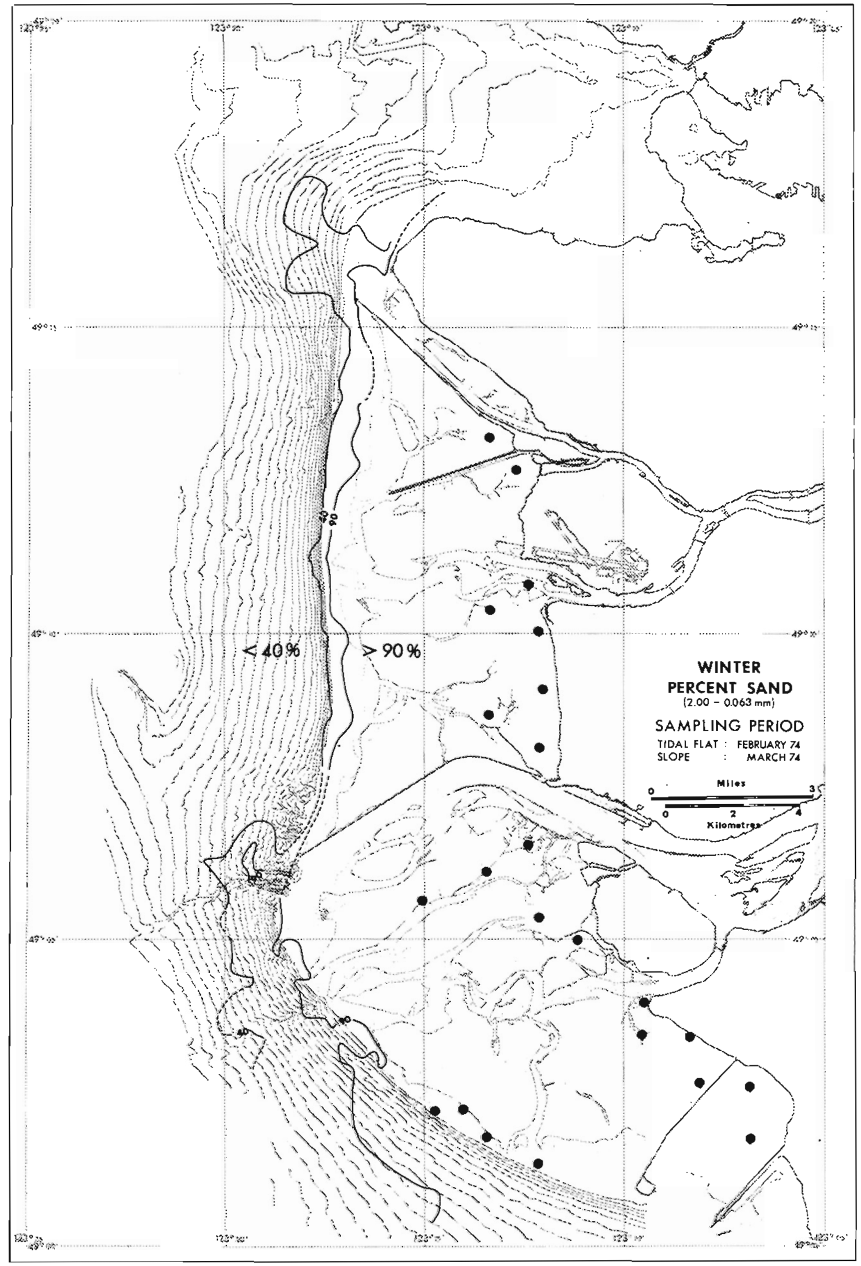


Figure 44. 2. Prefreshet sand distribution reflecting adjustment to the winter oceanographic/river discharge climate. Black dots represent sampling sites on the banks with less than 90 per cent sand. The sampling sites at the edge of Roberts Bank all have in excess of 50 per cent sand as do almost half of the inshore stations.

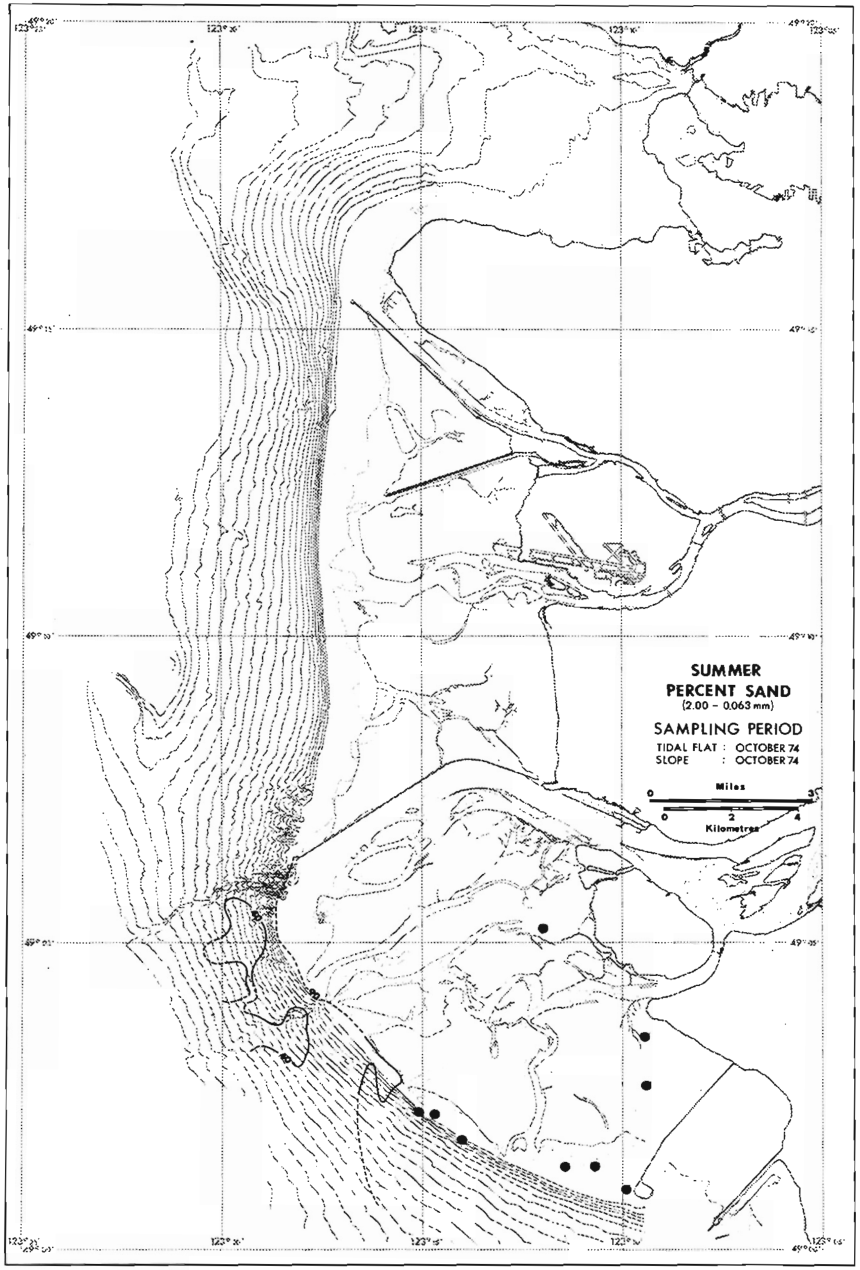


Figure 44. 3. Postfreshet sand distribution along Roberts Bank reflecting the influence of high river runoff. Black dots represent sampling sites on the bank with less than 90 per cent sand. All except one of the inshore sites has in excess of 50 per cent sand.

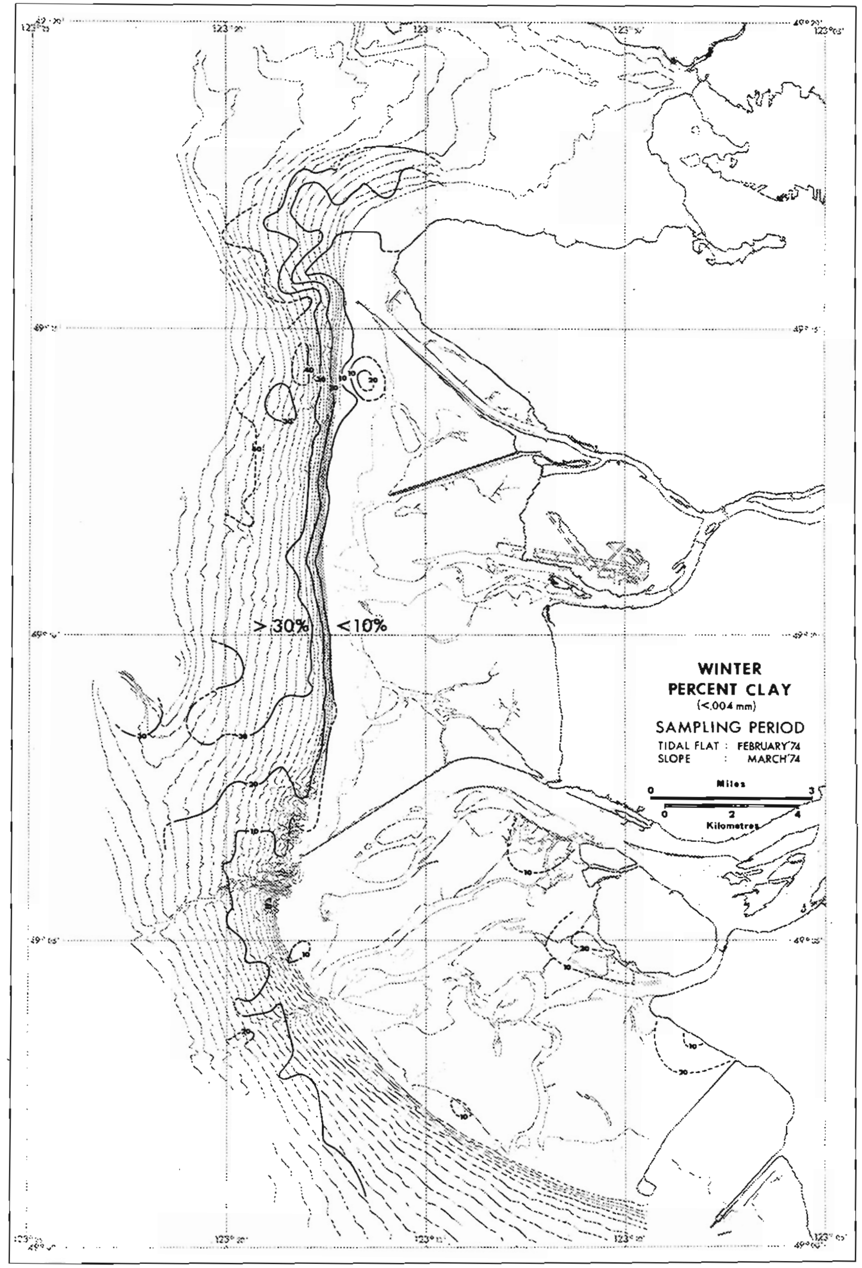


Figure 44. 4. Prefreshet clay distribution reflecting adjustment to winter oceanographic/river discharge climate.

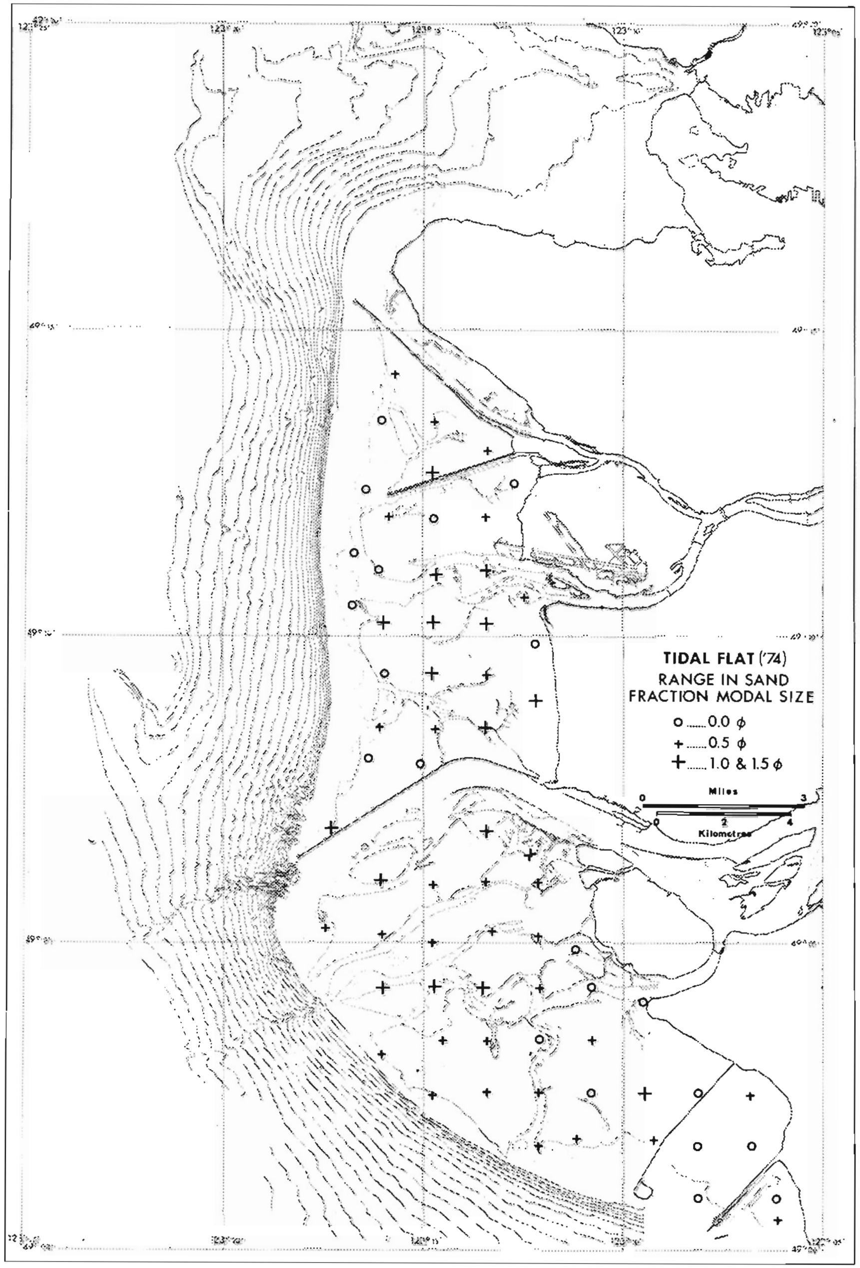


Figure 44. 5. Phi shift in modal size of sand fraction of tidal flat sediments during the 1974 sampling period. All samples analyzed at half-phi sieve intervals.

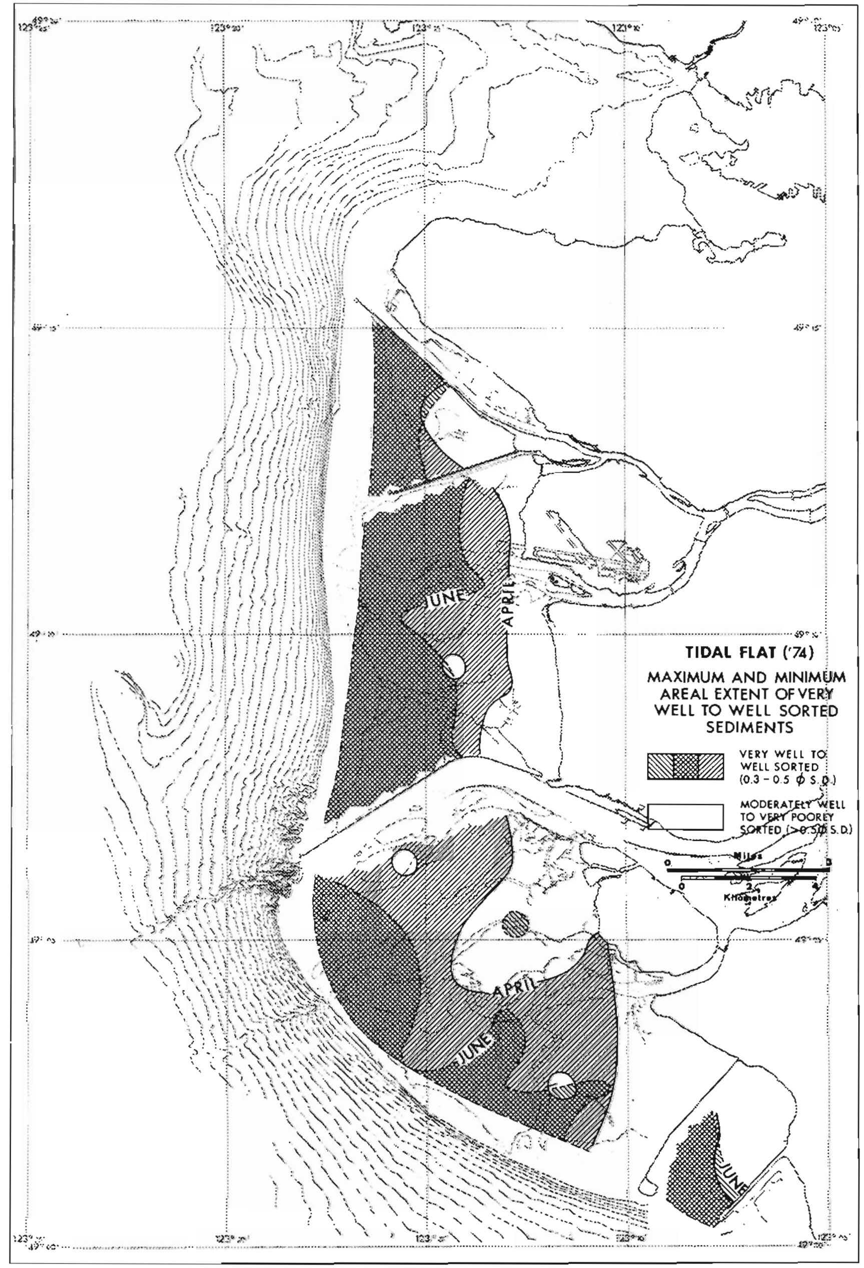


Figure 44. 6. Prefreshet and postfreshet (1974) variation in areal extent of very well to well sorted tidal flat sediments.

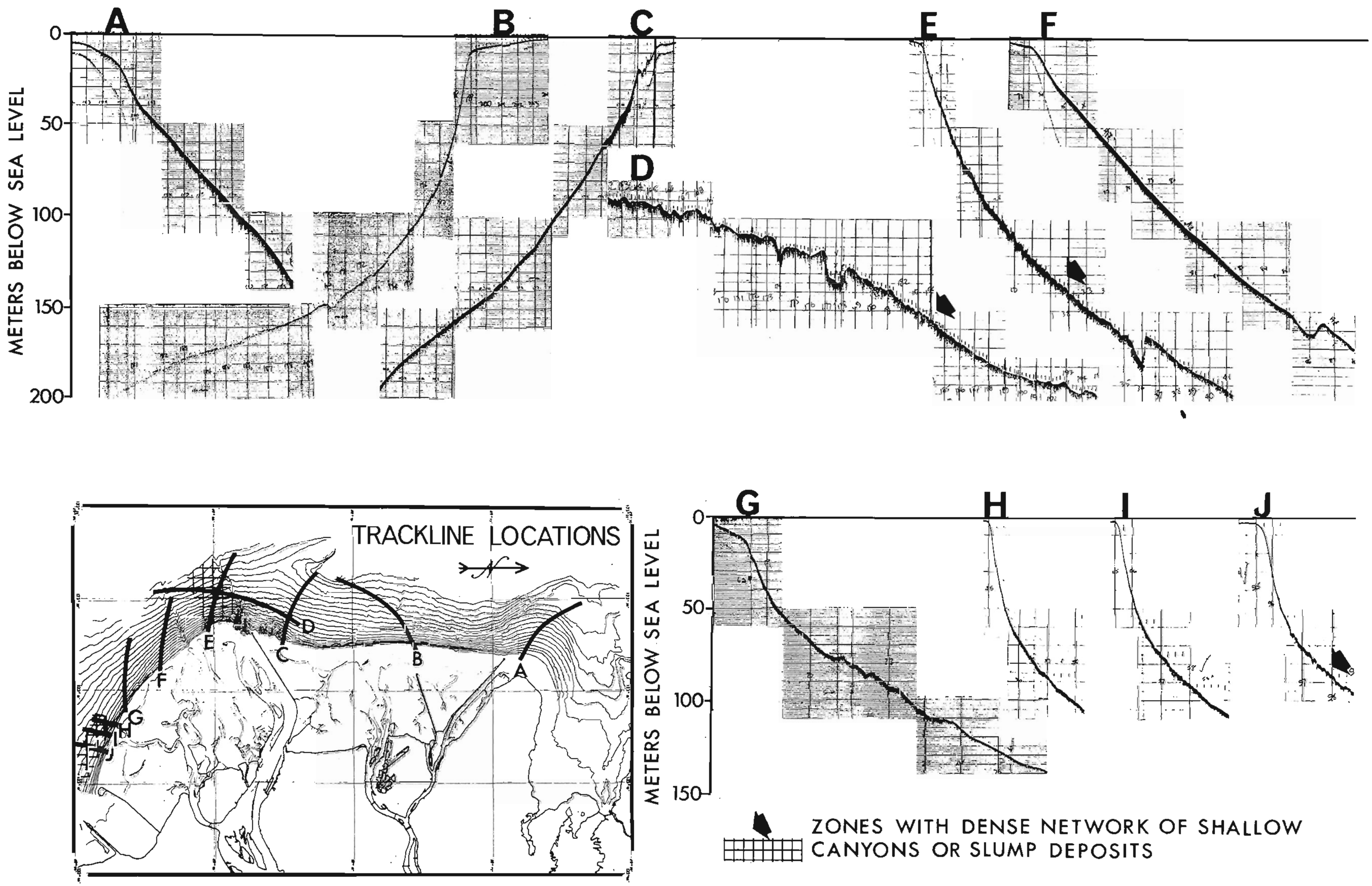
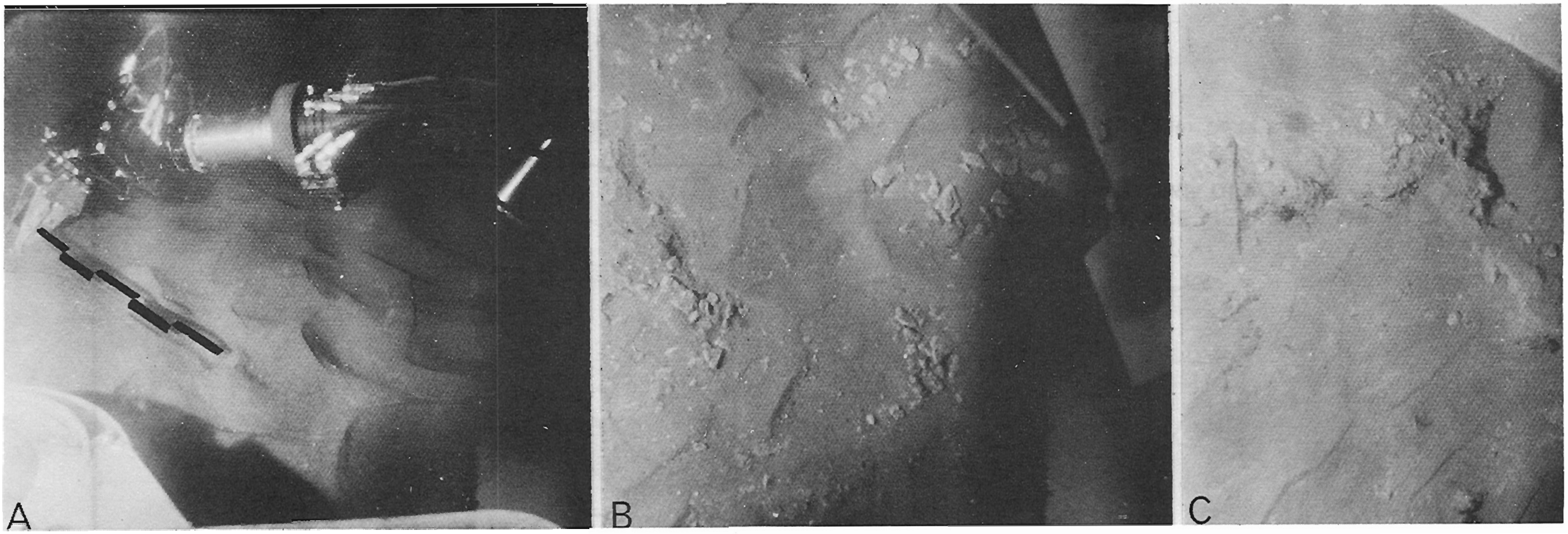


Figure 44. 7. Selected echo-sounding records from a 1974 Canadian Hydrographic Service survey of the Fraser Delta slope. Variable vertical exaggeration.



A. Rippled sand; scale marked off in 10 cm intervals.

B. Indurated (?) clay chips scattered on the sandy seafloor; photo at approximately twice the scale as (A).

C. A resistant ledge forming a small outcrop on the sloped sandy seafloor; photo at approximately twice the scale as (A).

Figure 44. 8. Underwater photos obtained from a Pisces submersible on March 22, 1975 at 75 m depth at a site on the delta slope seaward of Roberts Bank superport.

Figure 44. 9

Edge of marsh lying between the superport and ferry terminal causeways (Fig. 44. 1).

Channel was observed near a break in the otherwise continuous jetty. Along Roberts Bank the greatest variation in modal size is observed just south of the less confined southern margin of Main Channel and along the tidal channels extending west of Canoe Pass. These channels disgorge their sediment at the head of the second largest submarine canyon on the delta slope and not along that portion of the Roberts Bank slope which, as was previously noted, may be retreating.

It is apparent in Figure 44.6 that freshet discharge tends to compress the zone of well sorted sediments (although it must be noted that the areal extent of well sorted sediments was less in February than in April). The fact that this compression is more marked on Roberts Bank than on Sturgeon Bank suggests that the freshet influence is greater on the former platform. As man's alterations to the river system have tended to restrict sand flow to the channels, it is probable that the freshet contribution to Roberts Bank primarily has promoted the progradation of the high tide sedge grass zones and not the sand banks themselves.

The extent to which the superport causeway has contributed to the poor sorting of the sediments northwest of it cannot be defined at present because virtually nothing is known of the precauseway sediment distribution. Nevertheless, the causeway probably has contributed to the apparent retreat of the marsh lying between the superport and ferry terminal causeways (Figs. 44.1 and 44.9) by blocking and/or deflecting away from this vital biologic environment the suspended sediment which otherwise would have migrated into the area.

Acknowledgments

All grain size analyses were performed by Heather Mowat at the Terrain Sciences Laboratory in Vancouver. Preliminary data compilation was done by Wendy Symonds.

References

Grieve, D. and Fletcher, K.
1975: Trace metals in Fraser Delta sediments; *in* Report of Activities, Part B, Geol. Surv. Can., Paper 75-1B, p. 161-163.

Luternauer, J. L.
1975a: Fraser Delta sedimentation, Vancouver, B.C.; *in* Report of Activities, Part A, Geol. Surv. Can., Paper 75-1A, p. 467-468.

1975b: Fraser Delta sedimention, Vancouver, B.C.; *in* Report of Activities, Part B, Geol. Surv. Can., Paper 75-1B, p. 171-172.

HYDROLOGIC RECONNAISSANCE, THOMSEN RIVER BASIN, BANKS ISLAND, DISTRICT OF FRANKLIN

Project 740079

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As a part of a reconnaissance study of surface processes on Banks Island, undertaken by Day (1975), a program of basic hydrologic data collection was carried out in the Thomsen River basin of central Banks Island during summer 1975. The field season commenced on June 3 with the establishment of a fly camp on Thomsen River ("Thomsen Fly", 73°14'N, 119°32'W) in the area of special interest. From this location it was possible

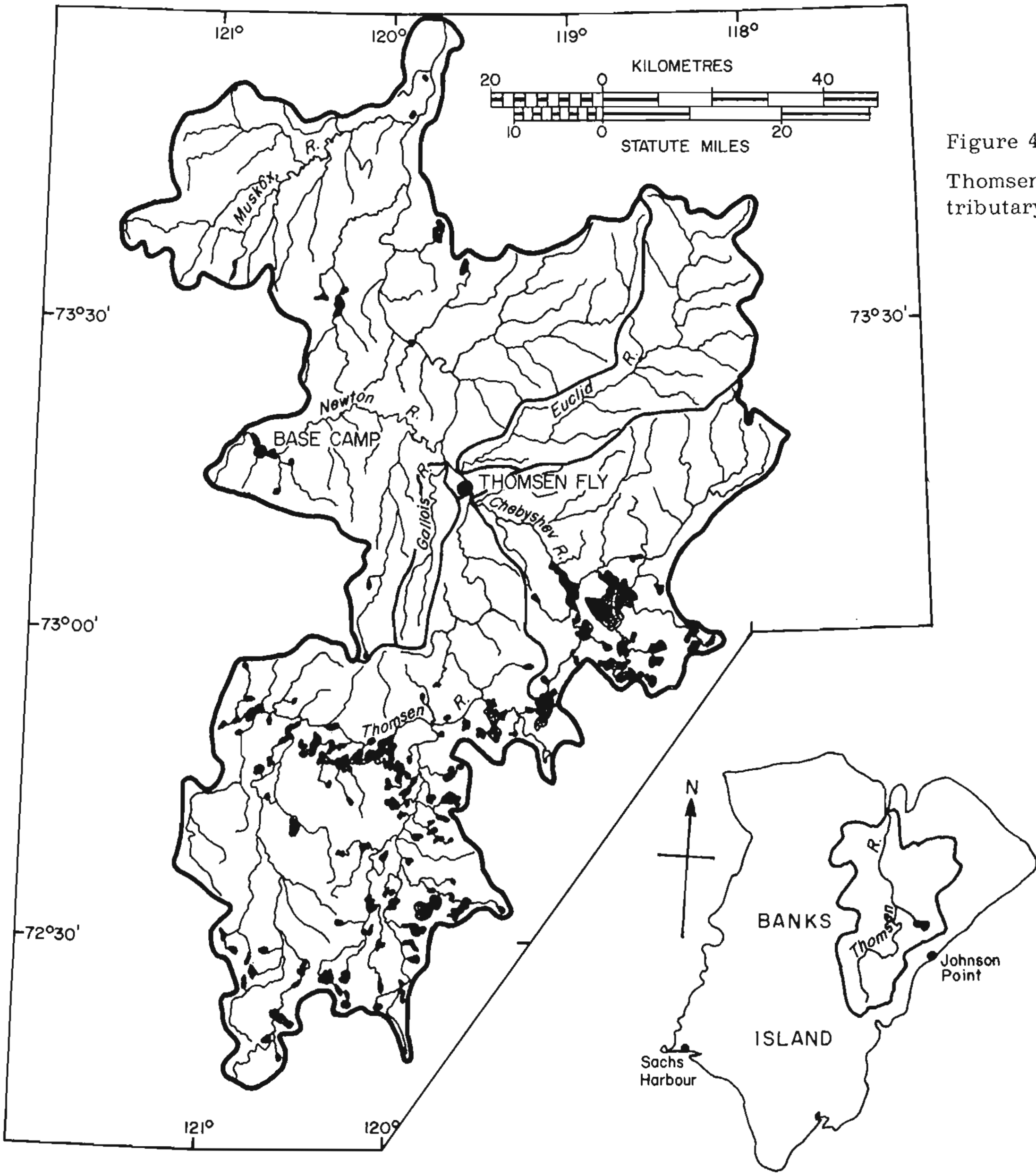

Figure 45. 1

Thomsen River basin, showing tributary basins studied in 1975.

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Table 45.1

Some Elementary Drainage Basin Statistics

Drainage Basin	Area (km^2)	Lake cover (%)	Approximate maximum relief (m)	Peak discharge (m^3/sec)	Peak unit area discharge ($m^3/sec/km^2$)
Thomsen above "Thomsen Fly"	5280	7.3	300	426	0.081
Thomsen above "Chebyshev"	3610	7.6	300	326E	0.090E
"Chebyshev"	1670	6.9	275	100E	0.060E
"Euclid"	1060	0.2	350	119	0.112
"Gallois"	221	1.9	150	69	0.313

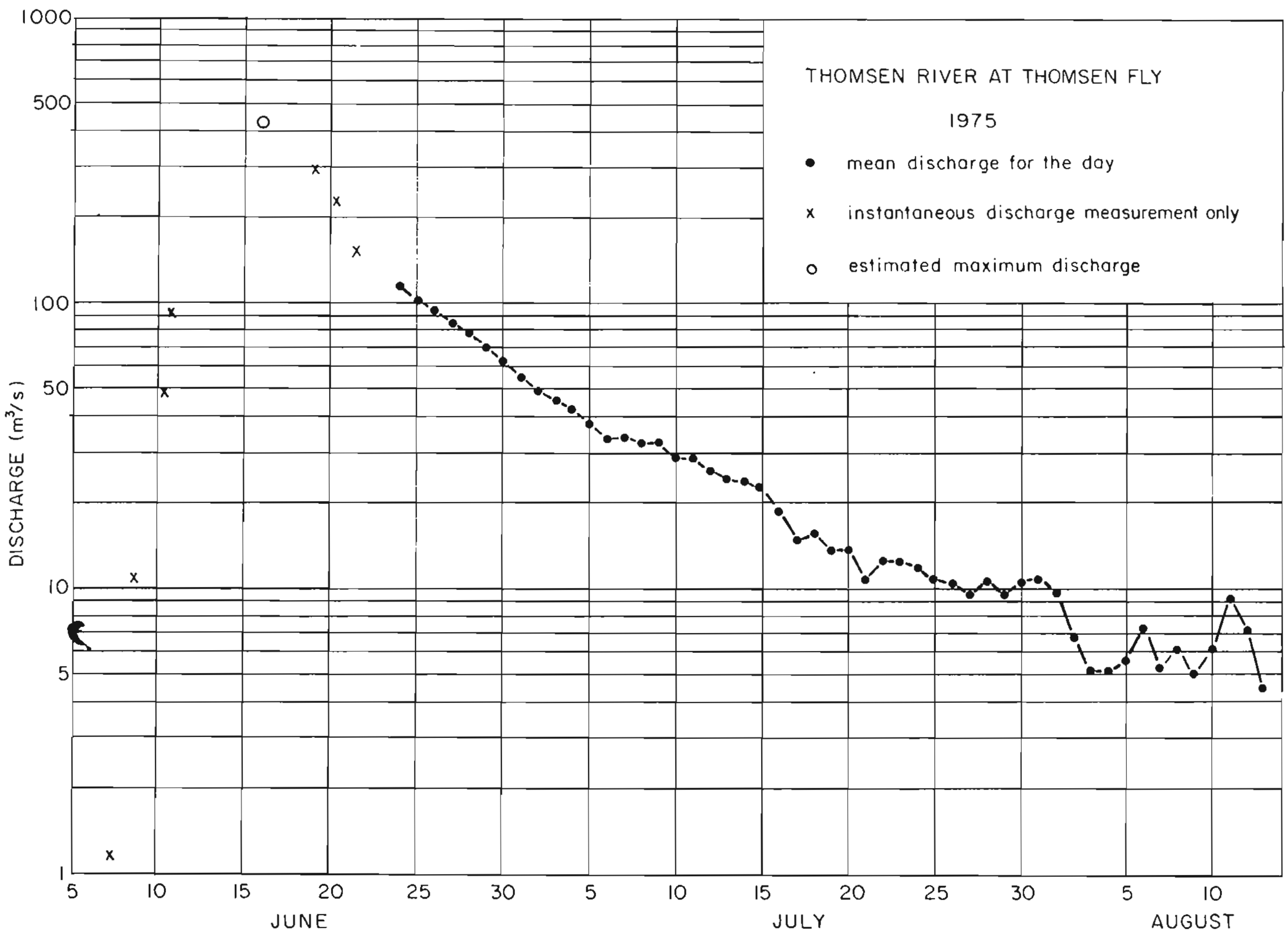

Figure 45.2. Hydrograph for Thomsen River basin above "Thomsen Fly", 1975.

to study the Thomsen River as well as three of its tributaries, which enter the Thomsen within a radius of 7.5 km (Fig. 45.1).

Snow and Ice

Prior to the work at "Thomsen Fly", observations were made of snow distribution in the area between Johnson Point and the base camp (73°17'20"N, 120°42'00"W). As expected, snow had been redistributed by the wind during the winter, such that most of it lay in large snowbanks in the lee of hills, in river valleys, and in gullies. Hilltops were swept clean of snow in areas of greater relief.

The snowpack ripened in late May when daily maximum temperatures began to rise consistently above 0°C. Owing to the extreme nonuniformity of snow distribution and the limited time available, it was not possible to obtain a value of the average snowpack water equivalent. The results of a snow course performed on "Elf Lake" near the base camp are given as a general indication only. That course yielded a mean water equivalent of 90 mm with a mean density of 0.43. The snow depth varied from 89 mm to 406 mm, with a mean value of 208 mm.

Before the commencement of spring runoff, observations were made of ice and snow conditions in the major river valleys in the vicinity of "Thomsen Fly". A major feature in all of the valleys were snowdrifts lying adjacent to valley walls, particularly where the walls were steep faced. A maximum snow depth of at least two metres in those drifts was not uncommon.

Along Thomsen River major pools of the channel were ice filled. Where the depth of the pool permitted, the ice could attain a thickness of up to 2.25 m. During early snowmelt, the initial runoff of Thomsen River flowed over this ice. In early June as the width of flowing water increased to cover most of the ice, the ice rotted, became free of the bottom, and rose in the current to float downstream in pans and irregular chunks. At "Thomsen Fly", the ice surfaced in large pans from a deep pool. These pans floated only a short distance downstream to the next riffle where they became grounded and formed a partial ice jam in mid-channel. The flow was diverted around both sides of this jam, and the situation persisted from late in the evening of June 10 until at least late in the morning of June 11.

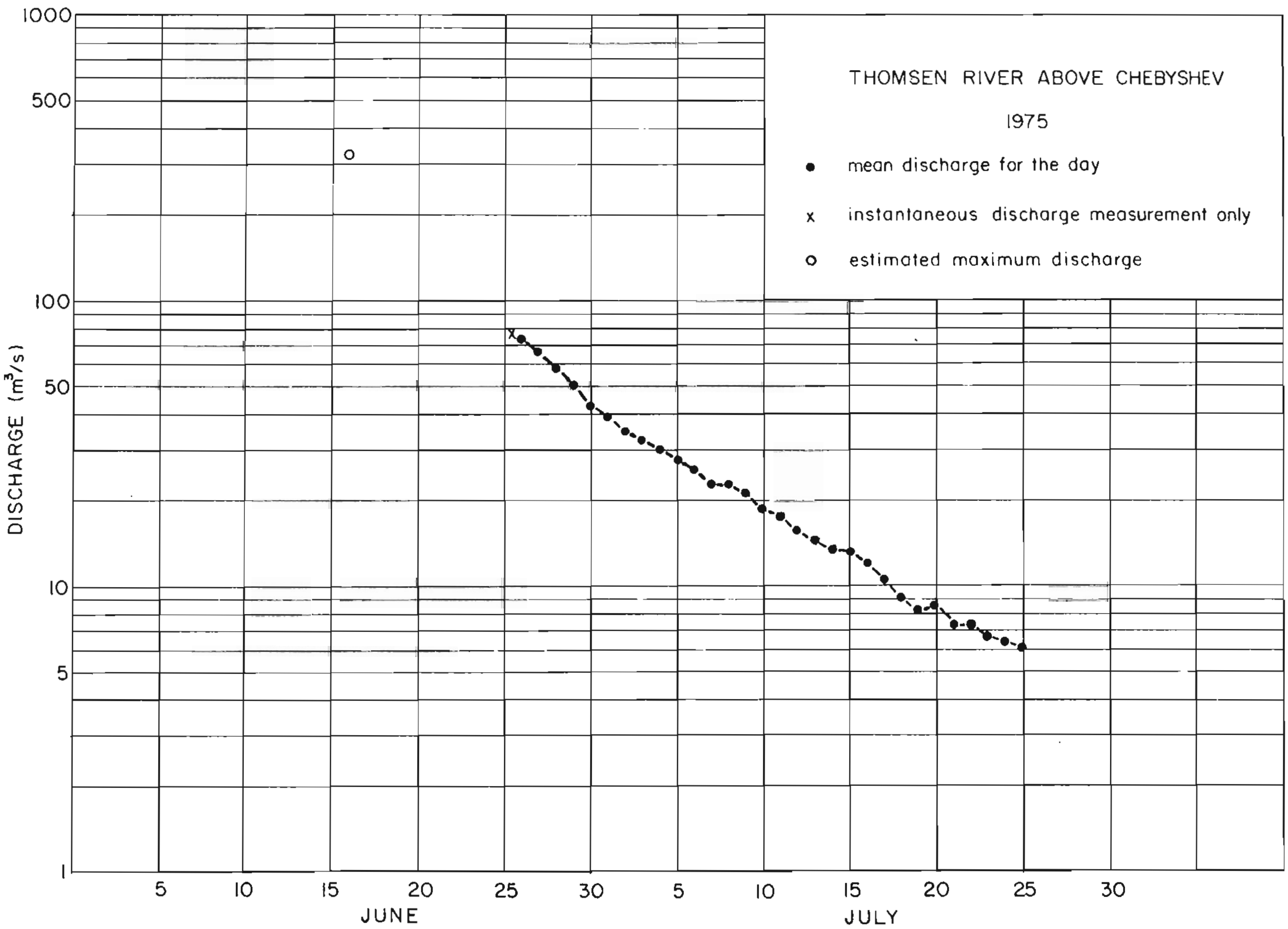

Figure 45.3. Hydrograph for Thomsen River basin above the "Chebyshev", 1975.

Discharge and Runoff

Runoff commenced during the first week of June and attained a peak value for the open-water season on June 16. A rapid rise to peak flow occurred during the afternoon and evening of June 15 in response to a sudden increase in air temperature. At "Thomsen Fly", a high of 15°C was recorded on June 15, far above the maximum of 6.5°C recorded in the previous two weeks.

During the week before the flood peak and for three or four days afterwards, there was a considerable amount of floating ice observed along Thomsen River. The amounts in the tributaries studied were minimal by comparison.

As water levels rose through early June, rivers widened in their channels to undercut the snowbanks that lay in the valleys adjacent to their banks. As the flood receded, these snowbanks were left to overhang the channel with no support from beneath by the water. Tension cracks developed and large blocks of snow calved into the water. In this manner, snowbank depletion was accelerated and river flow was augmented.

Peak flow occurred at a time when only five to ten per cent of each basin still had any snow cover. Most of this snow cover, however, lay in the valleys in close proximity to river channels or in hollows on steep slopes where the gradients encouraged rapid runoff. Consequently, it was still possible to have a significant response in runoff when the snow was melting rapidly.

Stilling wells with Stevens type F water-level recorders were used to measure river stage. These installations proved inadequate for use on rivers of the size studied. The recorders had to be removed during peak flow owing to large fluctuations in water levels as well as to the hazards imposed by floating ice. The maximum discharge of the rivers had to be estimated by a survey of high-water marks and extrapolation of stage-discharge relationships derived during the flood recession. These estimates are given in Table 45.1.

Note that the estimates of peak discharge are tentative at the time of writing, especially for "Chebyshev River" and for Thomsen River above "Chebyshev". In the case of "Chebyshev", a water-level recorder was not reinstalled following the spring flood. Instead, flow was derived as the difference in discharge on Thomsen River below and above the point where the "Chebyshev" enters. In estimating the flood maximum of the "Chebyshev", the assumption has been made that "Chebyshev" and Thomsen rivers crested at approximately the same time. In the case of Thomsen River

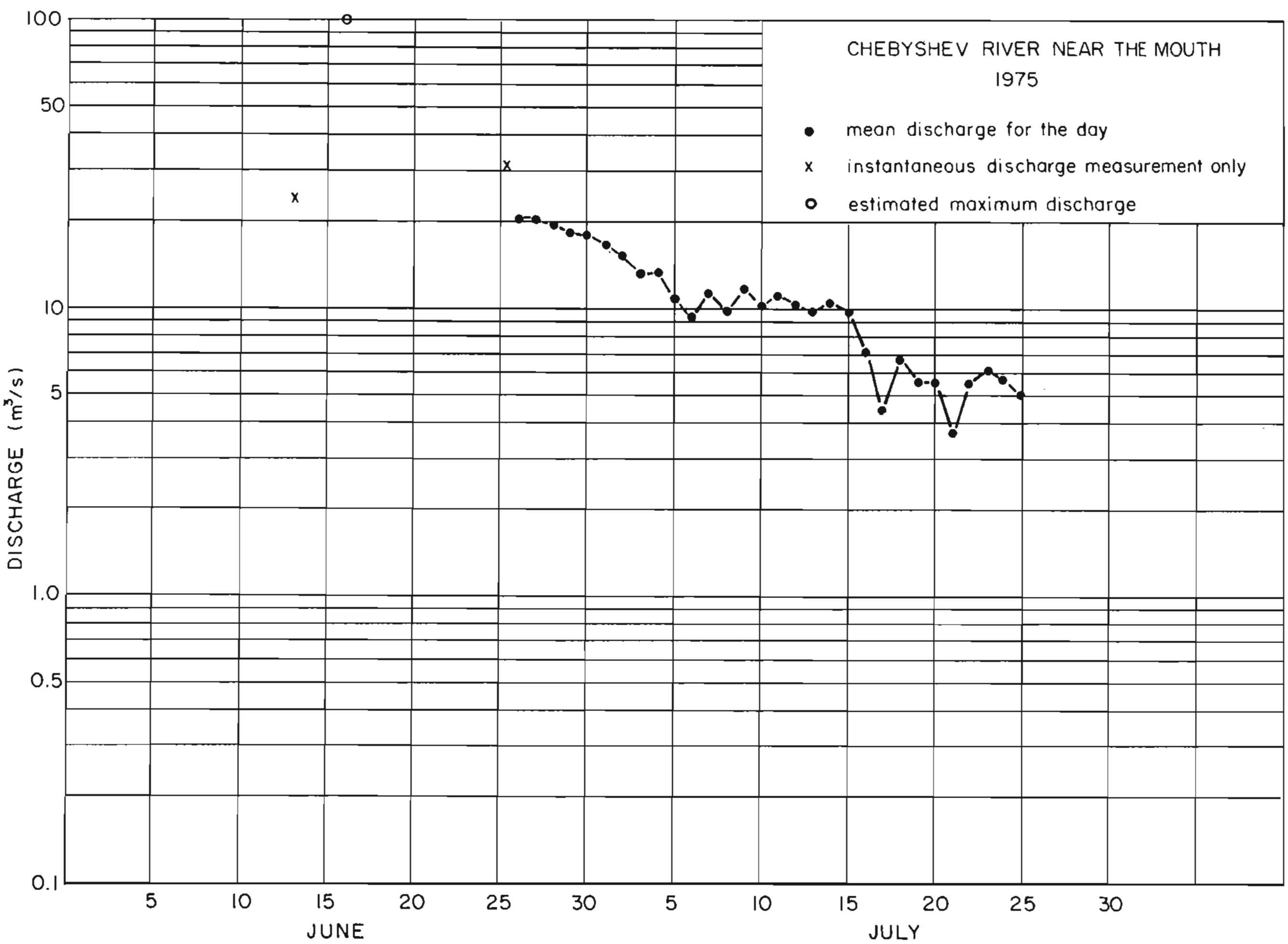

Figure 45.4. Hydrograph for the "Chebyshev River" basin, near the mouth, 1975.

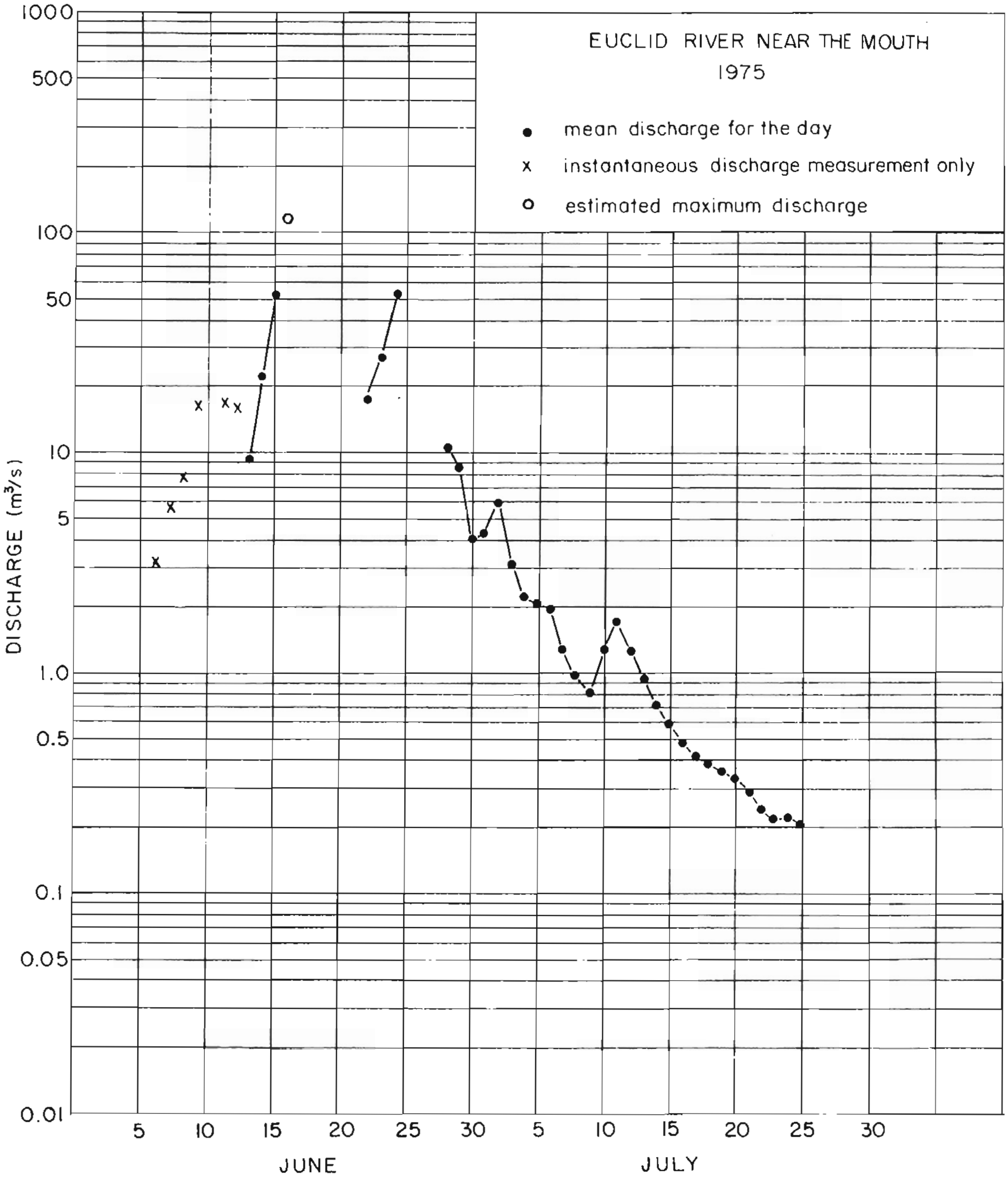

Figure 45.5

Hydrograph for "Euclid River" basin, near the mouth, 1975.

above "Chebyshev", extrapolation of the stage-discharge relationship yields an estimated peak flow of 376 m^3/sec. No discharge measurements however are available before June 25 to help verify this. Consequently, a more conservative estimate has been adopted.

Hydrographs for each river are presented in Figures 45.2 to 45.6. The snowmelt flood dominated in every case. A prolonged recession followed for the remainder of each record. By late July discharge was low, especially for "Gallois River", which had a flow of almost zero when the water-level recorder was removed.

Suspended Sediment and Water Chemistry

Suspended sediment samples are still undergoing preliminary analysis. Highest sediment concentrations apparently occurred in "Euclid" River during peak flow and again in mid-July, as well as in "Chebyshev" River during mid-July.

Hardness and pH were monitored during July. The trend in hardness was an increase over time, from a range of 61 ppm to 87 ppm as calcium carbonate early in the month to a range of 91 ppm to 157 ppm three weeks later. "Gallois" River had the highest concentrations, followed by the "Euclid", the "Chebyshev", and the Thomsen above the "Chebyshev".

In conjunction with this trend, pH increased in all cases except for the "Chebyshev", which varied only slightly within the range of 8.75 to 8.90. The other three rivers exhibited a rise within the range of 8.55 to 9.00.

Meteorologic Data

A weather station was set up at "Thomsen Fly" on June 5, 1975. Precipitation was measured using an Atmospheric Environment Service tipping-bucket rain gauge. Total precipitation (rain plus snow) recorded from June 5 to August 13 was 28 mm (this omits many trace events, as well as dewfall and frost). The normal total for June and July at Sachs Harbour has been 26 mm. On July 10 the heaviest precipitation event at "Thomsen Fly" produced only 4.3 mm of rain. Although snow fell during each summer month, it came mostly in flurries, and amounts were not commonly significant.

Temperature was recorded with a Lambrecht weekly thermohydrograph. Means for June 5 to 30, July 1 to 31, and August 1 to 12 were 5.7°C, 5.1°C, and 5.0°C, respectively. Means for Sachs Harbour in June, July, and August have been 2.2°C, 5.6°C, and 4.3°C. The highest temperature recorded was 19°C on June 25; the lowest was -2°C on June 21. Generally, the first half of June was cloudy and cool and the second half was sunny and considerably warmer. July was predominantly cloudy and cold.

Wind speed and direction data at a height of 1 m were obtained using a Lambrecht monthly wind recorder. The record shows a tendency for wind direction to remain consistent for several days at a time. The predominant directions were as follows: June 9 to 27, east through south; June 28 to July 26, northwest and north; July 27 to August 4, west and northwest. Extended periods of very light wind were rare. Conditions of actual calm are estimated to have happened for only one per cent of the time from June 9 to August 12. The highest wind speeds were recorded around mid-day June 24 with average hourly speeds of almost 40 km/h.

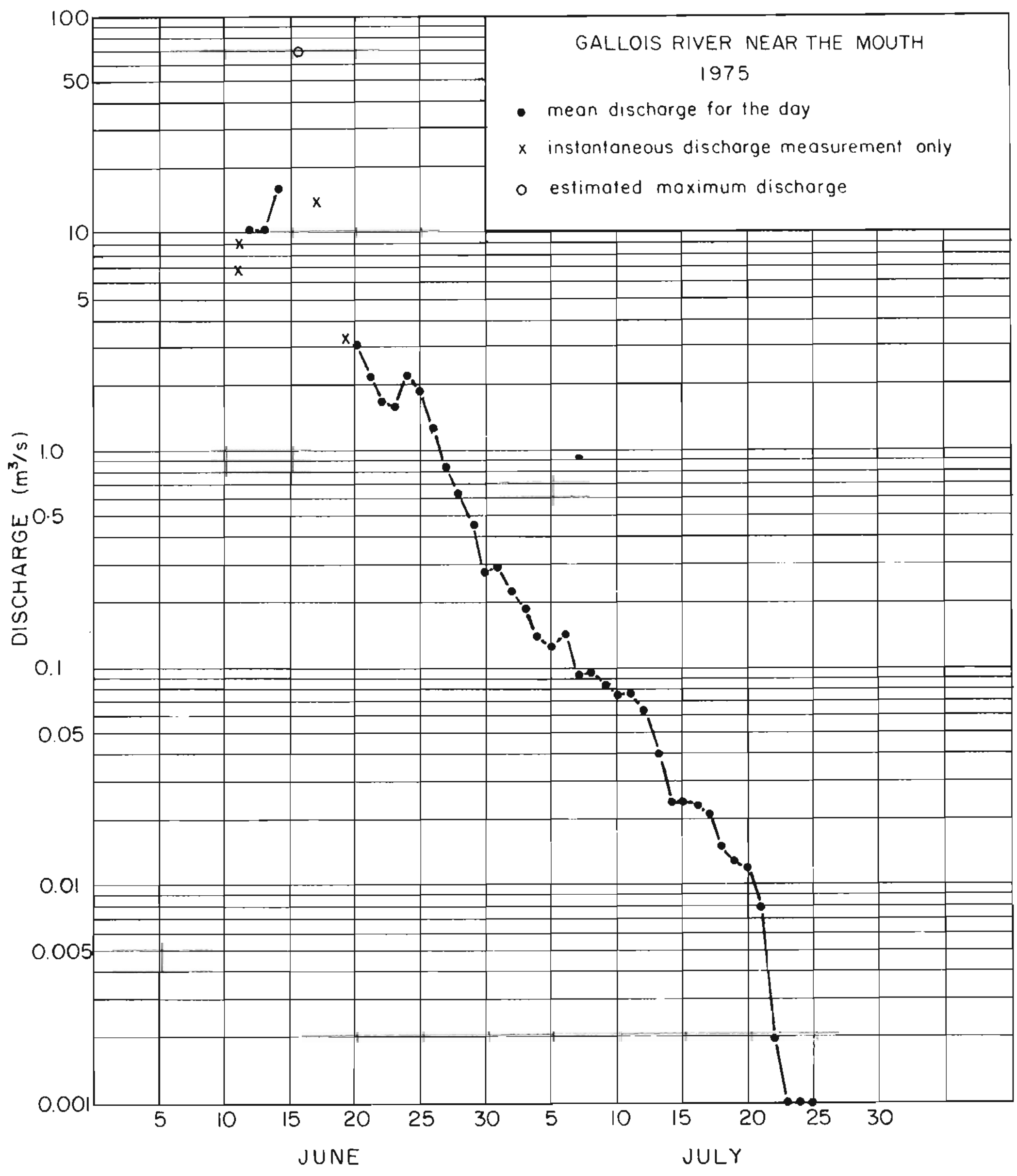

Figure 45.6

Hydrograph for "Gallois River" basin, near the mouth, 1975.

Acknowledgments

We wish to express our gratitude to the following individuals and organizations who assisted us with our needs in the field in terms of equipment and/or advice: from the Department of Energy, Mines and Resources, Ottawa, Dr. T.J. Day (Terrain Sciences Division), and from the Department of the Environment, the following people: Mr. J.N. Jasper (Glaciology Division, Ottawa); Mr. P. Engel (Head, Hydraulics Research Division, Burlington); Mr. W. Stichling (Head, Sediment Surveys Section, Ottawa); Mr. R.A. Halliday (Head), Mr. R.A. Terzi (Special Services and Surveys Section, Ottawa), and Mr. T. Chapman (Special Services and Surveys Section, Calgary); and from Water Survey Division, Mr. D.A. Davis (Regina), Mr. J. Dwyer (Calgary), and Mr. H.L. Wood (Inuvik). Special mention must go to the staff of Polar Continental Shelf Project, Tuktoyaktuk for logistical support provided throughout the summer field season.

Reference

Day, T.J.
1975: Reconnaissance studies of surface processes, Banks Island, District of Franklin; *in* Report of Activities, Part A, Geol. Surv. Can., Paper 75-1A, p. 457.

HYDRAULIC, MORPHOLOGIC, AND MORPHOMETRIC STUDIES OF SELECTED RIVERS ALONG THE ROUTE OF THE MACKENZIE HIGHWAY

Project No. 750077

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Terrain Sciences Division

Introduction

In response to requests by the Mackenzie Highway Environmental Working Group for environmental studies along the Mackenzie Highway, a fluvial geomorphic study was initiated with two main objectives: 1) to study the hydraulic and morphologic behaviour of selected rivers over which the Mackenzie Highway crosses with a view to (a) assessing the accuracy of design estimates and (b) evaluating any modifications to the channel regime and form caused by construction; and 2) to initiate long-term studies into river processes and form.

Site Description

Three rivers whose beds are unstable were chosen for intensive study: Little Smith Creek (highway mile 521. 1 – km 838. 7), Steep Creek (highway mile 511. 9 – km 823. 8), and Saline River (highway mile 521. 1 – km 838. 7). These rivers drain from the Franklin Mountains (McConnell Range), an area of highly folded Devonian limestones and shales (Fig. 46. 1).

Little Smith Creek is a fourth order stream (all ordering after Horton, 1945) with a drainage area of 476. 6 km^2 of moderate slope (1: 307). It has many gravel bars and crossovers. The majority of the basin is covered by a till plain, although considerable portions of the upper tributaries are dominated by organic bogs. The lower valley is wide with high, steep, silt and sand sides which are subject to thermal erosion. The bed, in the vicinity of the highway crossing is 75 per cent boulder and 25 per cent sand; however, the proportion of sand increases with proximity to the mouth.

Steep Creek is a fourth-order stream that drains a 147. 6 km^2 area; as the name suggests, the stream gradient is high (1: 75). Ninety per cent of the bed material is in the boulder-size range. A relatively high discharge, in comparison to the other rivers, was maintained throughout the summer months suggesting that a relatively large proportion of the stream discharge originates as groundwater flow from the Franklin Mountains.

Saline River is a fifth-order stream of intermediate gradient (1: 141). It drains 303. 0 km^2 of land, the majority of which is a till plain. The till is of variable thickness and in the upper tributaries forms a thin veneer 1 to 2 m thick. The bed ranges from 95 per cent boulder and gravel in the vicinity of the highway crossing to 80 per cent boulder and gravel near the mouth. Gully and thermal erosion occurs along portions of the steep north bank, and undercutting is common on both banks. The meandering stream contains several vegetated point bars. Evidence of higher flow conditions include the presence of several abandoned channels and multi-levelled alluvial terraces.

Field Methods and Results

Cross-sectional stream profiles were surveyed on each of the three streams and were resurveyed periodically throughout the summer to monitor the

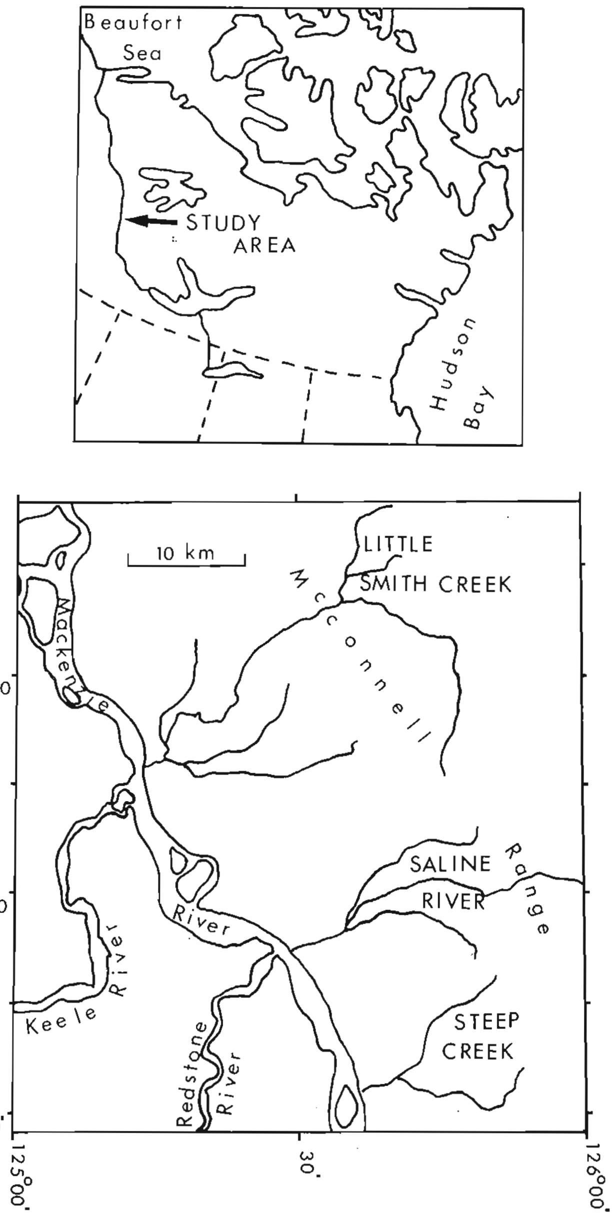

Figure 46. 1. The location of the study area.

Table 46. 1. Sample data collected from monitored streams

Profile	Date	Discharge m^3/s	Mean Height of Bed Below Datum m	Standard Deviation m	Variance m^2	Mean Stream Width m	Mean Stream Depth m	Mean Stream Velocity m/s	Mean Size Bed Material m	Standard Deviation m	Variance m^2
SA-MP5	July 15	0. 43	1. 010	0. 128	0. 054	3. 41	0. 253	0. 51	0. 028	0. 025	0. 002
SA-MP5	Sept 10	0. 50	1. 080	0. 106	0. 037	2. 44	0. 168	1. 22	0. 056	0. 052	0. 003
SA-MP5	Sept 13	1. 24	0. 865	0. 260	0. 223	11. 30	0. 143	0. 77	---	---	---
SA-MP4	July 15	0. 43	0. 829	0. 064	0. 013	9. 30	0. 105	0. 44	0. 049	0. 036	0. 001
SA-MP4	Sept 10	0. 50	0. 832	0. 063	0. 013	9. 14	0. 102	0. 53	0. 056	0. 044	0. 002
SA-MP4	Sept 13	1. 24	0. 802	0. 088	0. 026	11. 58	0. 165	0. 65	0. 039	0. 037	0. 001
SA-P1A	July 15	0. 43	0. 410	0. 158	0. 082	20. 42	0. 232	0. 10	0. 072	0. 114	0. 013
SA-P1A	Sept 10	0. 54	0. 386	0. 158	0. 083	18. 29	0. 226	0. 13	0. 059	0. 095	0. 009
SA-P1A	Sept 13	0. 98	0. 360	0. 176	0. 102	18. 90	0. 268	0. 19	0. 091	0. 091	0. 008
ST-MP1	July 18	0. 91	1. 30	0. 092	0. 027	7. 89	0. 147	0. 80	0. 116	0. 081	0. 007
ST-MP1	Sept 6	0. 80	1. 47	0. 108	0. 038	7. 89	0. 151	0. 68	0. 139	0. 130	0. 017
ST-MP1	Sept 8	0. 95	1. 48	0. 109	0. 039	7. 89	0. 188	0. 64	0. 150	0. 114	0. 013
ST-P6	July 16	0. 91	*0. 408/0. 152	0. 102/0. 103	0. 014/0. 011	5. 48/11. 52	0. 16/0. 08	0. 48	0. 063	0. 038	0. 001
ST-P6	Sept 6	0. 80	0. 399/0. 104	0. 111/0. 101	0. 012/0. 010	5. 48/11. 89	0. 13/0. 07	0. 50	0. 079	0. 052	0. 003
ST-P6	Sept 8	0. 95	0. 408/0. 101	0. 091/0. 095	0. 027/0. 009	5. 42/11. 89	0. 14/0. 06	0. 64	0. 075	0. 062	0. 004

* Two values are given as the stream occupied two channels.

N. B. All data were taken within the confines of the streams as they existed on the day of data collection.

change in bed configuration associated with changes in discharge. Bed material was sampled at each profile by a random walk method (after Wolman, 1954) while those portions of the bed that became exposed due to a drop in water level were sampled by sieving. Four kg samples generally were collected; however when the b axis of the bed material was greater than 7. 6 cm, 45 kg samples were taken.

The thalweg of each of the streams was mapped periodically to detect spatial and temporal shifts in the long profile of the streams. Lateral and vertical control was carried throughout these surveys. In addition, portions of each of the streams were mapped to provide quantitative information on the rate of bank sapping and undercutting.

Sample data showing the net change in the cross-sectional profiles and bed material over the field season are presented in Figure 46. 2 and Table 46. 1. Storm events in the study area this year were rare and of low intensity. However, 18 mm of rain fell in a 24-hour period commencing on September 12. A partial coverage of this event was possible and sample data are presented in Figure 46. 2 and Table 46. 1.

Mackenzie River greatly influences the hydraulic behaviour of the streams under study. The Mackenzie breaks up in the spring after its tributaries and, as a result, ice piles up in the mouths of the tributaries damming flow and causing a backup. Field evidence indicates that ice push is active 9. 9 m, 9. 4 m, and 9. 5 m above the September 10, 1975 water levels on Saline River, Little Smith Creek, and Steep Creek, respectively. In consideration of the average gradients presented earlier, it is readily appreciated that water backup to these heights would alter the flow characteristics of the streams for a considerable distance upstream from their mouths.

Water-level fluctuation on the Mackenzie River throughout the year is significant in determining flow conditions in the tributary streams. Upvalley precipitation events in late June 1975 caused the Mackenzie to rise 1. 3 m over a four-day period and brought about the deposition of silts and sands in the monitored mouth areas. Subsequently the Mackenzie water level declined 3. 8 m by September 10. This reduction of base level led to considerable scour in the mouth areas of the streams monitored (eg. SA-MP5, Fig. 46. 2).

In view of the importance of the Mackenzie River in determining the hydraulic characteristics of the streams under study, a depth-time curve was established for the Mackenzie River which will be used in conjunction with similar data from Norman Wells and Wrigley to produce data on flood frequency. Dendrochronology also is being used for this purpose.

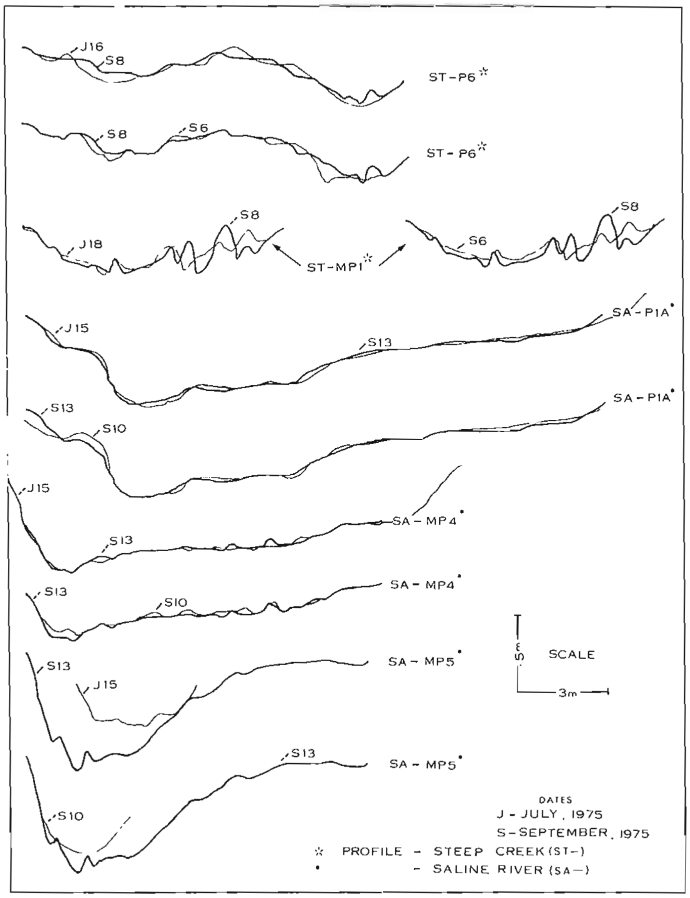

Figure 46. 2. Stream cross-section variation through time.

Long-Term Studies

To facilitate long-term studies of stream processes and form, a total of forty bench marks were emplaced along the monitored streams. The bench marks consist of 1. 5-cm-diameter, copper-clad rod driven into the ground by means of a portable 25. 8 kg jack-hammer to a maximum depth of 6. 7 m. The bench marks established this year will provide the horizontal and vertical control necessary for future monitoring studies.

References

Horton, R. E.
1945: Erosional development of streams and their drainage basins: hydrophysical approach to quantitative morphology; Geol. Soc. Am., Bull. 56, p. 274-370.

Wolman, M. G.
1945: A method of sampling coarse river bed material; Am. Geophys. Union Trans., v. 35, no. 6, p. 951-956.

ICE-WEDGES AS INDICATORS OF RECENT CLIMATIC CHANGE, WESTERN ARCTIC COAST

Project 680047

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Ice wedges are extremely abundant in many areas with continuous permafrost and unconsolidated materials. Ice wedges grow by thermal cracking in winter and infilling of the cracks by meltwater or hoar crystals in the spring. The lower portion of a veinlet, which lies within permafrost, then becomes preserved. Large ice wedges grow by repeated cracking at the same place. As ice wedges can only be preserved in permafrost, it follows that permafrost degradation will thaw the top of an ice wedge and, conversely, permafrost aggradation will cause a secondary wedge to grow if veins open along the same place. Such permafrost degradation and aggradation, with a thickening and thinning of the active layer, should follow a climatic warming and cooling trend. It is the purpose of this note to show that secondary and tertiary ice wedges have recently grown along the western Arctic coast in response, most probably, to the post-1950 cooling trend. In so far as is known, the secondary and tertiary wedges are not the result of vegetation change, the growth of peat, or burial by soil movement.

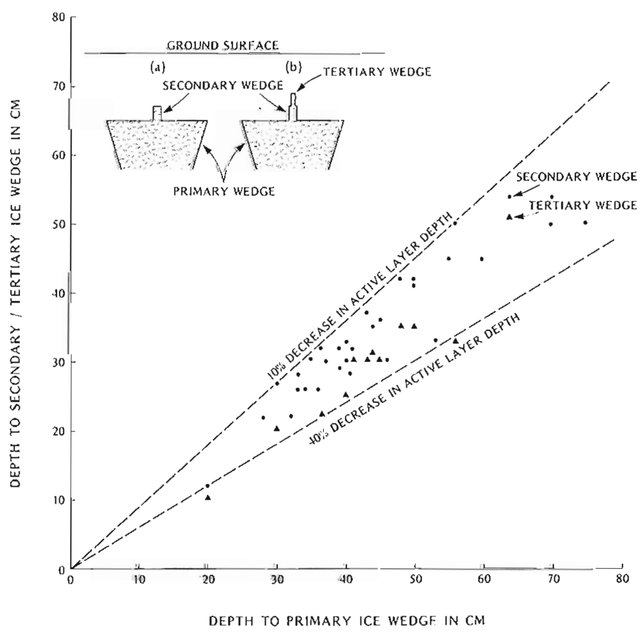

Figure 47.1. Graph showing the depth to the top of the primary wedge plotted against the depth to the top of the secondary wedge, and also the tertiary, if present.

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Field Evidence

During the 1975 field season, a systematic effort was made to examine the tops of all available ice wedges in order to see if the growth patterns reflected recent climatic changes (Burns, 1973; Mackay, 1975). About 125 ice-wedge exposures were examined at Garry Island, Hooper Island, Pelly Island, Richards Island, and along the coast 5 to 20 km southwest of Tuktoyaktuk, N.W.T. About 25 of the ice wedges were inactive, having been buried by soil creep or the accumulation of peat. Of the remaining 100 wedges, about 30 (Fig. 47.1) were active with small secondary or even tertiary ice wedges. The secondary and tertiary wedges typically were located in the centres of the wedges into which they penetrated (Fig. 47.2). Fresh cracks were not seen at the sides of the old wedges, thus showing that cracking tends to recur along the same vertical plane (i.e. at the rejuvenated wedge site). The annual veinlets averaged from 0.5 to 1.5 cm in width. The secondary wedges averaged 2 to 5 cm in width; the tertiary wedges were somewhat narrower than the secondary wedges. Although counting of the veinlets was most difficult, probably most of the secondary wedges had 5 to 10 annual growth veinlets, and the tertiary wedges had about 2 to 4.

Discussion

The case for a post-1950 climatic cooling trend is well documented for many areas. Burns (1973) has shown that in the lower Mackenzie Valley, January air temperatures decreased about 4°C from 1950 to 1970 and July temperatures were little changed. Thus mean annual air temperatures probably declined about 2°C. Ground temperatures would be expected to parallel, in a general way, those of air temperatures although other factors (e.g. snow cover) play a major role. In any event since present mean annual ground temperatures along the western Arctic coast are in the -8°C to -10°C range, a 2°C cooling trend should result in active layer thinning and upward permafrost aggradation. The rejuvenated ice wedges provide field evidence for a period of recent climatic cooling, a thinning of the active layer, and upward permafrost aggradation. As many of the secondary and tertiary wedges were split by fresh 1975 ice veinlets, the wedges obviously are growing. Ice wedges without secondary wedges appeared to be, with very few exceptions, inactive, because single ice veinlets extending downwards into the wedge tops could rarely be found.

Conclusion

Many active ice wedges along the western Arctic coast have been rejuvenated in response to a thinning

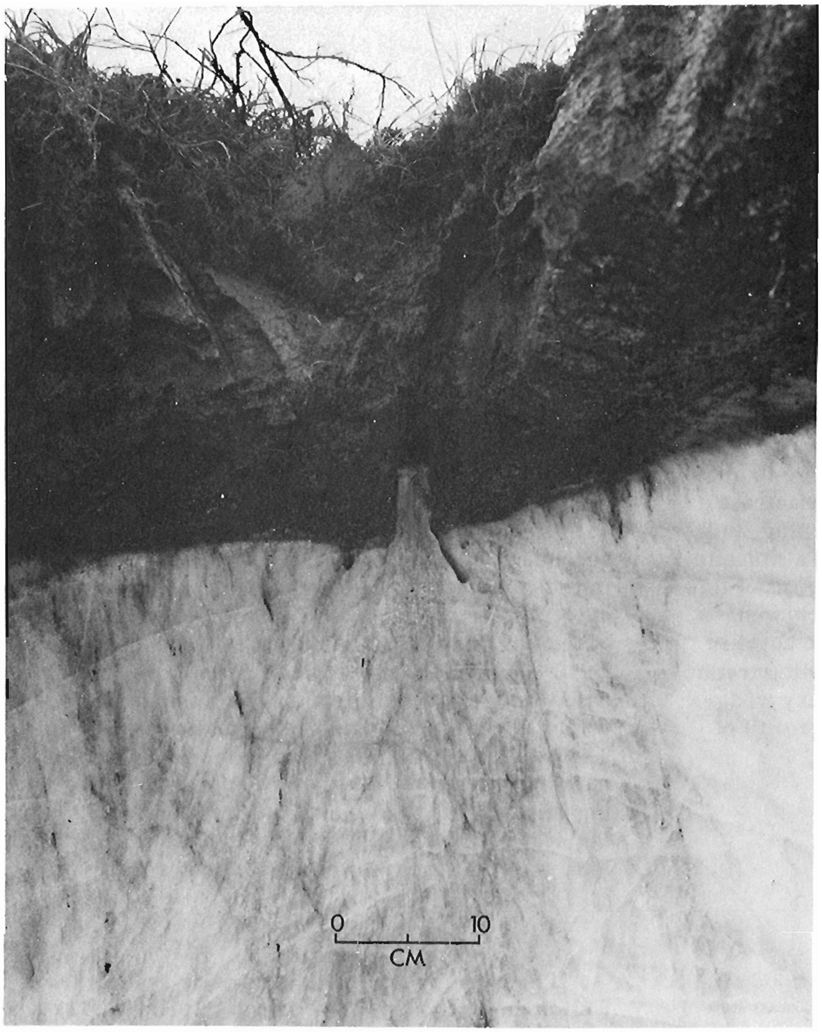

Figure 47. 2.

Secondary ice wedge, Pelly Island, N. W. T. ; a small tertiary wedge and a 1975 winter vein were present, but do not show in the photograph.

of the active layer and upward permafrost aggradation. When the general cooling trend was interrupted by warm periods, the tops of the new ice wedges were truncated, so that subsequent cooling and cracking produced tertiary wedges. The evidence from active ice wedges along the western Arctic coast suggests a thinning of the active layer of about 10 per cent to a maximum of 40 per cent in the recent past.

References

Burns, B. M.
1973: The climate of the Mackenzie Valley-Beaufort Sea, Vol. 1; Environ. Canada Climatol. Stud., no. 24, Toronto, 227 p.

Mackay, J. R.
1975: The stability of permafrost and recent climatic change in the Mackenzie Valley, N. W. T. ; *in* Report of Activities, Part B, Geol. Surv. Can., Paper 75-1B, p. 173-176.

PINGO INVESTIGATIONS, BANKS ISLAND, DISTRICT OF FRANKLIN

Project 640004

H.M. French[1]
Terrain Sciences Division

Field work was a continuation of studies initiated in 1974 in co-operation with A. Pissart (French, 1975; Pissart, 1975). In addition to compiling a pingo distribution map of the island, attention was concentrated upon groups of small pingo-like mounds, and various pingo-like ridges. The objectives were to establish the existence of ice cores within these features and, if successful, to remove ice samples for crystallographic and chemical analyses. Sections were cut using a portable Wajax fire pump.

Central Bernard River (73°23'N, 123°15'W)

Two weeks were spent examining a group of shallow pingo-like mounds on a low terrace of Bernard River. They had first been identified from the air in 1974. Thirty-one of these features were mapped in the field, and all were located within an area of approximately 1.5 km^2. The forms possess little vertical relief, the highest rising only 3.1 m above the level of the terrace. The majority are circular, less than 2.0 m high, and between 20 and 100 m in diameter (Fig. 48.1). They contain shallow central depressions which vary in width and depth; sometimes, the mound consists only of a circular rampart (Fig. 48.2). All are easily recognizable on the lowland tundra terrain because of the absence of vegetation on the ramparts. A few features are elongate or complex in form, the latter resulting from the coalescence of several adjacent mounds. One of the mounds is a true linear feature, being 166 m in length and only 20 to 30 m in width. This ridge exhibits a marked angular change in direction at one point, averages 1.5 to 2.5 m in height, and possesses a shallow enclosed depression along the axis of the ridge.

Sections were cut in the two highest mounds. In one, an ice core consisting of near vertically inclined ice layers was encountered at a depth of 1.5 m. In the other, although no ice was visible, the cross-section revealed that the two ramparts were composed of opposite-dipping beds of fluvial sand and gravel. These observations suggest the mounds to be the result of the growth of an ice core and the updoming of the overlying sediments. Samples of the ice were taken for laboratory analysis.

The age and origin of these pingo-like mounds is not yet clear. The possibility that the depressed centres are dead ice hollows can be dismissed since glacial materials are not present on the terrace. The uptilting of the enclosing sediments, the absence of 'candled'

Figure 48.1. Pingo-like mound with summit depression on a low terrace, central Bernard River.

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Figure 48. 2. A very small collapsed mound consisting only of a shallow circular rampart, central Bernard River.

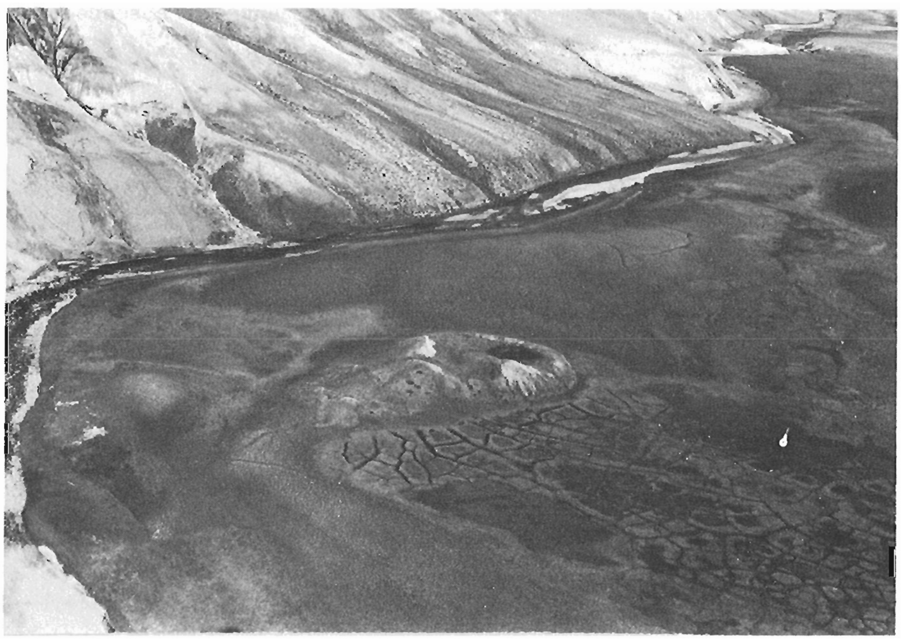

Figure 48. 3.

Oblique air view of pingo with summit crater and lake, upper Sachs River drainage basin. The feature is 7.5 m high.

ice, and the inclined banding of the ice indicate that these features are not simple ice mounds or 'naledi' as reported in the literature (e.g. Shumskii, 1964, p. 192-195; Church, 1972, p. 101-103). Furthermore, their density of occurrence and relatively small dimensions are not typical of mounds normally regarded as pingos. Finally, the features are different from, and much larger than, the very small hydrolaccoliths described from southern Banks Island (French, 1971) and elsewhere in the High Arctic (e.g. Bird, 1967, p. 201-203). The most comparable features so far described have been reported from northern Norway (Svensson, 1964) where they are regarded as a form of fossil thermokarst. For the moment, therefore, it seems best to regard the Bernard River features as a form of relic frost mound preserved within a permafrost environment.

Figure 48. 4. Oblique air view of pingo-like ridges, upper Sachs River drainage basin, illustrating the parallel alignment of the ridges, their central depressions, and the location of the ridges within an asymetrical valley. The pingo illustrated in Figure 48. 3 and the section excavated through one of the ridges are located by arrows.

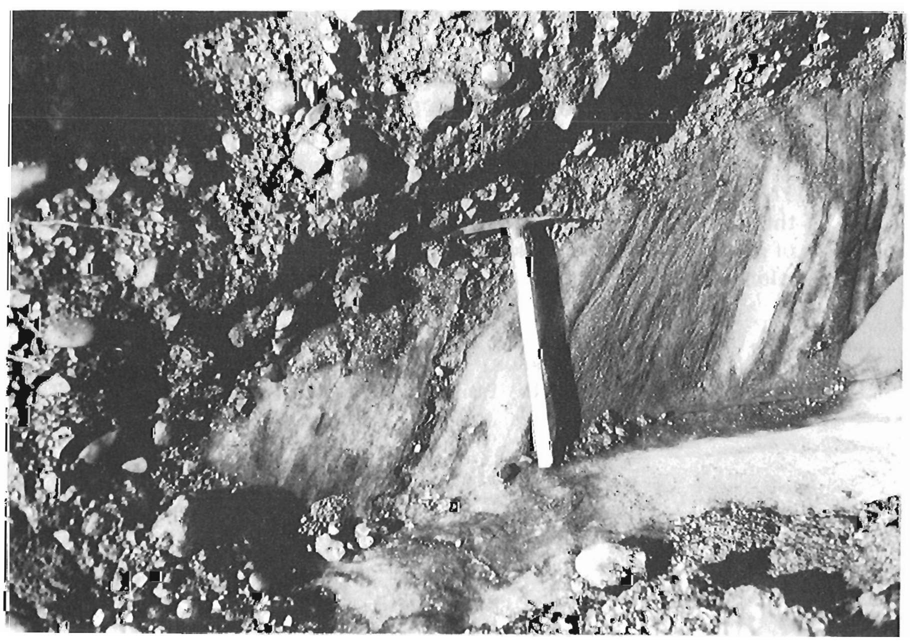

Figure 48. 5.

Illustration of the ice found in section excavated through one of the pingo-like ridges showing mineral material enclosed within the ice and the gravelly nature of the enclosing sediments.

Upper Sachs River (71°47'N, 123°12'W)

Air photographs (A16285 104-5) indicated the existence of two peculiar ridges and the presence of a partially collapsed pingo within a small stream valley in the headwaters of upper Sachs River drainage basin. The unusual location of the pingo, the puzzling nature of the ridges, and the possibility of excavating a section through one of the ridges prompted field investigation.

The pingo is located in a depression formed at the junction of a tributary valley (Fig. 48. 3). Judging by the ice-wedge pattern of the tundra terrain which surrounds it, the pingo grew at one end of a shallow lake which has now disappeared. Attention was then focused upon the two ridges which extend upvalley for a distance in excess of 2. 0 km (Fig. 48. 4). These ridges are approximately parallel to each other, varying from 50 to 200 m apart, and are located towards the foot of the northwest facing slope. The valley is strikingly asymmetrical with the present stream undercutting the steeper southeast facing slope. Neither ridge exceeds 10 m in height, and each varies in width between 20 and 200 m. The ridges possess little surface vegetation and are composed of coarse sand and gravel, giving rise to a relatively fresh, hummocky topography. Characteristic of both ridges are a number of shallow, elongate, and enclosed depressions which follow the long axes of the ridges (Fig. 48. 4).

A section cut partially through one of the ridges revealed large ice bodies existing at depths of 1. 0 to 1. 5 m beneath the depressions (Fig. 48. 5). the overlying sand and gravel was virtually structureless but relatively well sorted, with no boulders exceeding 30 cm in diameter. Samples of ice were taken for analysis and for comparison with known pingo ice.

No convincing explanation for these pingo-like ridges is advanced. Their elongate and hummocky form and their composition of coarse gravels and sands favour their interpretation as eskers. However, their location towards the bottom of a sharp asymmetrical valley, the shallow depressions aligned along their axes with ice bodies beneath, and the absence of other evidence for recent glacial activity within the valley do not support this interpretation. Moreover, the dual nature of the ridges and their relationship to the pingo immediately downvalley is problematical. If these ridges are of a pingo origin, their mechanism of growth appears obscure, and more detailed investigations are clearly required.

Thomsen River and Able Creek (73°45'N, 119°50'W)

Two ^{14}C dates obtained from the Thomsen River study area of 1974 provide information concerning the age of these pingos. Willow fragments enclosed within the sediments overlying the ice core of one of the Able Creek pingos (Pissart, 1975) have given an age of 4990 ± 90 years B.P. (GSC – 2117). This data provides a maximum age for the growth of the pingo. In Thomsen River valley, twigs and branches of willow found beneath 8. 0 m of windblown sands have given an age of 3460 ± 80 years B.P. (GSC – 2124). These sands occur on the terrace on which the pingos are located (French, 1975). The date indicates the minimum age for the abandonment of that terrace by the river and for the commencement of wind transportation and deposition. In all probability therefore, the pingos of Thomsen River and Able Creek localities developed in the time period bracketed by these dates.

References

Bird, J.B.
1967: The physiography of Arctic Canada; Johns Hopkins Press, Baltimore, 336 p.

Church, M.
1972: Baffin Island sandurs; a study of arctic fluvial processes; Geol. Surv. Can., Bull. 216, 208 p.

French, H.M.
1971: Ice cored mounds and patterned ground, southern Banks Island, western Canadian Arctic; Geogr. Ann., v. 53A, p. 32-38.

1975: Pingo investigations and terrain disturbance studies, Banks Island, District of Franklin; *in* Report of Activities, Part A, Geol. Surv. Can., Paper 75-1A, p. 459-464.

Pissart, A.
1975: Banks Island, N.W.T.: pingos, wind action, periglacial structures; *in* Report of Activities, Part A, Geol. Surv. Can., Paper 75-1A, p. 479-481.

Shumskii, P.A.
1964: Principles of Structural Glaciology; Dover Publications, New York, 479 p.

Svensson, H.
1964: Pingo-like frost mounds; Svensk Geogr. Arsbok, v. 40, p. 93-106.

49. SKEENA DELTA SEDIMENTATION, BRITISH COLUMBIA

Project 740062

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It is evident from a recent Environment Canada report (Hoos, 1975) that the Prince Rupert-Skeena estuary area has the potential for continued, large-scale development. It is equally clear, however, that although environmental stresses imposed by this growth may be considerable, they cannot adequately be assessed on the basis of our current knowledge of prevailing coastal physical, chemical, and biological regimes.

The Prince Rupert-Skeena estuary region is a major centre for both the fishing/fish processing and forest products industry. As Prince Rupert is a transportation centre for the northwestern part of British Columbia, not only are docking facilities currently being expanded, but consideration also has been given to the erection of a major bulk loading facility. Furthermore, as the city is relatively close to coal resources in the interior of the province, it has been suggested as an alternative site for the development of a major steel mill proposed for British Columbia by Japanese interests. The growing industrialization and urbanization of the area has, in turn, markedly increased the sources of water pollution which now include sewage, fish processing plants, pulp mills, log handling and storage facilities, a chemical and cement plant, and shipping. In addition, burgeoning construction within the Prince Rupert area has increased demand for gravel and sand, local resources of which are rapidly dwindling (Clague, 1976).

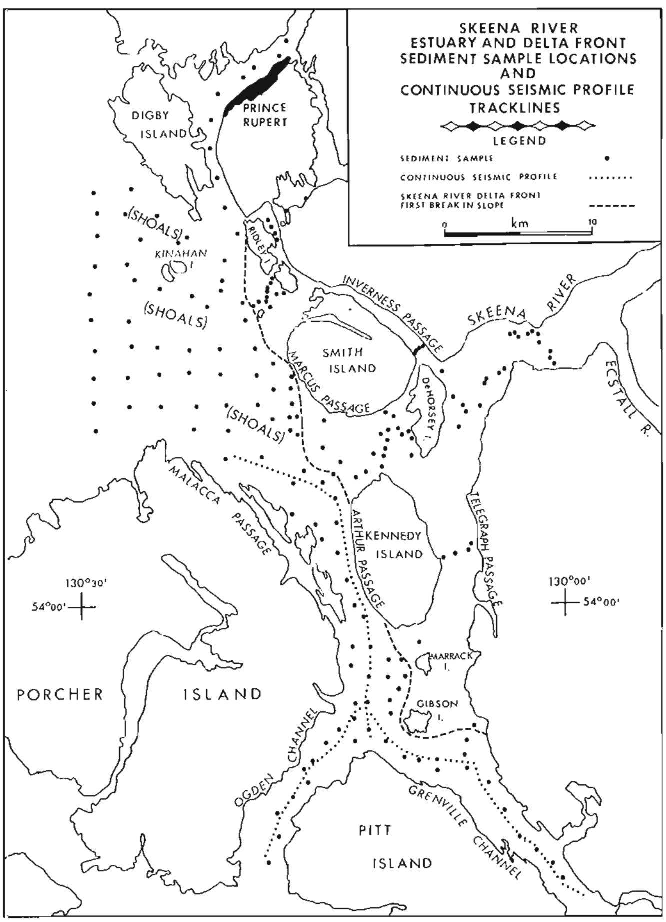

Figure 49.1. Location map.

A marine geology research program of the Prince Rupert-Skeena estuary area was initiated in 1974 to produce information, which ultimately can be applied to studies of vital coastal ecosystems, coastal and estuarine land use, pollutant dispersal and accumulation, and permit an assessment of offshore sand and gravel resources. A total of 210 sediment samples were collected from the sea floor off the delta front, from the lower reaches of the Skeena River channels, and from small beaches and basins in the vicinity of Ridley Island (Fig. 49.1). In addition, approximately 50 km of continuous seismic profiles were obtained along Grenville and Ogden Channels and along Arthur and Malacca Passages (Fig. 49.1).

Operations were carried out from the C.S.S. William J. Stewart (June 17 to 23) during a Canadian Hydrographic Service survey of Skeena River, from the C.S.S. Laymore (August 12 to 24) during a biological survey of various estuaries (under the direction of Dr. C. Levings of the Pacific Environment Institute), and from the C.F.A.V. Endeavour (August 25 to 30).

Over half of the collected samples have now been grain size analyzed, and a few remarks can be made about the estuarine-marine geological environment of the Prince Rupert-Skeena River area.

The lower reaches of the Skeena River channels cut across plutonic, metamorphic, and sedimentary rocks (Hutchison, 1967). In general they are floored by 0.250 mm mode sands which accumulate to form extensive bars. Megaripples, commonly having amplitudes and wave lengths as great as 0.5 m and 10 m, respectively, are produced on the surface of these bars by currents which can attain velocities in excess of 3 knots (150 cm per sec.) (Figs. 49.2 and 49.3 D, E, F). Scouring along the deeper portions of some channels is sufficient to sweep away all but gravel-sized matter (Figs. 49.2 and 49.3 I).

The margins of the river channels generally are floored by organic-rich silty or plastic clay or muddy gravels (Figs. 49.2 and 49.3 C, G, H). Broad, sandy flats lie at the edge of Skeena Delta, which extends in a discontinuous fashion for approximately 30 km from Ridley Island to Gibson Island (Figs. 49.1 and 49.3 A, B). Along the delta front a generally sharp transition is evident from channel sand to ocean basin muds (Figs. 49.4 and 49.5). At the mouth of the river between Kennedy Island and Marrack Island, however, sandy sediment extends farther into deeper water,

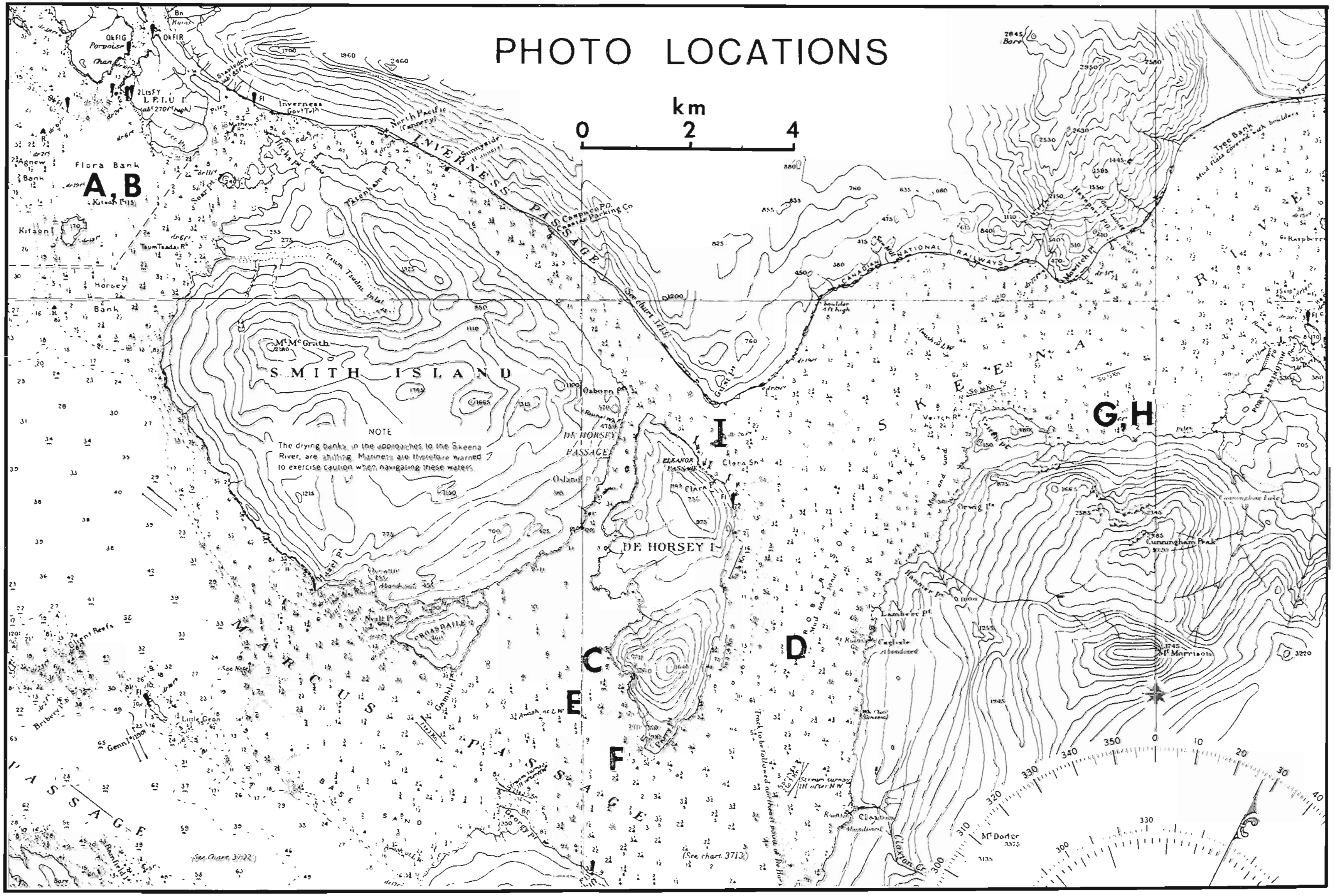

Figure 49. 2. Locations of photographs in Figure 49. 3. Base map: Canadian Hydrographic Service Chart No. 3735.

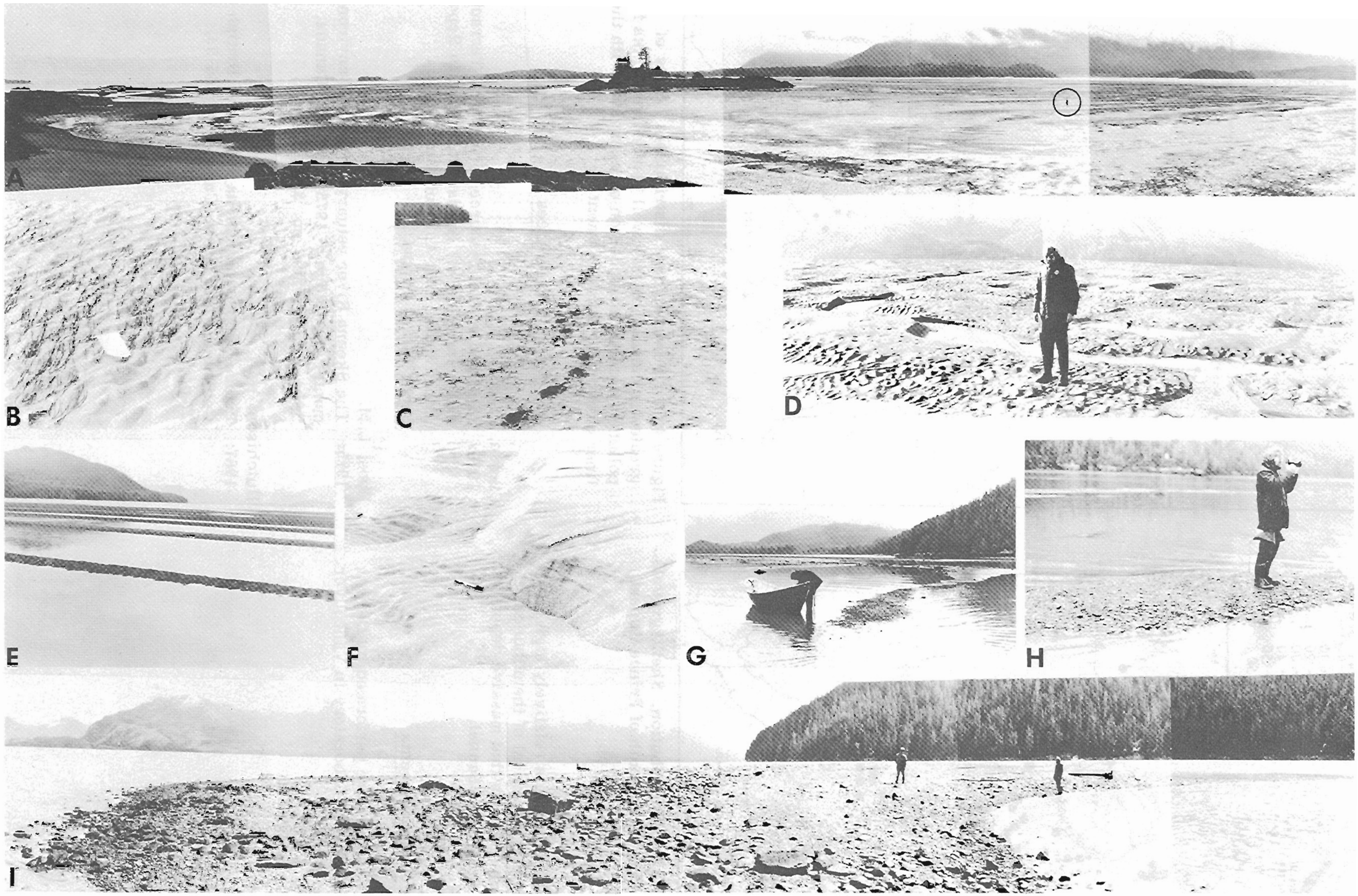

Figure 49.3 (A) Looking north along Flora Bank. This extensive sand bank lies at the mouth of Inverness Passage and is considered to be a vital nursery and feeding ground for fish (Hoos, 1975). The surveyor is circled.

(B) Representative view of the surface character of Flora Bank showing rippled sand surface and ecologically vital eelgrass (Zostera marina). A Fisheries Service study indicates that Flora Bank supports over 50 per cent of the total eelgrass in the Skeena estuary (Hoos, 1975).

(C) Typically flat, muddy beach common along margin of Skeena estuary river channels.

(D, E) Megarippled sands along channels.

(F) Avalanche face of megaripple with fine gravels at its base.

(G, H) Compacted muddy gravel beach at margin of channel.

(I) Mudflat on which has been dumped fill from a nearby source, most likely the floor of Inverness Pass (R. W. Sandilands, pers. comm.).

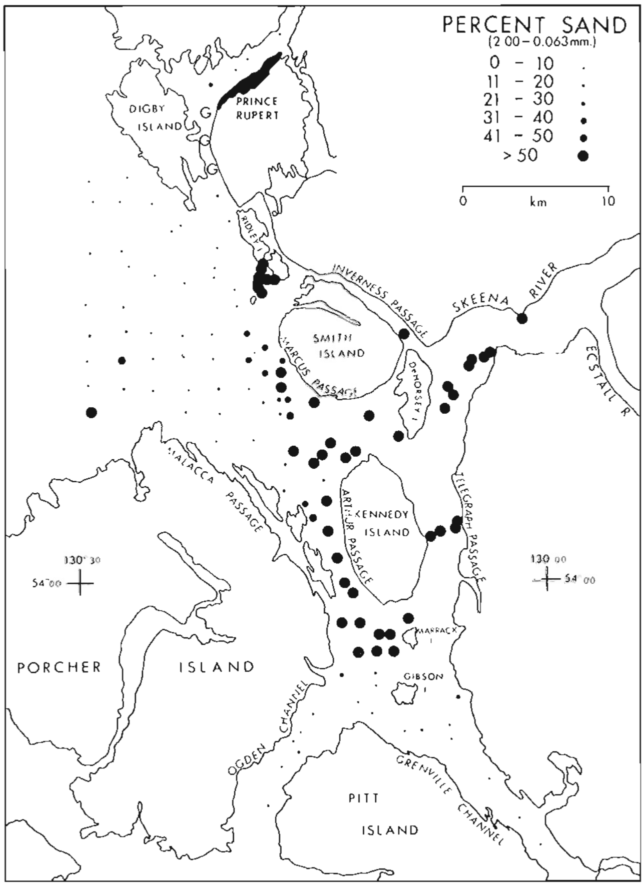

Figure 49. 4. Per cent sand distribution. Shelly gravels were collected at 40 m depth south of Prince Rupert at sites denoted "G".

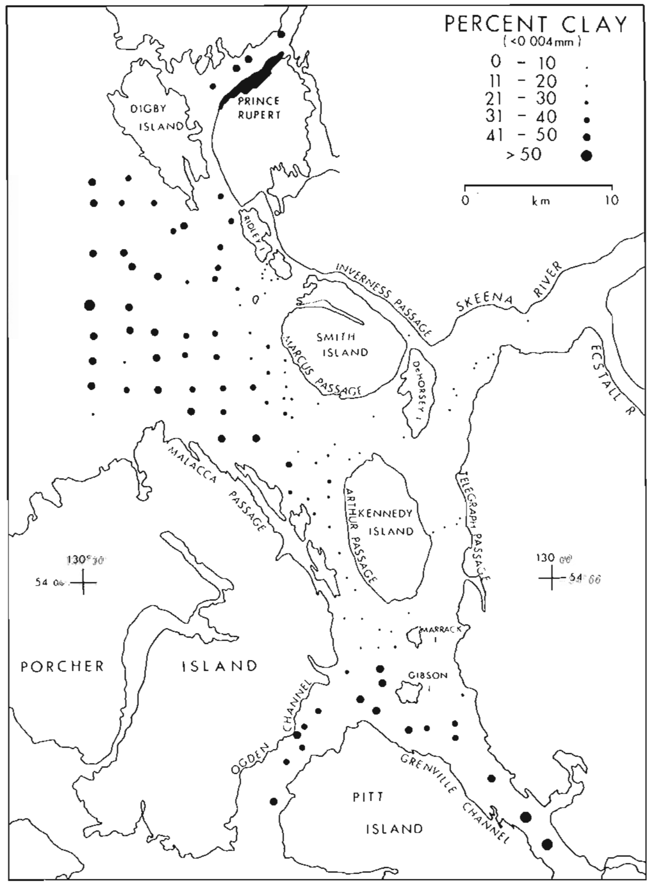

Figure 49. 5. Per cent clay distribution. Sites of greatest accumulation may well be dominant sinks for pollutant material absorbed on or associated with the finer component of the sediment.

suggesting that discharge here is relatively high. This agrees with a preliminary appraisal of the obtained seismic profiles which suggests that massive deposition has occurred all along Ogden Channel.

Acknowledgments

All grain size analyses were performed by Davis Swan at the Terrain Sciences Laboratory in Vancouver.

References

Clague, J. J.
1976: Sand and gravel resources of Kitimat, Terrace, and Prince Rupert, British Columbia; *in* Report of Activities, Part A, Geol. Surv. Can., Paper 76-1A, report 57.

Hoos, L. M.
1975: The Skeena River estuary – status of environmental knowledge to 1975; Environ. Canada, Spec. Estuary Ser. no. 3, p. 418.

Hutchison, W. W.
1967: Prince Rupert and Skeena map-area, British Columbia; Geol. Surv. Can., Paper 66-33, 27 p.

ENVIRONMENTAL GEOLOGY – HAMILTON URBAN AREA, ONTARIO

Project 730029

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Terrain Sciences Division

Since the end of the 1974 field season, data collected prior to that time have been corrected and entered into the Hamilton geoscience data bank (Morin, 1975). Experience from this phase of the study indicates that extensive manual corrections are necessary in order to eliminate inconsistencies and errors in coding so that the quality of the information is high enough to allow appropriate computer analysis and retrieval. Belanger (1975) has developed a new input format to overcome this problem in the future. In addition to the 5234 corrected records entered into the bank, 1347 seismic points and 6677 water well records from the Ontario Water Resources Commission were incorporated in the data base. During the 1973 and 1974 field seasons, 63 boreholes were drilled and almost 300 soil samples obtained for analysis.

Four maps thus far have been produced using the more than 13 000 data records in the data bank. These maps pertain to the entire area shown in Figure 50. 2, and consist of: a high water-table map, a penetration resistance map, a bedrock topography map, and a type of borehole map. The two first maps consist of data

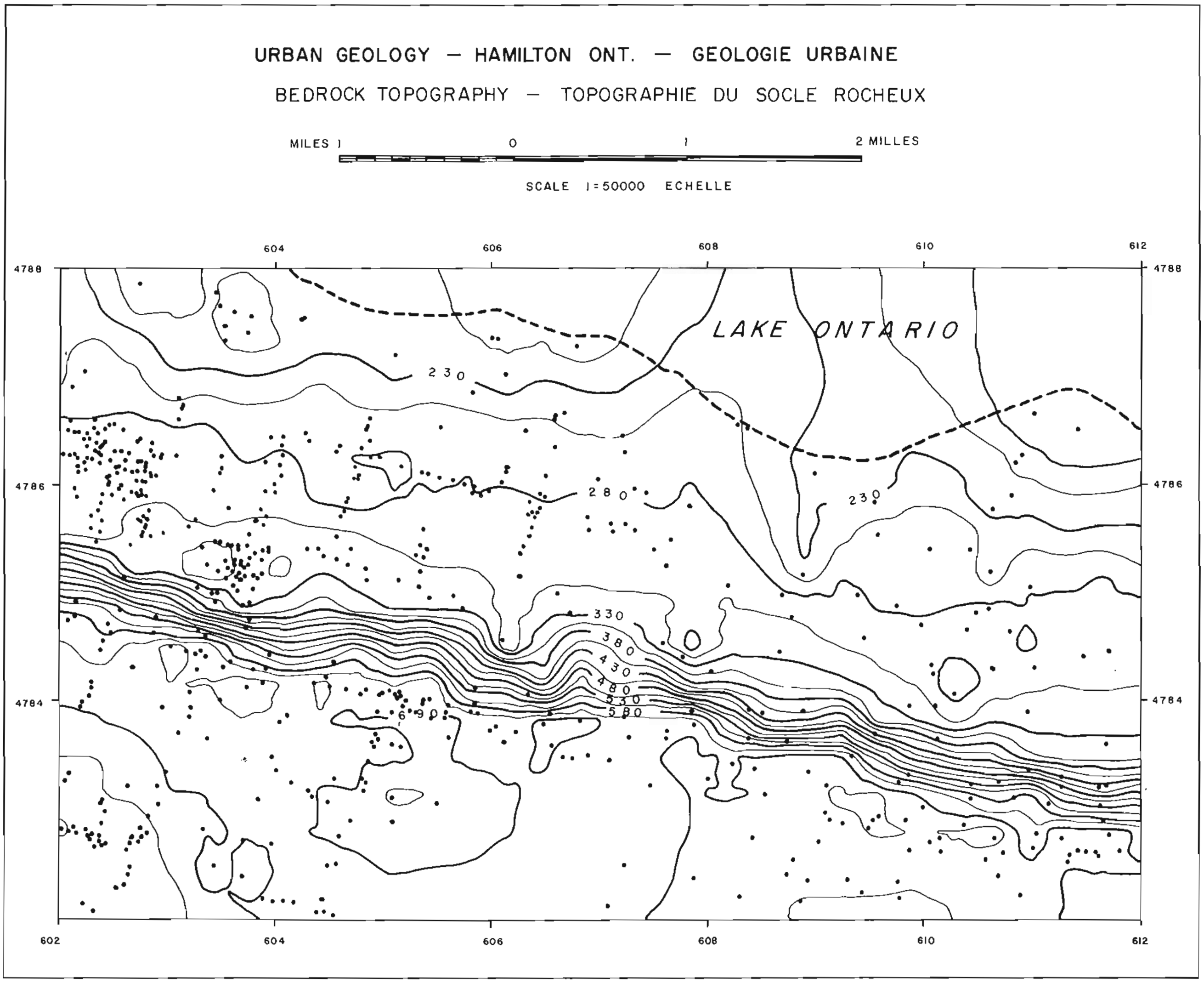

Figure 50. 1. Computer-drawn contours of the bedrock topography (trend surface map) for a small portion of NTS map-sheet 30 M/4g (1: 25 000). The dashed line represents the shoreline of Lake Ontario. Numbers around the border refer to the UTM grid system (zone 17). Dots indicate data points.

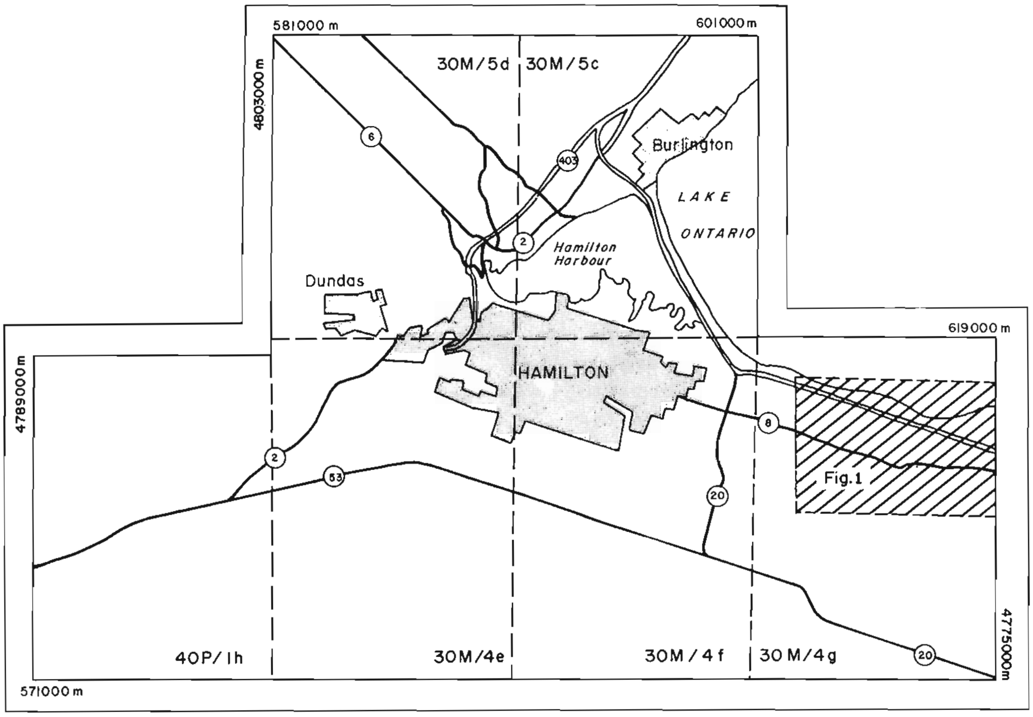

Figure 50. 2. Environmental geology, Hamilton, project area (1: 250 000). Hachures indicate area covered by Figure 50. 2. Numbers between borders refer to the U. T. M. gird limits (zone 17). The 1: 250 000 NTS map-sheets are indicated in dashed lines and their number shown.

retrieved as points (location of the boring). No extrapolation is attempted. A combination of colours and symbols is used to convey information. For example; the high water-table map symbols indicate the location and the season that the boring was made. The colour of the symbols indicates depth ranges at which the water table was first encountered (0 to 5 feet = red, 5 to 10 feet = yellow orange, etc.).

A portion of the bedrock topography map is reproduced in Figure 50. 1. A second contoured map is being produced to show drift thickness trends.

A number of other maps are in preparation or are planned. A program to assign the soil descriptions compiled in the data bank to one of the major known soil formations described by Karrow (1963), Cowan (1972), and Feenstra (1975) is in the last stage of completion. This computer program will permit a wide series of retrievals consisting of useful geotechnical documents and a better statistical appraisal of the physical properties of these formations.

References

Belanger, J. R.
1975: U. G. A. I. S. data record instruction manual; Geol. Surv. Can. , Open File 292.

Cowan, W. R.
1972: Pleistocene geology of the Brantford area, southern Ontario; Ont. Dep. Mines and Northern Affairs, IMR 37.

Feenstra, B. H.
1975: Quaternary geology of the Grimsby area, southern Ontario; Ont. Div. Mines, prelim. map, p. 993.

Karrow, P. F.
1963: Pleistocene geology of the Hamilton-Galt area; Ont. Dep. Mines, G. R. 16, 68 p.

Morin, F. J.
1975: Environmental geology – Hamilton urban area, Ontario; *in* Report of Activities, Part A, Geol. Surv. Can. , Paper 75-1A, p. 369-370.

51. PHYSICAL, CHEMICAL, AND STRATIGRAPHIC ASPECTS OF SEDIMENTATION IN LAKE BASINS OF THE EASTERN ARCTIC SHIELD

Projects 730013 and 700014

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Introduction

In 1975 physical and chemical parameters that control sedimentation in eastern Arctic lakes were studied and integrated with work begun by Klassen, Shilts, and Dean in 1973 (*see* Klassen, 1975; Klassen *et al.*, 1975; Shilts and Dean, 1975). In addition to the original objectives of defining 1) controls of trace-element fixation in lake sediments and 2) processes and genesis of patterned ground in shoal areas of lakes, detailed studies of the sediment budget of lakes and of their subbottom sediments were initiated. The objectives of these studies are: 1) to establish baseline chemical information for unconsolidated sediments in lake basins located in areas of potential mining activity. This information can be used to monitor potential environmental dislocations caused by mining and to compare the relative effectiveness of lake sediments and drift as geochemical exploration media; 2) to define the extent of near-bottom permafrost within lake basins; and 3) to define the present-day chemical and physical processes of sedimentation in lakes so that effects of future physical disturbances in their drainage basins may be better predicted. The principal disturbances anticipated are construction activities associated with mine-site development and installation of an Eastern Arctic gas pipeline.

To achieve these objectives, several tens of kilometres of subbottom profiles were completed in Yandle, Spi, and Carr lakes, several hundred samples of lake and associated glacial sediments were collected from Yandle, Henninga, and Spi lakes, and chemical and temperature profiles were measured in 22 lakes in the Baker-Yandle Lake area (Fig. 51.1).

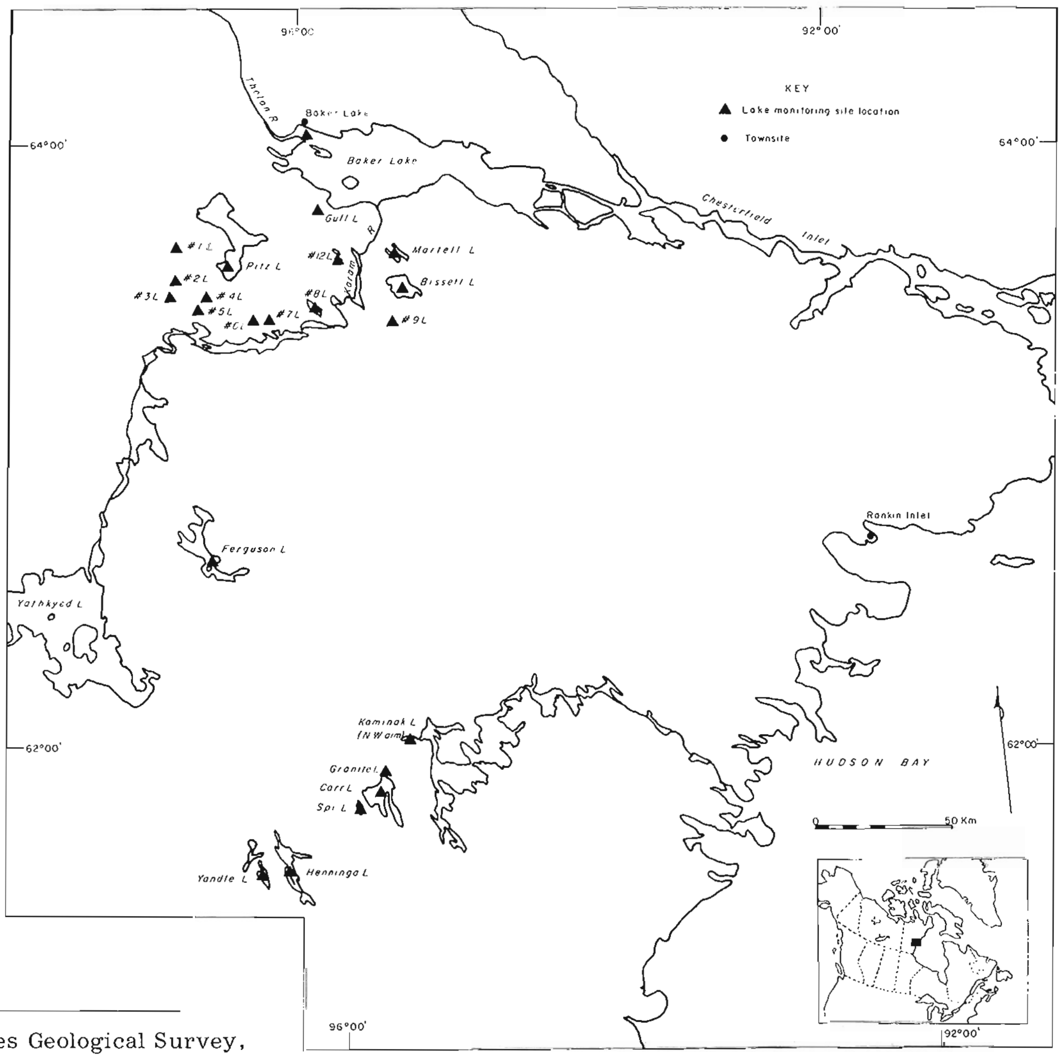

Figure 51.1 Map showing location lakes mentioned in report; triangles locate lakes where chemical and thermal profiles were measured (*see* Table 51.1).

[1]United States Geological Survey, Denver, Colorado, U.S.A.

Preliminary Results

Lake water temperature, pH, dissolved oxygen concentration, electrical conductivity, and reduction-oxidation (redox) potential were monitored with depth (Table 51.1). Measurements were made on August 8 and 9, 1975 using a Hydrosonde 6D Surveyor that was transported and operated from a float-equipped Beaver aircraft. The unit was inspected and calibrated prior to shipment into the field. Temperature and dissolved oxygen sensors were calibrated in the field after reassembly and were not found to require adjustment. Conductivity and redox probes were not checked against standards in the field after reassembly. All units performed well and there was no reason to suspect malfunction.

Thermal Data

Areas of lakes monitored ranged from approximately 100 to >10 000 ha and water surfaces were from 2 to 175 m above sea level. Sites monitored were less than 20 m in depth, except for Kaminak and Carr lakes. Surface water temperatures ranged between 10° and 14.5°C. Thermal stratification was found only in Baker, Kaminak, and Carr lakes, although the latter lakes could not be monitored adequately due to the 20 m lower operation depth of the sensor unit.

Dissolved Oxygen Data

The levels of dissolved oxygen were at or near saturation at approximately 12 ppm and did not decrease significantly with depth in any of the lakes, indicating either little oxygen uptake by organic matter in the sediment or thorough mixing at the air-water interface during the summer.

Conductivity

Electrical conductivity is an indirect measure of the concentration of material in solution. Conductivities ranged between 20 and 40 µmho/cm in the lakes (Baker Lake excepted) and indicate that levels of dissolved material are low. Measurements made in Kaminak Lake area during July and August 1973 were approximately ten times higher (Klassen *et al.*, 1975), but these levels were not duplicated either in this survey or in a second, independent survey (L. Johnson, pers. comm.) and are considered to be in error. Baker Lake is chemically as well as thermally stratified. The chemical contrast in dissolved material between surface and bottom waters is probably due to a layer of brackish water that enters its basin from Chesterfield Inlet.

Redox Potential

Redox potentials of all lakes were positive and ranged between 130 and 300 mv. These measurements confirm that depositional environments are oxidizing in all basins studied.

pH

Values of pH greater than 7 and slightly higher conductivities, as compared with other lakes, were found in four adjacent lakes (lakes 4, 5, 6, and 7) and Martell Lake in the Baker Lake area. These characteristics may be due to higher levels of carbonate material in either bedrock or surficial deposits within their drainage basins. Similarly, elevated pH values were found in an area of Kaminak Lake that is underlain by syenitic-carbonatitic intrusive rocks (Klassen *et al.*, 1975).

Subbottom Profiling and Sampling

Profiling was done using a Raytheon RTT-1000 Survey System, with a low frequency (7kHz) transducer, mounted on a 16-foot Canova boat (Fig. 51.2). A modified Livingston corer and dredge-type grab sampler with a calibrated winch also were mounted on the profiling instrument board. All systems are highly

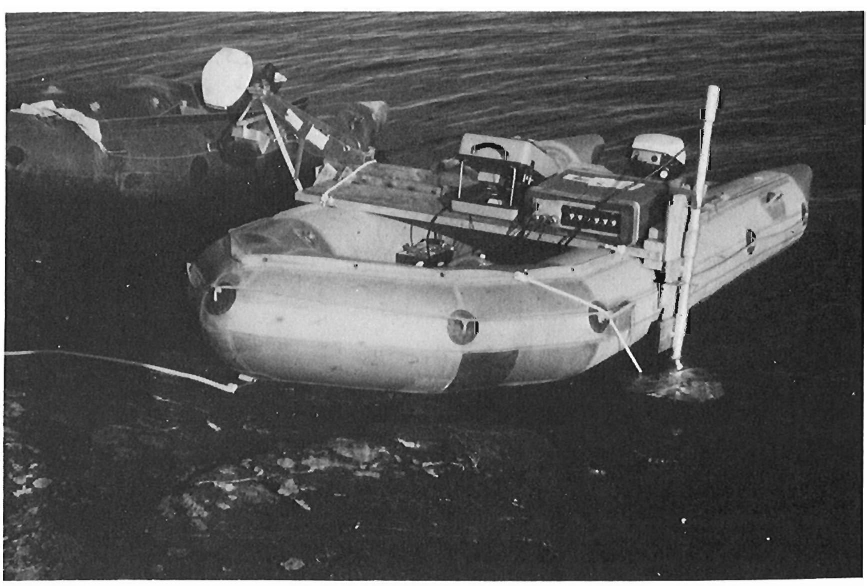

Figure 51.2.

A 7kHz transducer mounted on 16-foot Canova boat; note dredge sampler mounted on left side of instrument board. Power source is a 12-volt wet-cell battery with AC converter.

TABLE 51.1

LAKE WATER DATA, BAKER-KAMINAK LAKE AREA

LAKE	ALTITUDE (m)	LAKE AREA (approx.) (ha)	DEPTH (m)	LAKE WATER PARAMETERS TEMP. (°C)	pH	DISSOLVED OXYGEN (ppm)	CONDUCTIVITY µmho/cm	REDOX POTENTIAL (mv)
Baker L.	2.6	>>10 000	2	10.0	6.9	14.0	30	235
			3	10.0	6.9	14.0	30	235
			4	9.5	7.0	14.4	14	230
			5	7.0	7.0	15.0	650	235
			6	5.5	6.9	14.8	850	240
			8	6.0	6.9	14.4	900	240
			10	5.0	6.9	14.5	1150	250
			12	4.3	6.85	14.6	1300	245
			14	4.0	6.85	14.4	1300	250
			15.8 (bottom)	4.0	6.85	14.4	1350	250
Baker L.	2.6	>>10 000	1 (shore area)	7.0	6.8	14.7	930	170
Baker L.	2.6	>>10 000	1 (shore area)	6.0	6.8	15.0	1000	160
Gull L.	33	750	2 (bottom)	12.5	6.4	13.0	20	180
No. 1 L.	175*	200	1	13.0	6.6	13.4	15	180
			5.4 (bottom)	13.0	6.6	13.0	15	190
No. 2 L.	120	425	1	13.5	6.8	12.4	20	175
			5 (bottom)	13.0	6.7	12.4	20	185
No. 3 L.	120	250	1	13.0	6.7	13.0	15	170
			4 (bottom)	13.0	6.7	13.0	15	130
No. 4 L.	115	850	1	13.0	7.3	12.8	30	160
			4 (bottom)	12.5	7.3	13.0	30	150
No. 5 L.	115	600	1 (bottom)	13.5	7.2	13.0	20	180
No. 6 L.	110	125	1	13.0	7.2	12.0	40	200
			1.8 (bottom)	13.0	7.3	11.8	40	195
No. 7 L.	110	100	1	12.5	7.2	12.8	40	205
No. 8 L.	107	1 350	1	13.0	6.8	12.6	25	220
			8.2 (bottom)	12.5	6.8	12.4	20	210

* above the limit of postglacial marine transgression

Table 51.1 (cont.)

LAKE	ALTITUDE (m)	LAKE AREA (approx.) (ha)	DEPTH (m)	LAKE WATER PARAMETERS TEMP. (°C)	pH	DISSOLVED OXYGEN (ppm)	CONDUCTIVITY µmho/cm	REDOX POTENTIAL (mv)
No. 9 L.	78	375	1	13.0	6.8	12.3	25	220
			7.2 (bottom)	12.5	6.8	12.2	20	210
Bissett L.	68	6 250	1	13.0	6.8	12.6	30	220
			4.8 (bottom)	13.0	6.8	12.4	30	225
Martell L.	75	2 000	1	13.0	7.0	12.4	40	230
			5 (bottom)	13.0	7.1	12.4	40	220
No. 12 L.	58	375	1 (bottom)	13.0	6.9	12.8	20	220
Henninga L.	90	5 500	1	14.5	6.7	12.4	25	265
			6	14.5	6.7	11.9	30	235
			8 (bottom)	14.5	6.7	11.9	30	235
Yandle L.	110	1 700	1	14.0	6.9	12.2	35	250
			5 (bottom)	14.0	6.9	12.0	25	260
Spi L.	68	950	1	14.0	6.5	12.1	20	265
			6	14.0	6.5	12.0	20	250
			8.6 (bottom)	14.0	6.4	11.9	20	245
Carr L.	61	9 500	1	10.0	6.5	14.4	30	285
			20	7.5	6.3	14.2	30	290
'Granite' L.	65	450	1	14.0	6.8	12.6	20	285
			10.2 (bottom)	13.5	6.7	12.0	20	290
Kaminak L. (NW arm)	54	550	1	13.3	6.4	12.0	30	295
			15	10.5	5.9	10.6	30	315
Ferguson L.	117	10 000	1	13.0	6.4	12.2	20	300
			11.6 (bottom)	13.6	6.3	11.7	20	300
Pitz L.	60	> 10 000	1	12.5	6.9	12.8	20	245
			6.4 (bottom)	12.0	6.8	12.8	20	250
Pitz L.	60	> 10 000	1 (shore area)	12.5	6.8	12.6	20	300
Pitz L.	60	> 10 000	1	13.0	6.9	12.0	20	180
			3	12.5	6.9	12.0	20	185
			5.4 (bottom)	11.5	6.9	12.0	20	185

Figure 51. 3. Location of traverses and sample-sounding points, Yandle Lake.

Figure 51. 4. Bathymetry in Yandle Lake. Note troughs resulting from concentrated glacial erosion around central islands. Numerous isolated depressions serve as sediment traps.

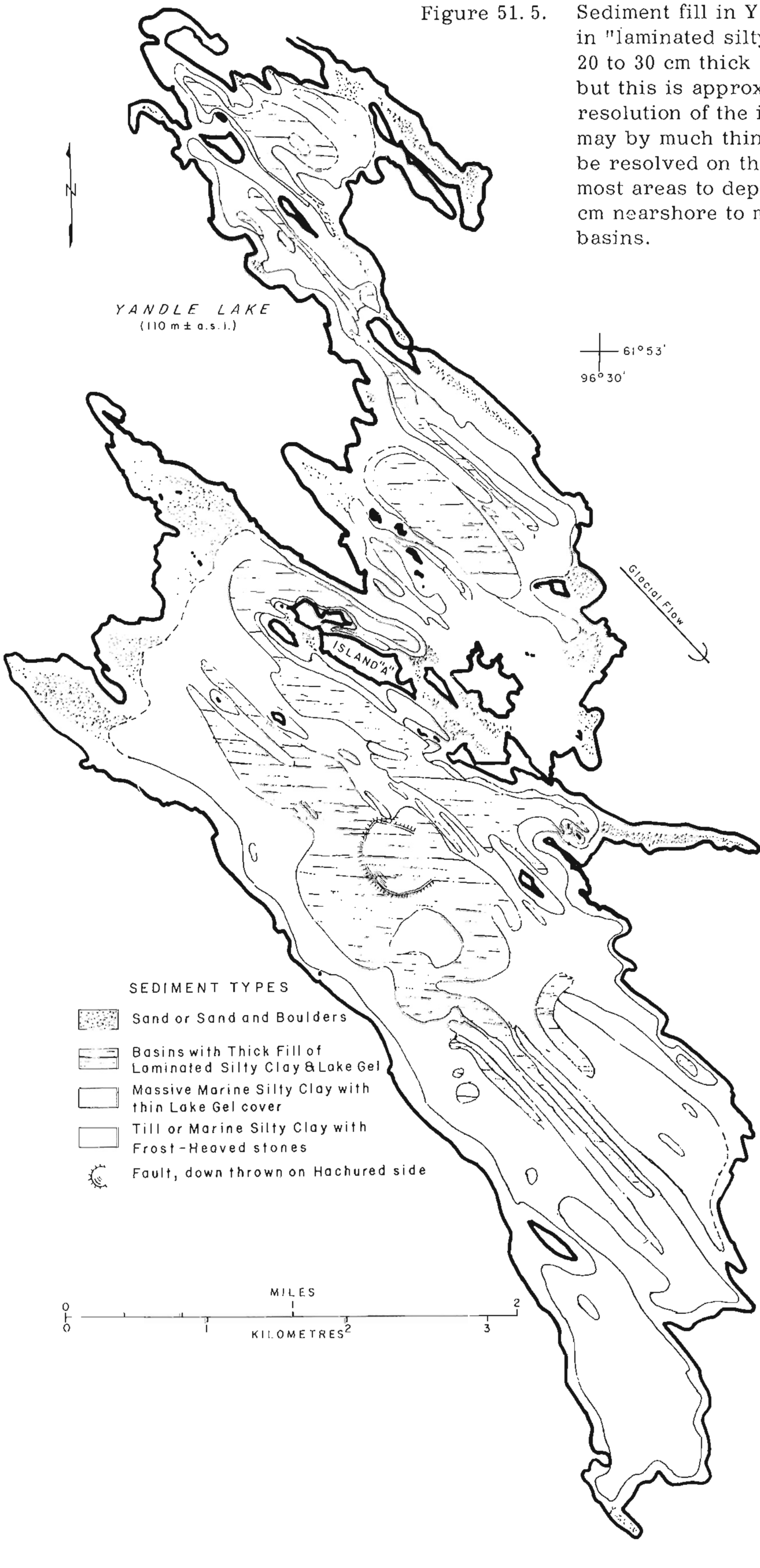

Figure 51.5. Sediment fill in Yandle Lake. Laminae in "laminated silty clay" appear to be 20 to 30 cm thick on subbottom profiles, but this is approximately the limit of resolution of the instrument and laminae may by much thinner. Lake gel cannot be resolved on the record but covers most areas to depths of less than a few cm nearshore to more than 1 m in offshore basins.

portable and performed well under rugged field conditions. Figures 51.3, 51.4, and 51.5 are preliminary drawings based on information obtained from these activities on Yandle Lake. Profiling and sampling indicate that most of the unconsolidated sediment in this basin is of either glacial (till) or marine origin (most basins studied were below the limit of postglacial marine submergence).

Lake gel

A stiff to watery, gel-like sediment that flocculates, rather than dispersing, in lake water is thought to represent true, modern lake sediment in this and other lakes studied. Based on observations by Klassen (1975) and of 1975 Livingston cores, this gel is thought rarely to exceed one metre in thickness, to be absent or very thin (<<10 cm) over large parts of the basin margins, and to contain 10 to 20 per cent organic matter by dry weight. The surface of the sediment is almost always a brilliant reddish-orange colour, and black or orange, 1 to 10 mm-thick horizontal bands commonly are observed through the length of a core. Round manganese nodules of 2 to 10 mm diameter are found within and at the surface of several samples. In shallower areas with major inlets, such as the southern bay of Yandle Lake, the gel may be admixed with considerable medium to coarse grained sand and granules; sand-sized to boulder-sized debris, apparently ice rafted onto the gel, is found sporadically throughout the basins. Many boulders protruding from the gel were highly weathered on their exposed surfaces but were fresh on buried surfaces. In Henninga Lake pods of greensih-yellow silty clay were found in two places below a black manganese-cemented surface layer or pavement. Radiography of cores of gel collected by Klassen (1975) reveals that delicate laminations of 1 mm or less are typical of many gel cores (Fig. 51.6); these laminae are thought at present to be chemical in origin and to represent some basic, possibly yearly, cycle in the sedimentation process.

Biogenic components of lake gel are generally microscopic in size, but few macroscopic plant remains, shrimp-like arthropods, and abundant benthic worms and worm tubes were found on or in the gel. Although fish are abundant in all lakes studied, no recognizable vertebrate remains were noted.

Massive and Laminated Silty Clay and Sand

In Yandle Lake subbottom profiling and coring revealed that significant thicknesses of massive to laminated, grey to pink silty clay underlie the gel and overlie a unit that is interpreted as till. In places over 7 m of this type of fill were noted. The silty clay presently is interpreted as marine sediment because of its strong physical similarity to subaerially exposed, fossiliferous marine silty clay around the sides of the lakes, its great thickness, and its general lack of biogenic components. Subbottom profiles show it to have some lamination, particularly where it has filled in depressions on the relatively rough original glacial floor of the basins. This lamination is thought to reflect turbidity flows triggered by slumping of soft sediment from the sides of the irregular drumlinized lake floors, particularly as marine wave base approached present lake levels during isostatic rebound. Also associated with the shoaling period of marine deposition in the lake basins are sandy pockets or strips that occur in shallow water adjacent to drumlinized peninsulas or their shallow extensions into the lakes. The sand is thought to have been derived by wave washing of the till surfaces and presently is overlain by at least several centimetres of the modern lake gel. These deposits are particularly common in Spi Lake where they form continuous, interdrumlin strips that run the length of the lake.

In Yandle and Carr lakes subbottom profiles revealed the presence of slump faults and faulting or kettles resulting from sediment collapse over melting buried ice blocks (Fig. 51.7). The laminated and massive silty clays were displaced or, in Yandle Lake, seemed to be closely associated with this faulting, further suggesting their formation during the 2000 to 3000 years during which marine waters filled these basins after glacial retreat.

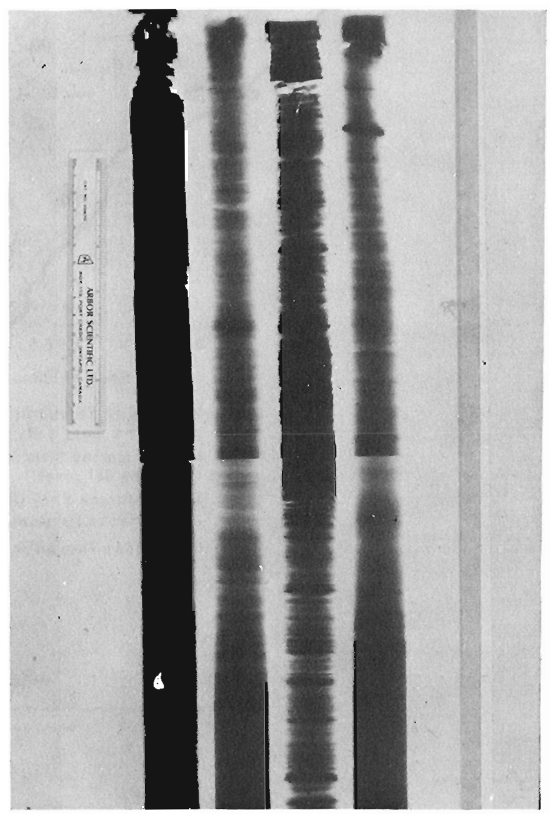

Figure 51.6. X-radiographs of typical gel from 'Granite' and Kaminak (northwest) lakes. Tapering of cores is due to drying in core tubes. Shrinkage is proportional to water content of sediment.

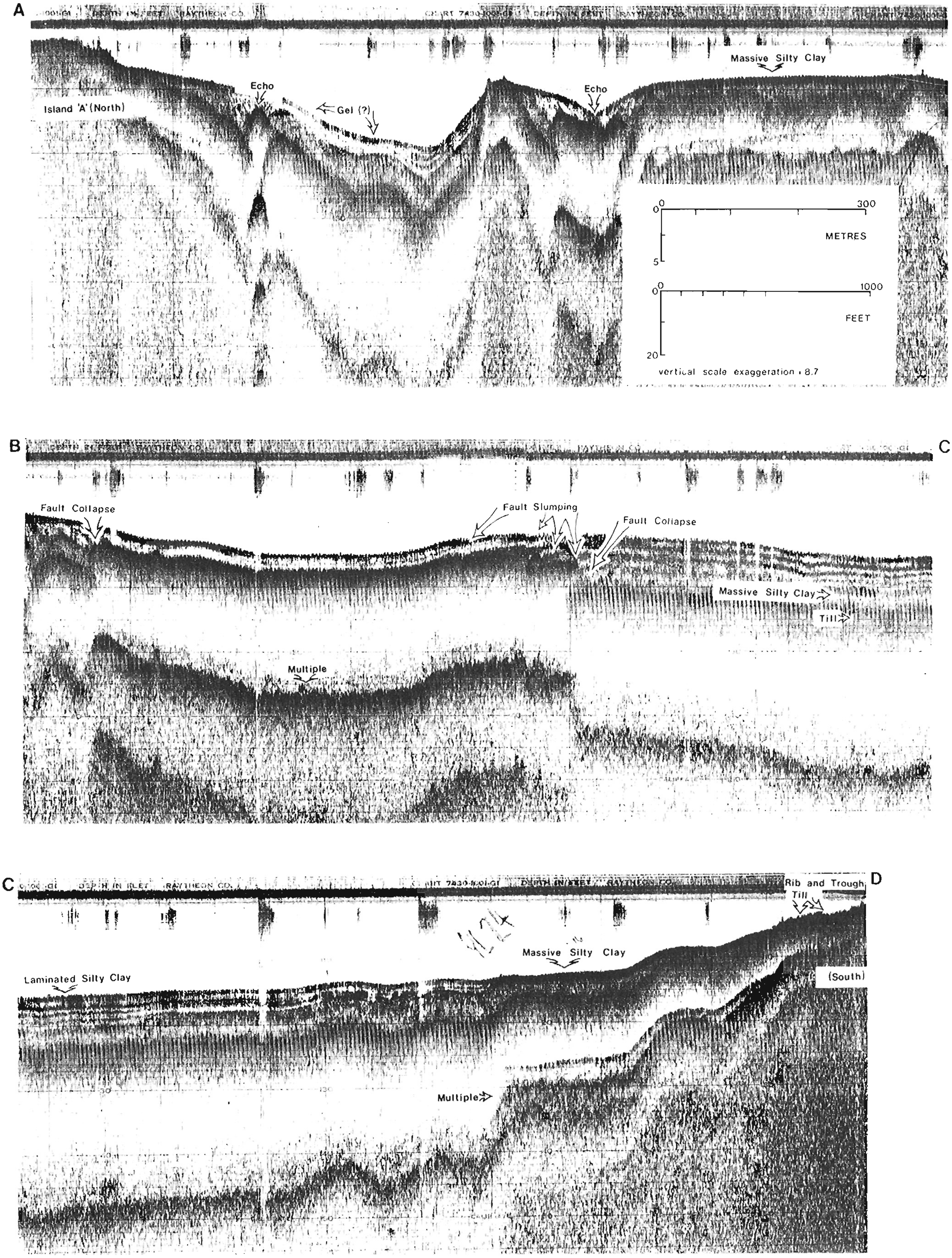

Figure 51. 7. A typical north-south subbottom profile extends from central island (A) to southwest shore of Yandle Lake.

Water depths in Yandle, Henninga, and Spi lakes are rarely greater than 13 m; but the southwestern arm of Carr Lake is a glacially scoured trench that is 100 to 115 m deep, its bottom lying more than 50 m below modern sea level. Remarkable profiles across this trench reveal that a 15 m-high, subaqueous extension of the Carr Lake esker lies in the deepest portions of the trench where it is flanked and interbedded with 3 to 5 m of horizontally laminated marine sediment.

Preliminary Conclusions

1) High levels of dissolved oxygen in lake waters are associated with intense oxidation or weathering of sediment at the sediment-water interface in this part of Keewatin.

2) Modern sedimentation rates are extremely low, and much of the basin fill is glacial or marine in origin.

3) Biogenic sediment may be severely degraded by oxidation, leaving only its most resistant components as part of the lake sediment. This may account for its low affinity for trace elements as noted by Klassen *et al.* (1975).

4) The shape and distribution of individual sedimentation basins within a lake are closely related to glacial erosional and depositional features.

5) Ice-rafting is probably an important mechanism of redistributing nearshore sediment; very thick drifting ice pans may cause ice-shove damage to shorelines thus providing at least temporary sources of sediment that can be redistributed by wave action.

References

Klassen, R.A.
1975: Geochemistry of lakes and lake sediments in the Kaminak Lake area, District of Keewatin, N.W.T.; unpubl. M.Sc. thesis, Queen's University, Kingston, 205 p.

Klassen, R.A., Nichol, I., and Shilts, W.W.
1975: Lake geochemistry in the Kaminak Lake area, District of Keewatin, N.W.T.; Verh. Internat. Verin-Limnol., v. 19, p. 340-348.

Shilts, W.W. and Dean, W.E.
1975: Permafrost features under Arctic lakes, District of Keewatin, Northwest Territories; Can. J. Earth Sci., v. 12, no. 4, p. 648-662.

DRIFT PROSPECTING IN THE DISTRICT OF KEEWATIN – URANIUM AND BASE METALS

Projects 700014 and 750074

W. W. Shilts and R. A. Klassen

Studies of trace metal distribution in till were continued in the Yandle-Henninga Lake area; sampling at one-mile intervals was extended to the south and west of already completed one-mile grids (Shilts, 1974, 1975). In this area detailed sampling at spacings of less than one quarter mile was undertaken within selected areas that had been previously determined by one-mile sampling to be anomalously rich in copper, nickel, or zinc. Follow-up sampling was designed to determine how representative the anomalies, defined by many widely spaced samples, are of the several square miles of terrain they are supposed to characterize.

A new project was begun to evaluate the effectiveness of drift sampling in uranium exploration. Detailed till sampling was carried out around known uranium mineralization within units of the Dubawnt Group and some Archean-age dykes in the Baker-Pitz Lake area, and at one site within a conglomerate near the base of Hurwitz Group sediments, about 5 miles southwest of Yandle Lake. Till sampling at one-mile intervals was completed over approximately 1200 square miles of terrain underlain by the Dubawnt Group and adjacent Archean-age bedrock.

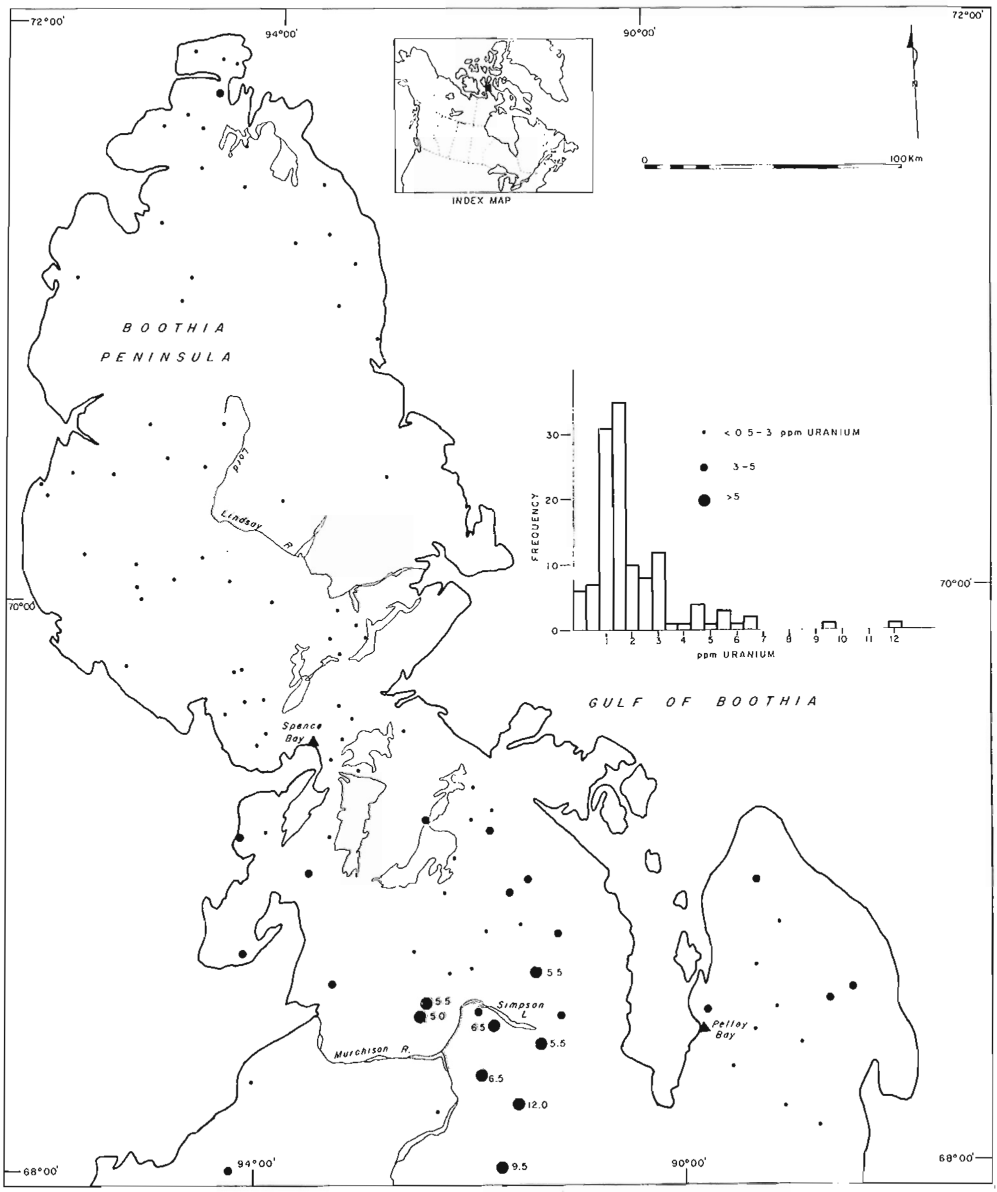

Figure 52. 1.

Uranium in till samples from mudboils on Boothia Peninsula, < 2μ fraction. Samples collected by A. N. Boydell, J. A. Netterville, K. A. Drabinsky, 1974 (project 740074).

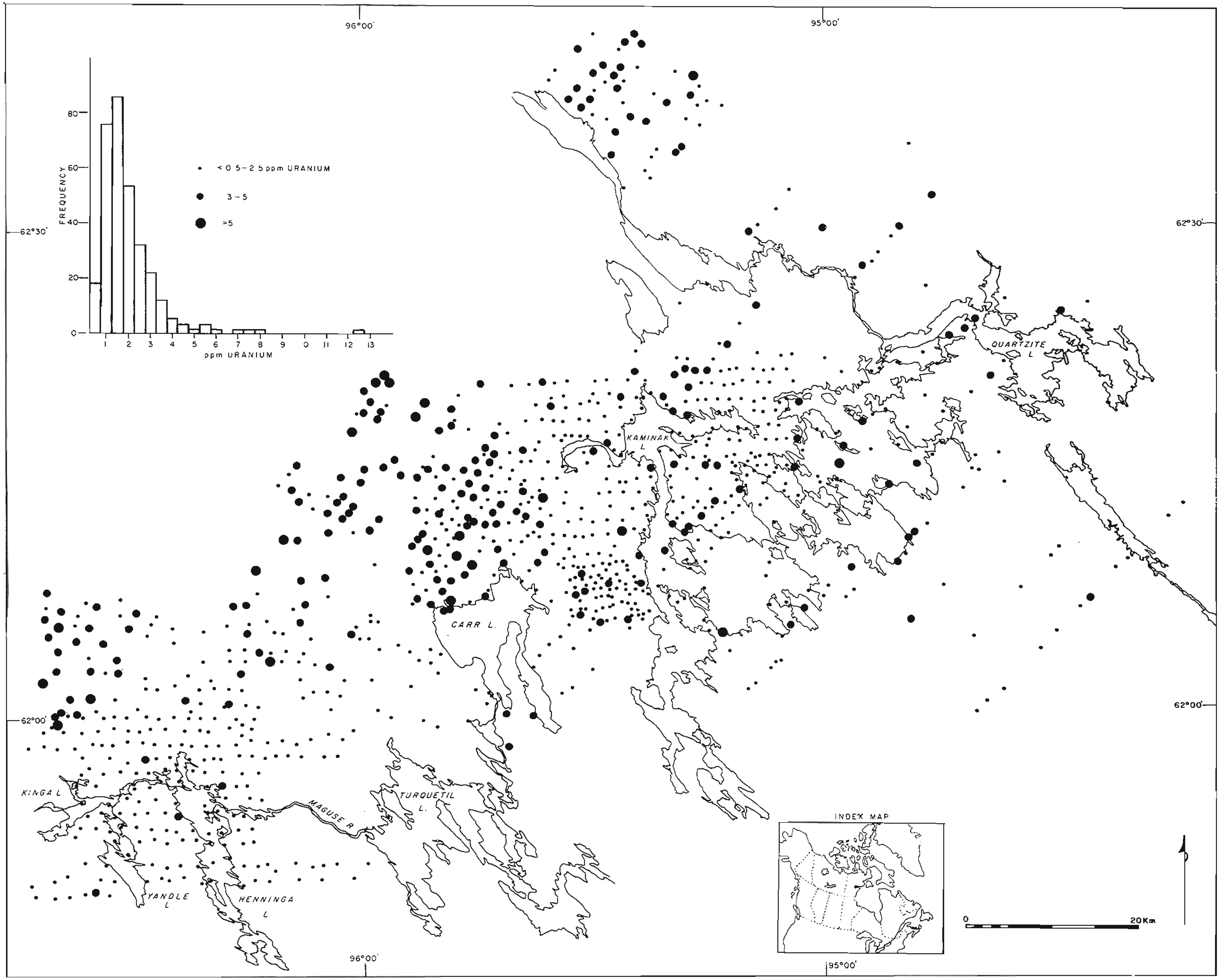

Figure 52. 2. Uranium in till samples from mudboils of Kaminak Lake region.

In addition to the drift sampling, R. W. Wein directed the collection of large samples of 10 to 15 species of tundra plants, peat, and mineral sediment at each of six sites. Two sites were directly over uranium mineralization, one was over copper-lead-zinc mineralization, one was in a dispersal train 2 miles down-ice from copper-lead-zinc mineralization, and two were background sites. This pilot project is designed to establish which, if any, typical tundra plant species might be useful as a medium in exploration for uranium or base metals. Ashed residue from these samples is presently being analyzed for several trace elements.

In preparing for the uranium sampling program over 1500 till samples, previously collected during drift prospecting and terrain mapping programs in the Kaminak Lake and Boothia Peninsula areas, were analyzed for uranium. Figures 52.1 and 52.2 illustrate the results of these analyses but are difficult to evaluate at present because of uncertainty about the magnitudes of uranium values to be expected in till from mudboils in close proximity to uranium mineralization.

In Boothia Peninsula a cluster of high values in widely separated samples occurs near the base of the peninsula in terrain mapped as an Archean-age gneissic complex (Fig. 52.1). Consistently high uranium values (5 to 12 ppm) were found in tills of the Simpson Lake-Murchison River area; the highest occurs near an area of modified granitic rocks (Heywood, 1961). The distribution of uranium values on this map merely may reflect the varying average background values of the different rock types from which the till was derived and at present should not be construed to represent areas that are favourable for uranium mineralization.

In the Kaminak Lake area groups of high uranium values are particularly prominent north and northwest of Carr Lake and north of Kinga Lake. Some of the high values that occur immediately north of Carr Lake are from till resting on or near the Hurwitz Quartzite. Both groups of high values, however, appear to the authors to relate to gneissic terrain mapped north and northwest of the Kaminak-Turquetil igneous complexes and their flanking metavolcanic and metasedimentary units. The clusters of high values may reflect a generally elevated background for uranium throughout the gneissic terrain, which in turn may be reflected by the till on and down-ice (southeast) from the gneissic areas. As with Boothia Peninsula, these clusters of high values should not necessarily be construed to indicate areas that would be favourable for uranium mineralization.

References

Heywood, W.W.

1961: Geological notes, northern District of Keewatin; Geol. Surv. Can., Paper 61-18, 9 p.

Shilts, W.W.

1974: Cu, Zn, and Ni concentrations in till of the Henninga-Kaminak-Quartzite Lake area, District of Keewatin; Geol. Surv. Can., Open File 211.

1975: Principles of geochemical exploration for sulphide deposits using shallow samples of glacial drift; Can. Min. Metall. Bull., May 1975.

Project 750063

W. Blake, Jr.
Terrain Sciences Division

The main objective of this project, initiated in 1975, is to establish a chronostratigraphic framework for as much of Quaternary time as possible throughout the Arctic Archipelago. Related objectives are: 1) to investigate the suitability of various methods of age determination, especially those which can be utilized beyond the range of radiocarbon dating (approximately 50 000 years); 2) to determine rates of crustal movement; and 3) to reconstruct past environments and events.

During the 1975 field season four weeks were spent in the Arctic, and collections were made on Bathurst, Ellesmere, and Banks islands (Fig. 53.1). On Bathurst Island particular emphasis was placed on resampling the organic deposits in the Stuart River valley; in southern Ellesmere Island the stratigraphic sequence exposed along a river north of Cape Storm was examined in more detail than had hitherto been possible; and on Banks Island attention was focused on the area near Nelson Head at the southernmost extremity of the island. Each of these three sites will be described in more detail in the sections that follow.

Stuart River Valley, Bathurst Island

This site was revisited in order that further study and collections could be made of a sequence of deposits occurring on the south side of the valley, approximately 1 km west of the mouth of Cut Through Creek. At the section exposed along the edge of a terrace, which McLaren discovered in 1955 during the course of 'Operation Franklin' (McLaren, 1963) and which the writer investigated briefly in 1963 (Blake, 1964), a sequence of peat layers is preserved beneath a veneer of silt (Fig. 53.2). This dark greyish brown (2.5Y 4/2) calcareous silt contains marine molluscs which ^{14}C dating has shown to be of postgalcial age; *Mya truncata* shells gave a value of 8670 ± 100 years (GSC-1854).[1]

The organic deposits at this site are made up of two main peat layers, separated by a gravel intercalation (probably a fan deposit) of varying thickness. Also present, mainly beneath the lower peat, are banded layers, which presumably are pond deposits. Some of these banded layers are calcareous. The uppermost 2.5 cm of the peat, which is separated from the overlying postglacial silt by a thin layer of shale pebbles imbedded in the surface of the peat, is more than 50 000 years old (GSC-165-2). On the basis of moss flora, the vascular plant remains, and the fossil arthropods found in the various peat horizons, the organic deposits here were assigned to the Stuart River Interglaciation (Blake, 1974). The fact that the entire sequence occupies a depression in the surface of the terrace is in keeping with the interpretation that a pond or small lake formerly existed at this site. It is also important to note that the lower peat layers are dominated by the aquatic moss *Scorpidium scorpioides* (determined by M. Kuc; unpublished G.S.C. Bryological Report Nos. 188 and 189), that the diatoms in the silt beneath the lowermost peat are all freshwater species dominated by *Fragilaria pinnata* (identified by S. Lichti-Federovich; unpublished G.S.C. Diatom Report No. 73-8), and that remains of water beetles are present in the peat layers (*Agabus moestus* and at least two species of *Hydroporus;* determined by J.V. Matthews, Jr.; unpublished G.S.C. Fossil Arthropod Report Nos. 1-73 and 6-73).

Details of the deposits along Stuart River, based on the writer's 1963 collections, have been published already (Blake, 1974). A fuller description of this interesting site will be prepared after the constituents of the 1975 collections have been examined in the laboratory. It is hoped that the new studies will shed further light on conditions during the Stuart River Interglaciation.

Cape Storm, Ellesmere Island

An analysis of radiocarbon age determinations on samples collected in 1967 and 1970 from the raised beaches a few kilometres north of Cape Storm, together with a discussion of the pattern of postglacial emergence, has been presented recently by the writer (Blake, 1975a). The purpose of the visit to Cape Storm in 1975 was to devote more time to a study of the deposits exposed along the river which bisects Andersrag Beach. The need for further study here was prompted by the fact that pelecypods collected near the base of an exposure in 1970 had turned out to be "old" rather than postglacial in age. The 1975 season proved to be a good one for re-examination. Not only had many of the perennial snowbanks along the north side of the river valley melted away by early August, but the river had been active on the north side of the valley as well so that fresh sections were exposed in a number of places.

Most of the time spent near Cape Storm was devoted to the site where shells had been found in 1970 (Fig. 53.3). Radiocarbon determinations on aragonitic pelecypod shells collected at 51 to 54 m a.s.l. and at a few metres above river level had resulted in the following values: *Mya truncata* — 27 700 ± 460 years (GSC-1409); *Hiatella arctica* — 38 300 ± 1360 years (GSC-1880). In addition to these two species other pelecypods present are *Macoma calcarea* and *Serripes groenlandicus*, plus barnacle fragments (*Balanus* sp.) and echinoid spines. The same four pelecypod species were found higher in the section in 1975, and a sample

[1] All identifications of marine pelecypods by the author.

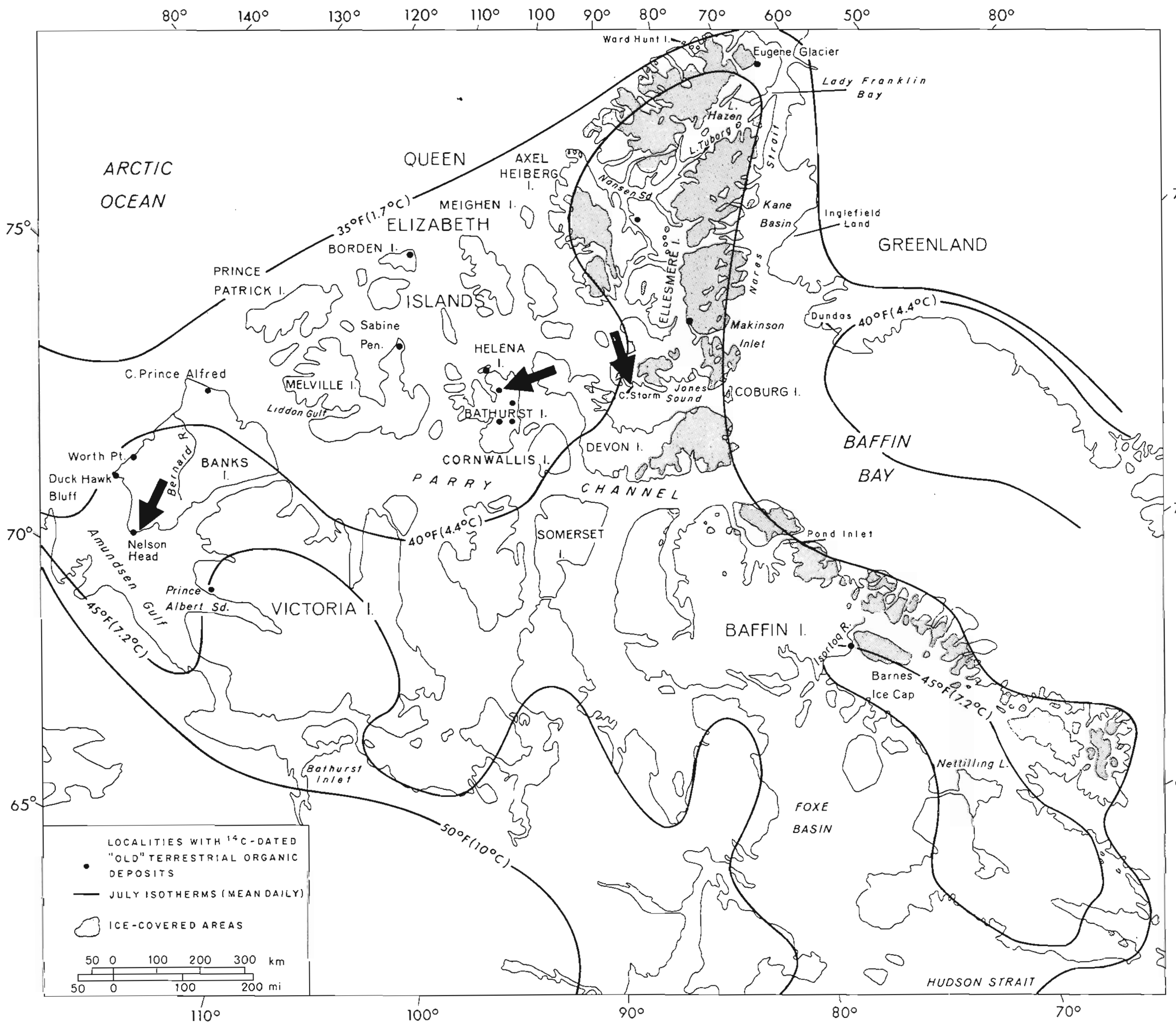

Figure 53. 1. Location map, Canadian Arctic Archipelago. Black arrows indicate the three main sites discussed in the text.

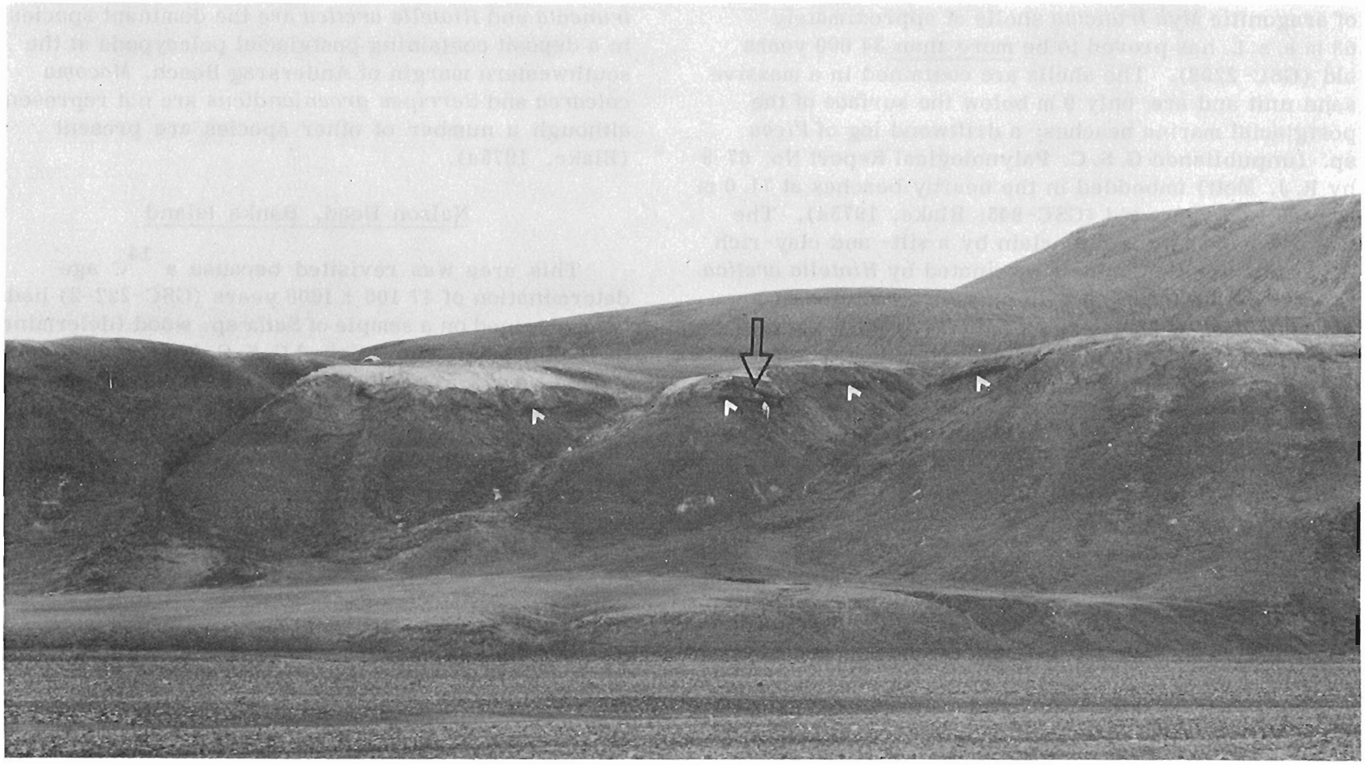

Figure 53.2. Peat layers (white arrows) under light-coloured postglacial marine silt on a terrace along the south side of Stuart River valley, northern Bathurst Island. The uppermost 2.5 cm of peat (open arrow) is more than 50 000 years old (GSC-165-2). July 31, 1975 (GSC 202936).

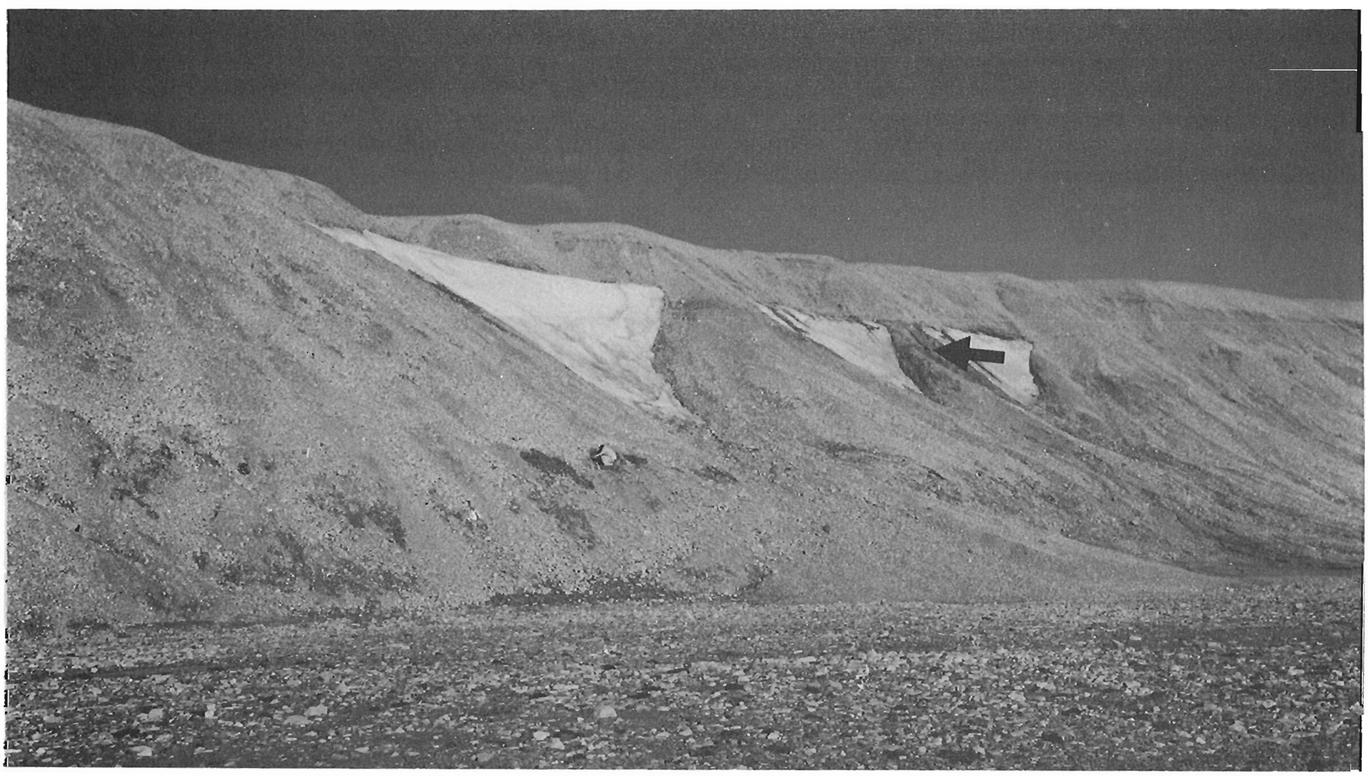

Figure 53.3. Upper site along river bisecting Andersrag Beach near Cape Storm, southern Ellesmere Island. R. Richardson is excavating at the 54 m level, where pelecypods collected in 1970 gave ages of 27 700 ± 460 years (GSC-1409; *Mya truncata*) and 38 300 ± 1360 years (GSC-1880; *Hiatella arctica*). Black arrow indicates position, at 63 m, of 1975 sample more than 34 000 years old (GSC-2209; *Mya truncata*). Surface of overlying postglacial beaches is at 72 m a.s.l. August 6, 1975 (GSC 202937).

of aragonitic *Mya truncata* shells at approximately 63 m a.s.l. has proved to be more than 34 000 years old (GSC-2209). The shells are contained in a massive sand unit and are only 9 m below the surface of the postglacial marine beaches; a driftwood log of *Picea* sp. (unpublished G.S.C. Palynological Report No. 67-8 by R.J. Mott) imbedded in the nearby beaches at 71.0 m is 8300 ± 70 years old (GSC-845; Blake, 1975a). The massive sand unit is underlain by a silt- and clay-rich bed, in which the fauna is dominated by *Hiatella arctica*. Downslope movement of both units has resulted in a concentration of large, robust shells near the base of the section, and it is this material on which determinations GSC-1409 and -1880 were based. Thus, most of the strata comprising the 24-m-high section predate the last glaciation. No deposit that definitely could be called a till was seen at this locality, but a cobble-boulder layer which overlies the massive sand unit and underlies the postglacial beach deposits may be related to a glacial episode.

About 1 km downstream at a 10 to 13 m-high section newly exposed this past summer (Fig. 53.4) a thin horizon, interpreted as being a till, was discovered within a sequence previously believed to comprise only beach and nearshore deposits. This "till" is 1 to 1.5 m thick, is composed of rounded cobbles in a sandy-silty matrix, and forms miniature cliffs. It is overlain by beach deposits and is underlain by bedded sands and by a silty-clayey unit. These underlying marine beds also contain echinoid spines as well as the same pelecypod fauna - *Mya truncata, Hiatella arctica, Macoma calcarea,* and *Serripes groenlandicus* — that was found at the section described in the previous paragraph. It seems reasonable to postulate that the cobbly "till" corresponds to the layer of cobbles and boulders which is present above the massive sand unit farther upstream. The overlying postglacial beach deposits at both localities contain only tiny fragments of shells, and although *Mya truncata* and *Hiatella arctica* are the dominant species in a deposit containing postglacial pelecypods at the southwestern margin of Andersrag Beach, *Macoma calcarea* and *Serripes groenlandicus* are not represented although a number of other species are present (Blake, 1975a).

Nelson Head, Banks Island

This area was revisited because a ^{14}C age determination of 47 100 ± 1000 years (GSC-222-2) had been obtained on a sample of *Salix* sp. wood (determined by L.D. Wilson; unpublished G.S.C. Wood Identification Report No. 73-51) collected in 1960 by J.G. Fyles from a sand-gravel-silt unit beneath till 5 km north of Nelson Head (Blake, 1974). At the time of writing this is still the only finite 'high pressure' (4 atm vs. 1 atm in the 5-L counter) age determination which has been obtained for the Canadian Arctic Archipelago in the Geological Survey's Radiocarbon Dating Laboratory (Blake, 1975b). Special care was taken with this sample, and the determination is based on a 9-day count; the three individual 3-day counts permitted ages of 47 400 ± 900, 48 300 ± 1000, and 45 600 ± 800 years to be calculated. It is important to remember, however, as has been emphasized earlier (Blake, 1974), that this sample showed only slight activity. Its counting rate is so close to the counting rate of the background that a small variation in either could result in the computation of a 'greater than' age.

Most of the stay in southernmost Banks Island was devoted to studying a complex succession of beds exposed along the coast some 4 to 5 km north of Nelson Head, but the present report is concerned only with a locality farther to the northeast, approximately 21 km from Nelson Head and 12 km southwest of Cape Collinson. At this site several tills, organic horizons, and massive silt units are present. In the brief time available only

Figure 53.4. Lower site along river at Andersrag Beach. Cobbly till is exposed above shovel held by R. Richardson and to left of the ice-axe (1.05 m long). The surface of the postglacial beaches above is close to 50 m a.s.l. August 7, 1975 (GSC 202938).

Figure 53.5. View northeastward towards Cape Collinson, southeastern Banks Island. J.-S. Vincent is indicating position of a thin organic layer where *Salix* sp. is more than 39 000 years old (GSC-2234). Above the woody detritus is a 3 m-thick cover of pinkish grey calcareous till. August 18, 1975 (GSC-202939).

the uppermost organic layer could be studied in any detail (Fig. 53.5). It is 2 to 5 cm thick, and in general the basal material consists of fine organic detritus (sapropel), followed by discontinuous banded silt and clay layers, some of which are calcareous. These are overlain in turn by coarser plant detritus including woody stems and branches with the bark still attached. The whole organic sequence is underlain by gravel and is overlain by: 1) a 5-cm-thick very dark greyish brown (2.5 YR 3/2) calcareous clay-rich horizon and 2) a calcareous till which is roughly 3 m thick and which is pinkish grey (7.5 YR 7/2) on the weathered surface.

A radiocarbon age determination has been carried out on the largest single piece of wood (5.1 g of *Salix* sp.; determined by R.J. Mott; unpublished G.S.C. Wood Identification Report No. 75-73) found in the plant detritus layer; it is more than 39 000 years old (GSC-2234; age based on two 3-day counts). The next step will be to take a larger sample from the same collection in order that an age determination can be carried out in the 5-L counter at 4 atm. This procedure will indicate whether the plant detritus layer is between 40 000 and 50 000 years old (e.g., a value such as that of 47 100 ± 1000 years obtained for wood (GSC-222-2 nearer Nelson Head) or whether it is beyond the limit of radiocarbon dating.

Acknowledgments

Support in the field was provided by the Polar Continental Shelf Project and by field parties from Terrain Sciences Division on Bathurst Island (D.M. Barnett) and Banks Island (J.-S. Vincent). The use of their base camp facilities and their help with transport to localities of interest on these two islands is acknowledged with gratitude, as is the assistance of R. Richardson. The Geological Survey's Radiocarbon Dating Laboratory, under the supervision of J.A. Lowdon, has determined the ^{14}C ages reported here. Identification of organic constituents has been carried out by staff members of the Paleoecology and Geochronology Section, as indicated in the text.

References

Blake, W., Jr.

1964: Preliminary account of the glacial history of Bathurst Island, Arctic Archipelago; Geol. Surv. Can., Paper 64-30, 8 p.

1974: Studies of glacial history in Arctic Canada. II. Interglacial peat deposits on Bathurst Island; Can. J. Earth Sci., v. 11, p. 1025-1042.

Blake, W., Jr. (cont'd)
1975a: Radiocarbon age determinations and postglacial emergence at Cape Storm, southern Ellesmere Island, Arctic Canada; Geogr. Ann., Ser. A, v. 57, p. 1-71.

1975b: Studies of glacial history in the Queen Elizabeth Islands, Canadian Arctic Archipelago; Naturgeografiska Institutionen, Stockholms Universitet, Forskningsrapport 21, p. 1-14.

McLaren, D.J.
1963: Stuart River area; *in* Geology of the north-central part of the Arctic Archipelago, Northwest Territories (Operation Franklin), Y.O. Fortier *et al.*, Geol. Surv. Can., Mem. 320, p. 596-620.

Project 730026

J.S.O. Lau and D.E. Lawrence[1]
Terrain Sciences Division

Introduction

The main objective of this project is to assemble all geotechnical data available in the Mackenzie Valley and Delta in order to assess the occurrence and distribution of permafrost and ground ice. In permafrost regions, engineering problems such as the thaw-consolidation-shear strength, slope stability, differential thaw settlement, and icing problems associated directly with permafrost result from disturbance in the thermal balance especially in those cases where fine textured soils exist. Influencing factors are the soil texture, thermal properties of soil, soil temperature, and the topographic, geological, hydrological, climatic, and vegetation conditions. An attempt will be made to correlate the occurrence and distribution of permafrost and ground ice with these factors in order to predict the presence of permafrost in areas where geotechnical data are sparse in order to prevent the engineering problems associated with permafrost.

Mackenzie Valley Geotechnical Data Bank

Over 10 000 boreholes have been drilled for the proposed highways and pipelines in the Mackenzie Valley and Delta. The principal sources of data are the geotechnical reports of the Department of Public Works of Canada and several consulting firms and surficial geology maps produced by the Geological Survey of Canada. The data are stored permanently in the Mackenzie Valley Geotechnical Data Bank (Lawrence, 1974a, b).

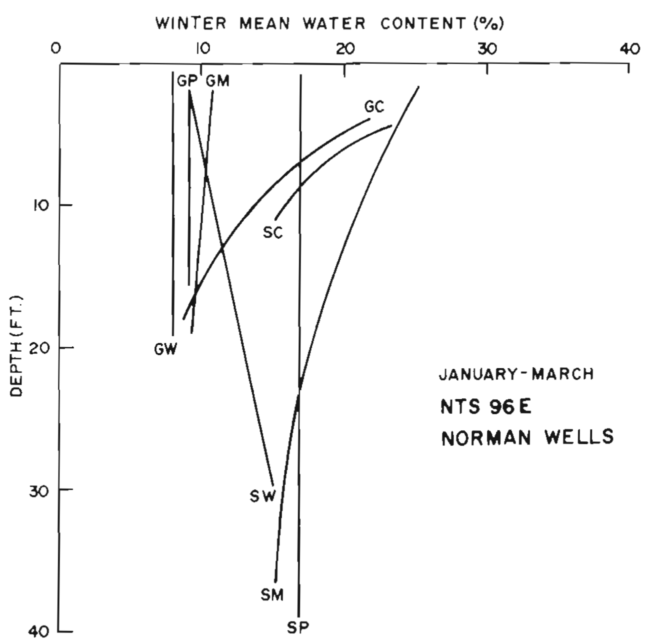

Figure 54.1. Winter mean water content vs. depth for coarse grained soil groups (Norman Wells NTS 96 E).

Data Analysis

A retrieval system using COBOL programing is utilized to retrieve data from the data bank for analysis. Information from a specified National Topographic System (NTS) map-area can be retrieved using the "GET-MAP" program. Selective retrieval on various parameters such as the genetic description and time span is permitted using the "RETRIEVE" program.

Lawrence (1975) reported a relationship existing between the natural soil moisture and depth for various fine grained soil groups in several areas within Mackenzie Valley (Norman Wells 96 E, Fort McPherson-Arctic Red River 106 M and N, and Aklavik 107 B). The water content of a soil is a dominant parameter in thermal calculations. The thermal conductivity, volumetric specific heat, volumetric heat capacity, and volumetric latent heat of both thawed and frozen saturated soil can be expressed as functions of the water content (Kersten, 1952; Nixon and McRoberts, 1973). In this study, special emphasis is placed on understanding the influence of dominant variables such as soil type, water content, and depth so as to predict response for any combination of conditions.

In the first stage of the investigation, a National Topographic System map-area is selected and studied with no preferential selection or refinement of the data being undertaken. Soils are grouped according to the Unified Soil Classification System which permits reliable classification on the basis of relatively few and inexpensive laboratory tests. The "SCATTER" computer program is used to generate a one-page cross-tabulation frequency plot of water content vs. depth for a specified soil group and to calculate the mean water content and the standard deviation at each depth. The mean water content subsequently is plotted against the depth. Particularly noteworthy is the dispersion about the mean which in some cases is considerable.

In the second stage, the data are further grouped in genetic categories and the dispersion about the mean is noticeably reduced. In some cases, the standard deviation is found to be large although most of the data are concentrated near the mean. The large standard deviation is undoubtedly due to the presence of a few data with exceptionally high water contents. These data in general are not truly indicative of the actual water content of the soil group at a particular depth. It is in these cases that preferential selection of data is desirable. A problem arises in deciding

[1]Department of Indian and Northern Affairs.

which data are insignificant. Several criteria, depending on the range of the sample, have been tried satisfactorily and are presented as follows: all data must fall within a specified number of standard deviations of the mean as indicated in Table 54. 1.

Table 54. 1

Number of standard deviations used in the selection of data

Range	Number of standard deviations
Less than 10	4. 0
10 – 30	3. 5
30 – 50	3. 0
50 – 70	2. 5
70 – 90	2. 0
90 – 110	1. 5
110 – 130	1. 0
Greater than 130	0. 5

Consequently, all data which fall outside the specified number of standard deviations are deleted from the calculation of the mean. About 2 per cent of the data have been deleted using these criteria. The refined mean water content is plotted against the depth.

Results and Discussions

Information from boreholes drilled during the period January to March was used to investigate the winter water content to depth relationship for various soil groups. Two NTS map-areas have been studied and some of the results presented in this paper are similar to those reported by Lawrence (1975). The difference is that in Lawrence's paper, the mean values were estimated, while in this paper the mean values were computed.

Norman Wells, NTS 96 E

The Mackenzie Plain in this area comprises deposits of morainal till, alluvial deposits, glaciolacustrine sands, silts, and clays, glaciofluvial sands and gravels, and eskers (Hughes, 1970). The results are illustrated in Figures 54. 1 to 54. 4. It should be noted that lack of data in some soil groups did not permit meaningful analysis, and therefore they did not appear in the plot. From the results, the following observations can be made:

(1) The winter mean water contents of fine grained soils are higher than those of coarse grained soils.

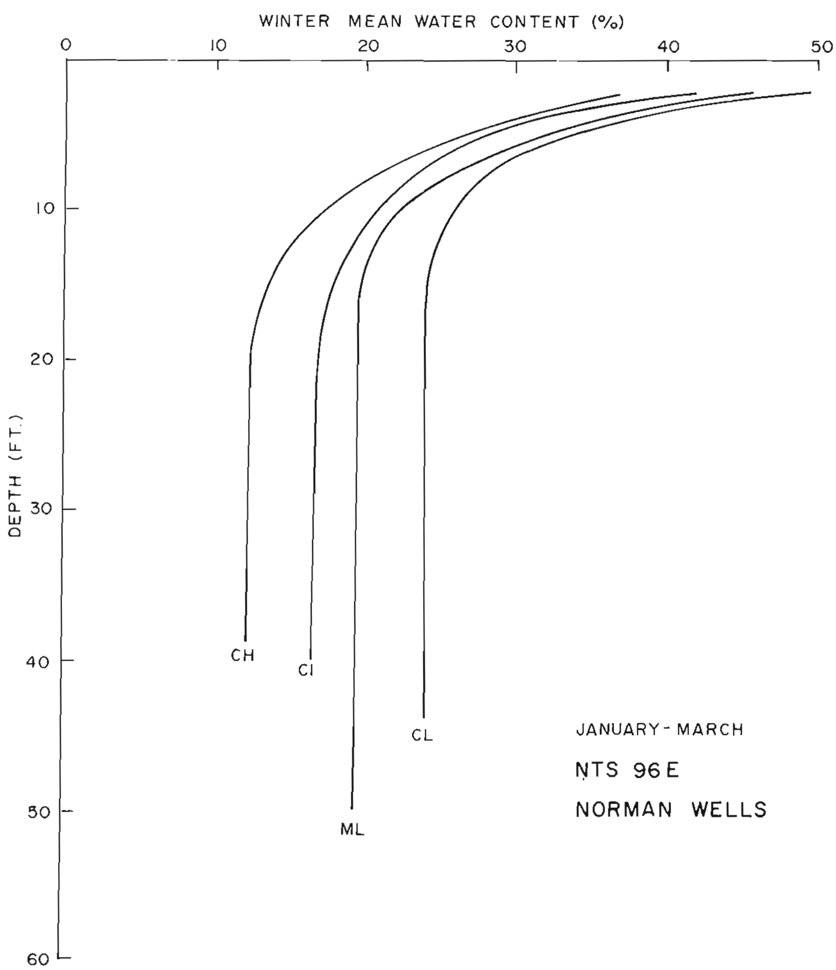

Figure 54. 2.

Winter mean water content vs. depth for fine grained soil groups (Norman Wells NTS 96 E).

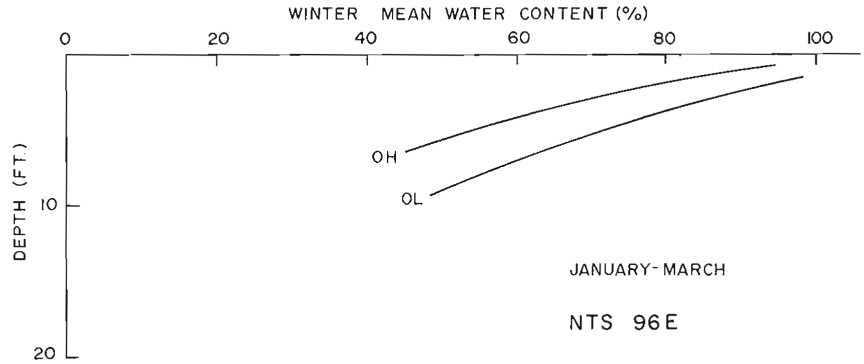

Figure 54. 3.

Winter mean water content vs. depth for organic soil groups (Norman Wells NTS 96 E).

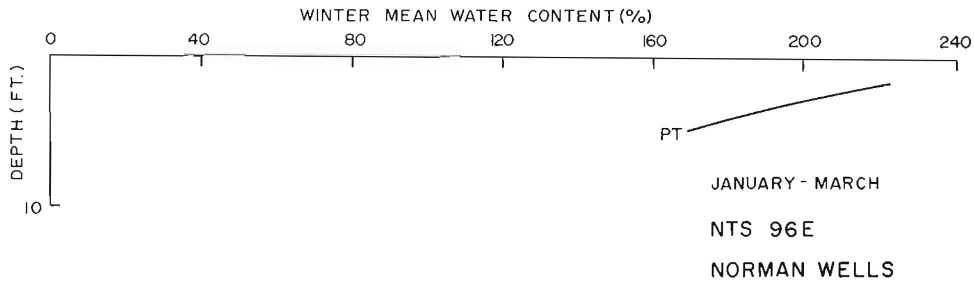

Figure 54. 4.

Winter mean water content vs. depth for peat (Norman Wells NTS 96 E).

(2) For coarse grained soils, the winter mean water contents increase as the amount of fines increase.

(3) The GC, SM, and SC curves show an exponential rise in the winter mean water content towards the surface.

(4) For the fine grained soils and organic soils, the winter mean water contents increase as plasticity decreases.

(5) The winter mean water content curves of fine grained soils show a limiting value at depths greater than about 18 feet. Above this depth the curves show an exponential rise in the winter mean water content towards the surface, while beyond this depth the winter mean water contents are relatively constant.

(6) Peat and organic soils have very high winter mean water contents.

A study of the permafrost and ground ice distribution in this area shows that most coarse grained soils are frozen in winter and contain some visible ice. High ice concentrations are observed only in the upper 6 feet in GM, GC, SM, and SC soils. Most fine grained soils are frozen and contain large concentrations of visible ice (greater than 20 per cent) in the uppermost 18 feet. There appears to be some correlation between winter mean water content and ground ice distribution. With the presence of high ice concentrations, the winter mean water content curve shows an exponential rise towards the surface.

For fine grained soils, alluvial soils always show greater winter mean water content values for a given depth as compared to glaciolacustrine soils, and glaciolacustrine soils always have higher winter mean water contents than morainal soils.

Camsell Bend, NTS 95 J

The Mackenzie Plain in this area contains lacustrine sediments and till (Rutter *et al.*, 1973). From the results (Figs. 54. 5 and 54. 6) the following observations can be made:

(1) The winter mean water contents of fine grained soils are higher than those of coarse grained soils.

(2) For coarse grained soils, the winter mean water contents increase as the amount of fines increase.

(3) The SM and SC curves show an exponential rise in the winter mean water content towards the surface.

(4) The ML and CL curves show an exponential rise in the winter mean water content towards the surface, while the winter mean water contents of CI and CH soils increase with depth.

(5) In the upper 4 feet, CL and ML soils have higher winter mean water contents than CH and CI soils, while at depths below 6 feet, CI and CH soils have higher winter mean water contents than CL and ML soils.

A study of permafrost and ground ice distribution in this area shows that most coarse grained soils are frozen in winter with little visible ice encountered in GW, SW, SM, and SC soils and no high ice concentrations. Most ML and Cl soils are frozen in winter

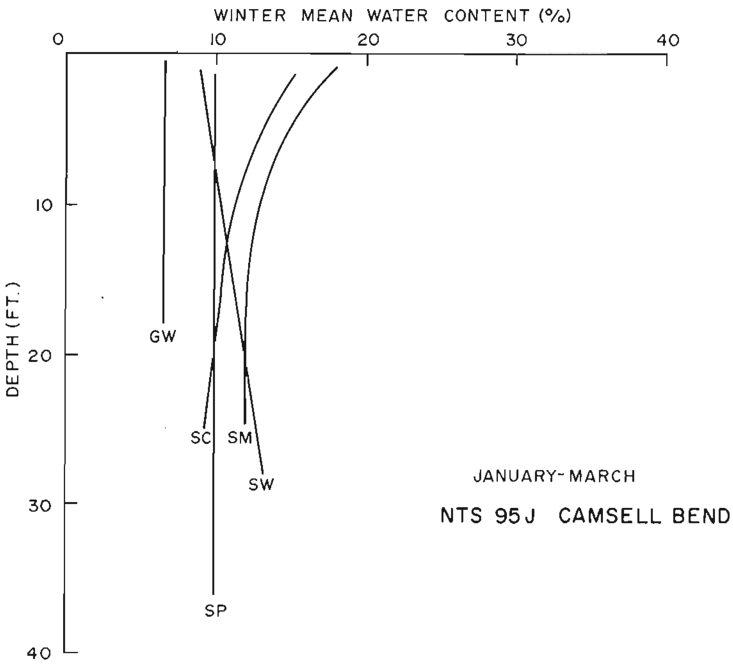

Figure 54. 5. Winter mean water content vs. depth for coarse grained soil groups (Camsell Bend NTS 95 J).

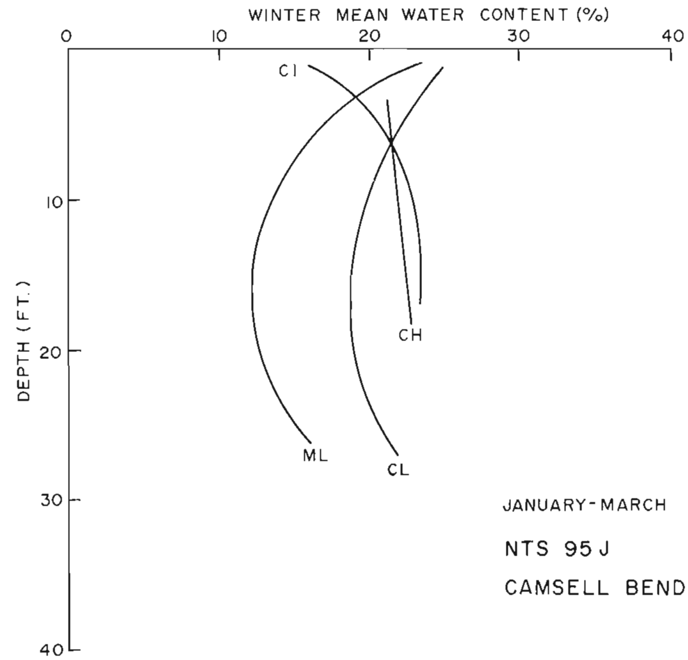

Figure 54. 6. Winter mean water content vs. depth for fine grained soil groups (Camsell Bend NTS 95 J).

with some visible ice in the upper 15 feet, while only less than 20 per cent of CI and CH soils are frozen with some visible ice. There appears to be some correlation between winter mean water content and ground ice distribution. The winter mean water contents of frozen fine grained soils with some visible ice decrease with increasing depth. Where less than 20 per cent of fine grained soils are frozen, the winter mean water contents increase with depth.

For fine grained soils, alluvial soils always show greater winter mean water content values as compared to morainal soils.

Conclusions

At this stage of the study, several conclusions can be drawn:

(1) The winter mean water contents of fine grained soils are higher than those of coarse grained soils.

(2) For coarse grained soils, the winter mean water contents increase as the amount of fines increase.

(3) There appears to be some correlation between the winter mean water content and the ground ice distribution.

(4) Soils in the Norman Wells area have higher winter mean water contents than soils in the Camsell Bend area.

References

Hughes, O. L.
1970: Surficial geology maps; Geol. Surv. Can., Open File 26, NTS 96 C-E, 106 G, H, scale 1: 125 000.

Kersten, M. S.
1952: Thermal properties of soils; Highway Res. Bd., Spec. Rep. 2, p. 161-166.

Lawrence, D. E.
1974a: Geological review of geotechnical data Mackenzie Valley; *in* Report of Activities, Part A, Geol. Surv. Can., Paper 74-1A, p. 281.

1974b: Geological review of geothechnical data Mackenzie Valley; prepared for Mackenzie Highway Environmental Working Group, DINA, 63 p.

1975: Soil moisture relationships, selected map-areas, Mackenzie Valley, N. W. T.; *in* Report of Activities, Part B, Geol. Surv. Can., Paper 75-1B, p. 183.

Nixon, J. F. and McRoberts, E. C.
1973: A study of some factors affecting the thawing of frozen soils; Can. Geotech. J., v. 10, no. 3, p. 439-452.

Rutter, N. W., Boydell, A. N., Minning, G. V., and Netterville, J. A.
1973: Preliminary drafts of surficial geology and geomorphology maps, District of Mackenzie; Geol. Surv. Can., Open File 158, NTS 85 D, E, 95 A, B, G-K, N, O, scale 1: 125 000.

55. PERMAFROST CORING EQUIPMENT

Project 730019

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Development activities related to permafrost coring equipment included planning and co-ordination of activities of outside participants and experimentation in the laboratory. Coring at the involuted hills test site in the vicinity of Tuktoyaktuk, Northwest Territories, from mid-March to early April 1975 and on Somerset and Prince of Wales islands, Northwest Territories during the summer constitutes the field component for the season. Approximately 220 m of cores resulted from these field programs. In addition to samples retained for engineering index tests, some cores were kept frozen and shipped to Ottawa for specific laboratory measurements by T. J. Katsube, Resource Geophysics and Geochemistry Division and P. J. Kurfurst, Terrain Sciences Division.

Inventory control

A need for a systematic maintenance check and inventory control of equipment and supplies was identified due to the accumulation of various items acquired outside Technical Field and Support Services store. A shelf and bin system was installed at 45 Spencer Street. Card files now exist for equipment stored in Ottawa and in field stations. This system facilitates closer control of all aspects of inventory and loans of equipment to field officers or others.

Equipment development

ATV-Drill

A light-weight portable drill, equipped with an extended mast and mounted on an all-terrain vehicle was developed with the purpose of increasing coring production in permafrost terrain. The ATV-drill along with field trials results are described by Veillette and Nixon (1975).

Hydrocyclone

Suspended cuttings in recirculated drilling fluid is recognized as a major obstacle to satisfactory diamond coring of frozen unconsolidated materials. Hydrocyclone units are in common use in the oil drilling industry to desand and desilt drilling muds. An investigation of the adaptability of the hydrocyclone concept to small diameter diamond coring tools began in fall 1974. A working model was built and tested in the laboratory. The results of these tests led to the selection of a commercial hydrocyclone which was tested in the laboratory and in the field during spring 1975. Some problems were identified during these tests but it is felt that the concept is promising and warrants further development.

Stihl 4308 earth auger

A Stihl 4308, hand-held, gas-powered auger adapted to CRREL augers was tested during the summer. The auger is better suited to shallow frozen ground coring than most hand-held augers in current use. The design of a portable ground coring kit, with the Stihl 4308 as a power source, is being considered.

Frozen core containers

Individual core containers were designed for the purpose of storing and preserving frozen cores while in transit from the field to permanent cold storage locations and to minimize disturbance to core samples. A container consists basically of PVC pipe inner and outer tubes sealed at one end, with an insulating air gap between the two tubes. A threaded plastic plug pressing against insulating material and the open ends of the tubes seals the assembly. Containers exist for cores 30 cm in length and diameters of 5.1 and 7.6 cm. Frozen cores inserted in the containers and placed in the coolers can be preserved for several hours at room temperature.

Reference

Veillette, J. J. and Nixon, F. M.
1975: A modified ATV-drill for shallow permafrost coring; *in* Report of Activities, Part C, Geol. Surv. Can., Paper 75-1C, p. 323-324.

A RECONNAISSANCE STUDY OF THE COASTAL PROCESSES ON BANKS ISLAND, DISTRICT OF FRANKLIN

Project 750079

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In conjunction with and as an extension of a study of the surface processes of Banks Island, N. W. T. (Day, 1975), a reconnaissance study of the coasts and coastal processes of Banks Island was undertaken. The study had four objectives:

a. to classify the coastline in terms relevant to the contemporary processes operating on them;

b. to select a number of representative beach profile sites so that their change over the open-water season could be measured and related to, for example, active layer depth;

c. to obtain samples of the beach deposits to enable any change in particle morphology over the summer to be detected. Additionally, sediment samples taken along the shoreline can be used as an indication of the direction of longshore sediment transport; and

d. to make an underwater examination of the nearshore zone at selected beach sites to inspect and, if possible, to measure shallow-water bottom features.

During the 1975 field season, two weeks were devoted to this work, most of which was done early in the season, in mid-June, and at the end, in late August when the camp was closed. At this writing, with the sediment samples still in transit from the field, only preliminary results are given.

The classification judged to be most appropriate for the Banks Island coasts is that of Shepard (1973, p. 111). It is comprehensive and places emphasis on the processes responsible for the present form of the coastline. Shepard classifies coasts according to whether they are shaped primarily by land erosion, subaerial deposition, wave erosion, or marine deposition. Although these are subdivided further in his scheme, only the four classifications have been used in this study. Identification was primarily from the air in the course of helicopter flights between profile sites, but supplementary information from airphotos and hydrographic charts also was used.

Most of the north, east, and south parts of the island from site 9 to site 1 (Fig. 56. 1) have wave-eroded coastlines, in spite of the fact that only along the south coast and the southern half of Prince of Wales Strait is there invariably an open-water season. Along most of the north coast and in some years in northern Prince of Wales Strait, pack or floe ice may remain close to shore over much of the summer, precluding protracted periods of vigorous wave attack. Locally, in the vicinity of large river mouths or in estuaries, the coast is best classified as one of subaerial deposition, but these areas are few. Shorelines shaped predominantly by marine deposition are also rare and are confined mainly to southern Prince of Wales Strait. Here, sediment is transported south along the Strait and accumulates in sediment sinks on the updrift side of promontories such as Cape Treadwell (site 14).

The west coast from site 1 to site 9 is classified almost entirely as one of land erosion. Most of it has been drowned by postglacial submergence and the nearshore zone is shallow and generously endowed with bars, spits, and barrier beaches.

There are 949 miles of coastline on Banks Island. Of these, 474 miles (50 per cent) are classified as wave erosion coasts, 266 miles (28 per cent) as land erosion coasts, 152 miles (16 per cent) are subaerial deposition coasts, and 57 miles (6 per cent) are coasts of marine deposition.

Sixteen beach profile sites were selected around the island (Fig. 56. 1) each was surveyed shortly after breakup and again at the end of the summer season using standard levelling techniques. Active layer depths and surface texture also were measured along each profile. Surface "channel" samples of the foreshore material were taken on each occasion and in some cases a tomato can was used to collect a 11. 8-cm-deep by 10. 0-cm-diameter "core" sample at what was judged

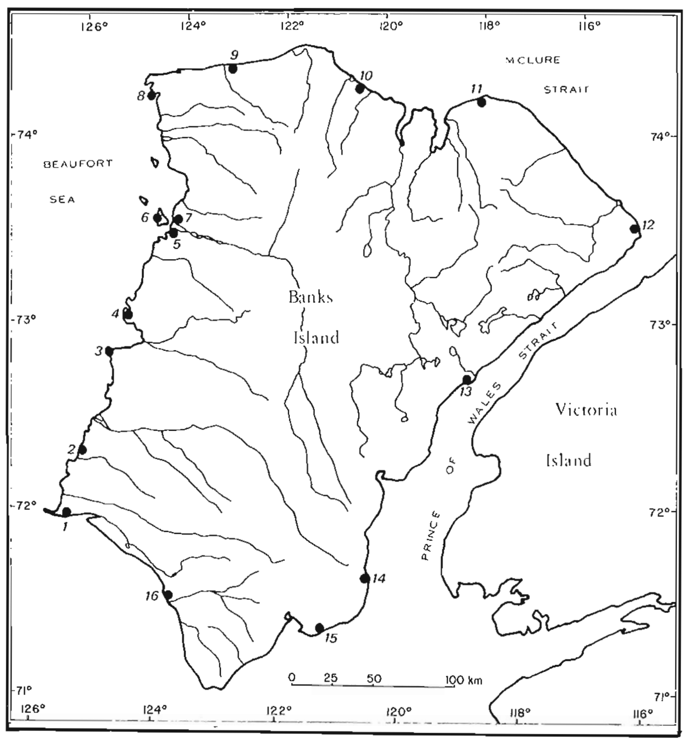

Figure 56. 1. Banks Island showing beach profile sites.

to be the mid-tide elevation. These will be analyzed by standard petrographic techniques with a view to describing the size characteristics of the deposits and the trends in size and size sorting on the foreshore during the summer period.

The profiles plotted for each site was superimposed and volumes of sediment accretion and erosion were measured with a planimeter. The results, which described either a net gain of sediment to the beach face or a net loss, reflect to a large extent the erosional nature, already described, of the coastal zone. Of the sixteen profiles, fourteen are exclusively marine influenced (as opposed to fluvially affected). Of these, nine (sites 2, 3, 4, 6, 9, 11, 12, 15, 16) show a net sediment loss ranging from 0. 08 to 3. 20 m^3 per metre of shoreline, and the remaining five (sites 1, 8, 10, 13, 14) show a net sediment gain ranging between 0. 05 and 1. 02 m^3 per metre of shoreline. These figures are typical of the magnitudes of erosion and accretion found on other western Arctic beaches (*see* e. g., Wiseman *et al.*, 1973).

The thickness of the active layer is an important variable in the coastal system because it defines the lower extremity of mobile sediment on the beach face. Active layer depths were measured along each beach profile at five pace intervals with a 120-cm-long steel probe, each time a profile was surveyed.

The sectional plots of active layers show that with distance normal to the shoreline, the depth of the permafrost table on the backshore is more consistent than that on the foreshore. This dichotomy is conspicuous on most of the profiles and is probable attributable to differences in the subsurface moisture regimes of the two zones. In similar circumstances Taylor and McCann (1974) found anomalously shallow frost table depths on lower foreshores at Radstock Bay, Northwest Territories, and speculate that their results may have been due to their ice auger encountering buried lenses of frozen material rather than the true permafrost horizon (R. B. Taylor, pers. comm.). It also may be true that in some cases removal of sediment by wave action and/or deposition by ice push are the processes responsible for the greater depth variation on the foreshore. Regardless of the processes involved, these kinds of variation raise the question of whether a probe is the best means of detecting the permafrost horizon on an unconsolidated and commonly water-saturated foreshore.

Active layer depths on the Banks Island coasts also increase substantially over the summer, with the rate of increase on the lower foreshores greater in most cases than the rate on their respective backshores. In the absence of results from grain size analyses of the foreshore sediment samples, speculation on the reasons for this would be premature. In this context, however, it is probably worth mentioning that interstitial pore space, a concomitant of grain size, governs the rate of swash and backwash percolation through the sediments, a process shown to have great influence on the stability of beaches in more temperature climates (*see* e. g., Kemp, 1963). It is not known to what extent these relationships are complicated by the added influence of permafrost.

Because the island has considerable latitudinal extent from 71°05'N to 74°34'N, an attempt was made to relate the rate of change in the depth of the active layer to latitude. Preliminary results suggest that the permafrost horizon on the foreshore deepens more rapidly during the summer in the northern latitudes than it does in the south. The active layer depths on the backshore do not show this trend. Grain size data, shortly to be incorporated in the study, will help to verify this relationship by permitting textural variation to be included in the description.

The last work undertaken in the field was a reconnaissance of the shallow-water submarine environment. Four shallow-water dives were made, one on each of the north, east, south, and west coasts. A number of sediment samples were collected and descriptive notes were taken on a diversity of bottom features including ice scour, ripple marks, and nearshore bottom topography.

References

Day, T. J.
1975: Reconnaissance studies of surface processes, Banks Island, District of Franklin; *in* Report of Activities, Part A, Geol. Surv. Can., Paper 75-1A, p. 457.

Kemp, P. H.
1963: A field study of wave action on natural beaches; Int. Assoc. Hydraul. Res., Proc. 10th Cong., p. 131-138.

Shepard, F. P.
1973: Submarine geology; 3rd ed., Harper and Row, New York, 517 p.

Taylor, R. B. and McCann, S. B.
1974: Depth of the frost table on beaches in the Canadian Arctic Archipelago; J. Glaciol., v. 13, no. 68, p. 321-322.

Wiseman, W. J., Coleman, J. M., Gregory, A., Hsu, S. A., Short, A. D., Suhayda, J. N., Walters, C. D., and Wright, L. D.
1973: Alaskan Arctic coastal processes and morphology; Coast Studies Inst., La. State Univ., Tech. Rept. no. 149, 171 p.

57. SAND AND GRAVEL RESOURCES OF KITIMAT, TERRACE, AND PRINCE RUPERT, BRITISH COLUMBIA

Project 750078

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Industrial and urban development is dependent, in part, on the availability of large amounts of sand (0.06 to 2 mm) and gravel (coarser than 2 mm). These materials are constituents of concrete and asphalt and are used extensively by the construction industry in buildings, roads, and as fill.

Because large volumes of construction aggregate must be transported at low cost, the most desired sources are those located close to utilization sites; the latter, in general, are expanding urban areas. Because of their location near cities, however, some of the largest and most accessible sand and gravel deposits are overrun by urban development. This is especially a problem in many parts of British Columbia where both municipalities and major deposits of unconsolidated sediment are restricted to intermontane valleys.

It is thus important that those individuals involved in urban planning reserve suitable sites for aggregate production. Ideally, the deposits should be close to areas of industrial and urban development, should be large in size and comparatively easy to "mine", and should contain a minimum of undesired constituents (such as clay, silt, and large boulders) which increase production costs.

One region in British Columbia which has experienced major urban and industrial expansion is the Skeena River area. Future development probably will be concentrated around the municipalities of Kitimat (54°03'N, 128°38'W), Terrace (54°31'N, 128°36'W), and Prince Rupert (54°19'N, 130°19'W); thus the demand for construction aggregate in these areas will increase.

The extent and geologic setting of sand and gravel deposits near Kitimat, Terrace, and Prince Rupert are the subjects of this report and represent the first results of a larger study of the surficial geology of the Skeena River area.

Sand and Gravel Deposits of the Terrace-Kitimat Corridor

Terrace and Kitimat are located in a major valley within the Coast Mountains which is referred to here as the Terrace-Kitimat corridor. The valley trends north-south, is 5 to 10 km wide in most places, and is crossed at Terrace by Skeena River.

The sand and gravel resources of this area are very large; some individual deposits are larger than 10^9 m^3 (Figs. 57.1 and 57.2). Thick unconsolidated sediments are restricted to elevations below about 300 m in the Terrace-Kitimat corridor, in the valley of Skeena River, and in a few of the larger tributary valleys such as the Zymoetz and Wedeene. Figure 57.2 shows the location and approximate size of the largest sand and gravel deposits in this region. In addition, sand and gravel of unknown thickness underlie the flood plains of the major rivers (e.g., Skeena, Kitimat, and Kitsumkalum). The various deposits are briefly described below in terms of their morphology and genesis (Table 57.1).

The largest deposits, each containing 10^9 to 10^{10} m^3 of sand and gravel, are located in the Terrace-Kitimat corridor from about 6 km north of Terrace to Kitsumkalum Lake (A of Fig. 57.2) and north and south

Figure 57.1.

Gravel platform south of Lakelse Lake (C of Fig. 57.2); view southeast across gently sloping, kettled surface underlain by sandy gravel.

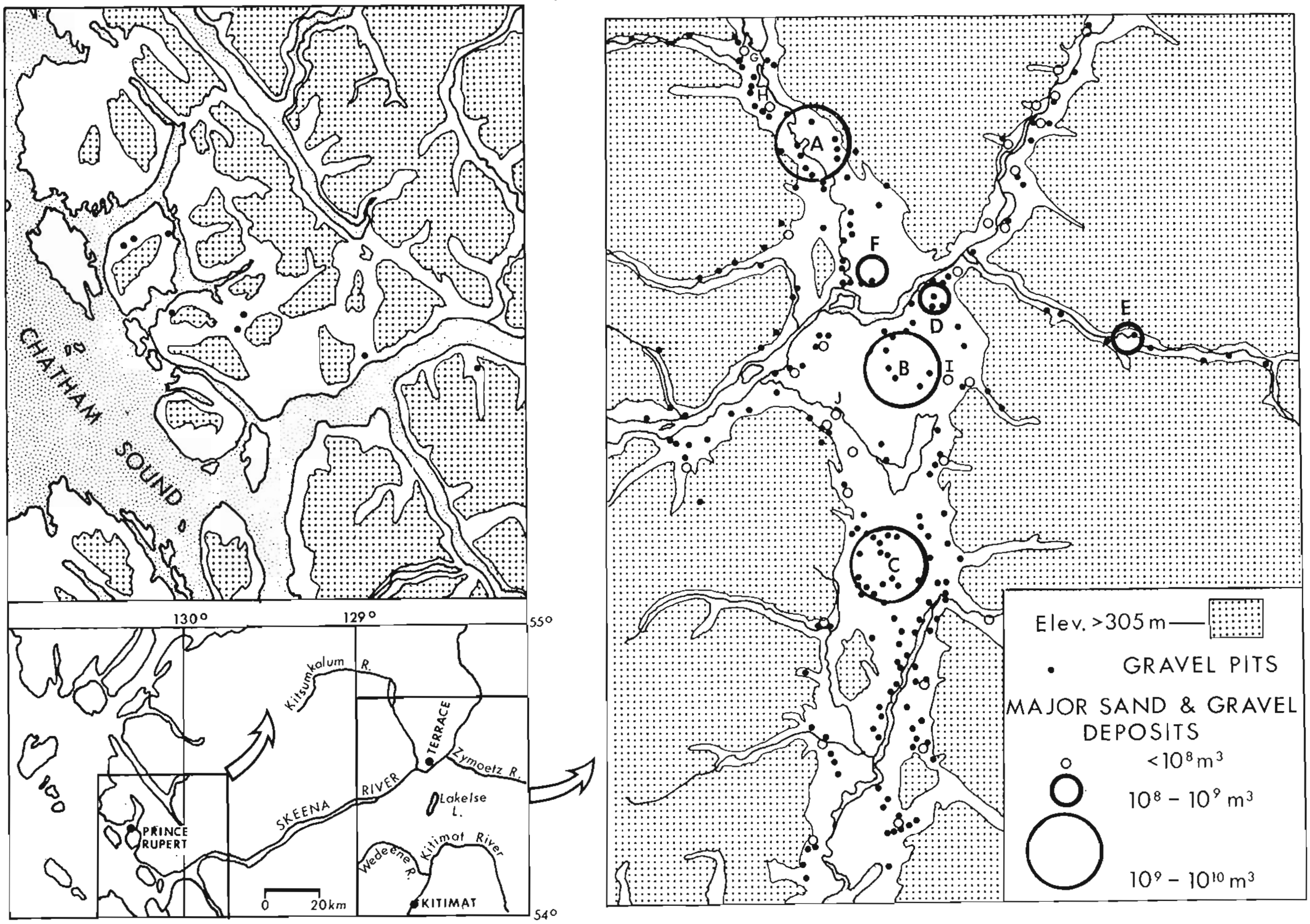

Figure 57. 2. Distribution of sand and gravel deposits and gravel pits, Skeena River area, British Columbia. Coarse stippled areas are higher than 305 m (1000 ft.); fine stippled areas are lakes, rivers, and sea. Lettered deposits are discussed in the text.

of Lakelse Lake (B and C of Fig. 57. 2). A smaller deposit of the same type is shown on Figure 57. 2 at D. The morphologic expression of this group of deposits is that of kettled platforms separated from surrounding areas by scarps up to 150 m high and sloping up to 35 degrees. The two large platforms south of Terrace have not been significantly modified since their formation, whereas the platform north of Terrace has been incised and terraced by the Kitsumkalum River during Holocene time.

The platform sediments, which locally may exceed 150 m in thickness, consist of horizontally stratified sandy gravel overlying inclined sandy gravel and gravelly sand. Representative sediment samples were analyzed for their grain size distributions, and the results are shown in Figure 57. 3[1]. For the seven samples analyzed, gravel constitutes 54 to 91 per cent of the sediment, sand 9 to 44 per cent, and silt and clay 0 to 2 per cent.

The platform-type deposits formed during deglaciation of the Terrace-Kitimat area at the end of the Pleistocene. After the mountainous uplands had become, in large part, ice free and as glaciers in the corridor receded north, arms of the sea entered the area from the west up Skeena River valley and from the south up the main corridor (Duffell and Souther, 1964; Armstrong, 1966). The base of each platform represents a marine delta built into the sea by meltwater flowing from glacier termini during glacier standstills or after minor readvances. These delta foreset beds are overlain by topset beds which directly underlie the platform surfaces and which were deposited in a glaciofluvial environment as the sediment piles were built above sea level.

During deposition of these platform gravels, marine fan deltas formed at the mouths of tributaries to the Skeena Valley and the Terrace-Kitimat corridor. Many of the gravel deposits in Figure 57. 2 with a volume of less than 10^8 m^3 are of this group. These landforms are fan shaped, with apices at the mouths of tributary valleys and surfaces sloping away from the apices into

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[1] Material coarser than -6 ϕ (64 mm) could not be sampled in a representative manner; therefore, the cumulative curves of Figure 57. 3 describe only those portions of the platform sediments finer than -6 ϕ. Cobbles and boulders, in general, are only a small fraction of the total sediments.

Table 57. 1

Sand and gravel deposits of Terrace, Kitimat, and Prince Rupert, British Columbia

Type of deposit	Size of largest deposit (m^3)	
	Terrace-Kitimat	Prince Rupert
Platform[1]	10^9-10^{10}	None
Fan delta[2]	10^7-10^8	None
Terraced valley fill[3]	10^8-10^9	None
Drift veneer and blanket	$<10^6$	$<10^6$
Fluvial fan[4]	10^7-10^8	None
Colluvial veneer and blanket[4]	$<10^6$	$<10^6$
Flood plain[4]	$>10^8$	None

[1] Late Pleistocene marine delta — outwash plain in Terrace-Kitimat corridor.

[2] Late Pleistocene marine delta – fluvial fan at mouth of tributary valley.

[3] Late Pleistocene valley train incised and terraced during Holocene time.

[4] Holocene sediments.

the main valleys. Each fan is bordered by a scarp sloping up to 35 degrees towards the main valley. Because streams and rivers flowing out of the tributary valleys have cut through the fan deltas, only remnants of the original depositional landforms are preserved.

The sediments of this group of deposits have grain size distributions similar to those shown in Figure 57. 3. Although much smaller than the platforms in the Terrace-Kitimat corridor, the fan deltas are suitable for major aggregate production because of the ease with which they can be mined.

Perhaps related in origin to the platform deposits are fills in a few major tributary valleys such as that of the Zymoetz River (E of Fig. 57. 2). These fills consist of glaciofluvial and deltaic sediments deposited in long, comparatively narrow valleys during deglaciation of the area. The fill in the Zymoetz River valley was extensively dissected and terraced during the Holocene, but there remains a large volume of sand and gravel, estimated to be between 10^8 and 10^9 m^3.

Glaciofluvial and fluvial gravels, in many places with a veneer or blanket of sand, underlie terraces flanking the Skeena River. Most of these terraces formed after the sea had withdrawn from the area; however one just north of Terrace (F of Fig. 57. 2) is capped by silt and clay which may be marine in origin. Sand and gravel underlying this surface total 10^8 to 10^9 m^3.

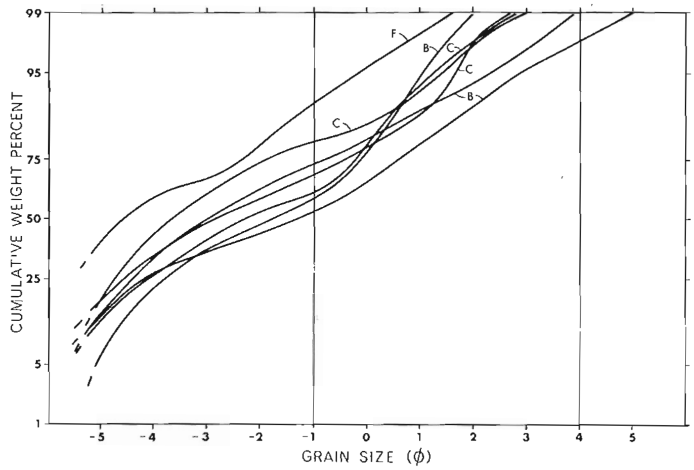

Figure 57. 3. Cumulative frequency curves of gravel samples collected in the Terrace-Kitimat corridor. Letters indicate deposits from which samples were collected (Fig. 57. 2).

The deposits described above presently supply most of the aggregate used by the construction industry in the Terrace-Kitimat area. The deposits, taken together, represent a large volume of sand and gravel, of the order of 10^{10} m^3. In addition, presently active fans and flood plains are underlain by sand and gravel. The larger contemporary fans are shown in Figure 57. 2 (G, H, I, and J). Each probably contains more than 10^7 m^3 of sand and gravel; however volumes are difficult to estimate because the thickness of fan sediments are not known. Like the late Pleistocene fan deltas, the Holocene fans are found at the mouths of tributary valleys; commonly the latter are inset into the former. Small gravel pits also have been opened on the flood plains of the Kitimat, Skeena, and Kitsumkalum rivers. Because of high water tables, however, these deposits probably will not be extensively developed.

Sand and Gravel of the Prince Rupert Area

Prince Rupert is located on the Pacific Coast of Canada approximately 25 km northwest of the mouth of Skeena River (Fig. 57. 2). The area is mountainous and is extensively dissected by fiords. Skeena River is confined by precipitous valley walls and is tidal in its lower reaches.

Although Prince Rupert is only about 120 km from Terrace, the sand and gravel resources of the two areas are completely different (Table 57. 1). Whereas Terrace possesses abundant deposits of unconsolidated sediment laid down at the close of the last glaciation, Prince Rupert has no sand or gravel deposits of importance (Fig. 57. 2). Peat overlies bedrock throughout much of the Prince Rupert area, and glacial deposits, in general, occur as a patchy veneer. A few gravel pits have been opened in pockets of till and outwash, in raised beach deposits, and in colluvium mantling steep slopes; but the operations are small in comparison to those near

Terrace and Kitimat, the deposits are difficult to work, and the product is of low quality. It is likely that there are no terrestrial gravel deposits in the Prince Rupert area which exceed a volume of 10^6 m^3; most known deposits are much smaller.

Most of the aggregate presently used at Prince Rupert is dredged west of the city from the sea floor. It is probable that marine sediments in Chatham Sound, Dixon Entrance, and Hecate Strait will become increasingly important as sources of sand and gravel for the Prince Rupert area. For example, the sea floor in much of Dixon Entrance (54 to 55°N, 131 to 134°W) is gravelly (Mathews, 1958). Alternatively, aggregate could be produced at Terrace and transported by rail to Prince Rupert, although at present this is not economically feasible.

Summary

Sand and gravel deposits which exceed 10^{10} m^3 in total volume exist in the vicinity of Kitimat and Terrace, British Columbia. These deposits were laid down during deglaciation of the area at the end of the Pleistocene and include: (1) large platforms in the Terrace-Kitimat corridor which are combined marine deltas and outwash plains, (2) marine fan deltas at the mouths of tributary valleys, and (3) dissected glaciofluvial and fluvial valley fills. Presently active alluvial fans and flood plains are also underlain by sand and gravel.

The abundance of sand and gravel in the Terrace-Kitimat area contrasts with the rarity of these materials at Prince Rupert which is only 120 km away. Aggregate in the Prince Rupert area presently is obtained west of the city from the sea floor. Sand and gravel deposits occurring on land are all smaller than 10^6 m^3 and include pockets of till and ice-contact sediments, raised beach deposits, and colluvial covers on steep slopes.

References

Armstrong, J.E.
1966: Glacial studies, Kitimat-Terrace area; *in* Report of Activities, Geol. Surv. Can., Paper 66-1, p. 50.

Duffell, S. and Souther, J.G.
1964: Geology of Terrace map-area, British Columbia (103 I E½); Geol. Surv. Can., Mem. 329, 117 p.

Mathews, W.H.
1958: Underwater gravel deposits of Dixon Entrance; unpubl. rept., Inst. Oceanogr., Univ. British Columbia, Vancouver, 6 p.

Project 740046

P. J. Kurfurst
Terrain Sciences Division

This project was initiated in spring 1974 (Kurfurst, 1975) as an extension of Project 710077, which was located in the Mackenzie Valley (Heginbottom and Kurfurst, 1972, 1973; Kurfurst, 1974). The main objectives of the field program were to add to the current knowledge of:

(1) terrain and permafrost conditions and ground ice content in various surficial materials;

(2) terrain sensitivity to traffic, engineering, and construction activities; and

(3) terrain susceptibility to man-made and natural disturbances.

A nine-week field program was carried out in the District of Franklin, N. W. T., during summer 1975 for the Environmental-Social Program (Terrestrial Environmental Project TEP-3) and as part of the Geological Survey's continuing program of acquiring geological, geomorphological, and engineering information on northern terrain. Two groups, with P. J. Kurfurst and J. Veillette (Project 730019) as party chiefs, were employed to cover Bathurst-Cornwallis and Somerset-Prince of Wales islands respectively. Both groups worked closely with the terrain mapping teams led by D. M. Barnett (Project 740067) and J. A. Netterville (Project 750071) and operated from their main base camps.

Objectives of the project were achieved by general mapping and sampling of all terrain units and geological formations and by detailed studies of man-made structures such as airfields, access roads, old oil well sites, abandoned camp sites, and related activities. Other features such as unstable slopes, flow and retrogressive slides, ice wedges and ice polygons and their size and distribution, peat, and ice-rich deposits also were studied in detail.

Nineteen sites, fifteen on Bathurst Island and four on Cornwallis Island, were selected and studied in detail to cover the wide range of surficial and subsurficial materials. Selection of sites was based on aerial photography evaluation and helicopter reconnaissance flights. General location of sites is shown in Figure 58. 1.

A hand-held auger drill with a modified CRREL core barrel was used to core to depths of 3 m. A total of 113 holes were drilled varying in depth from 0. 5 to 2. 4 m. Some 210 samples were collected and shipped back for further laboratory testing to determine their

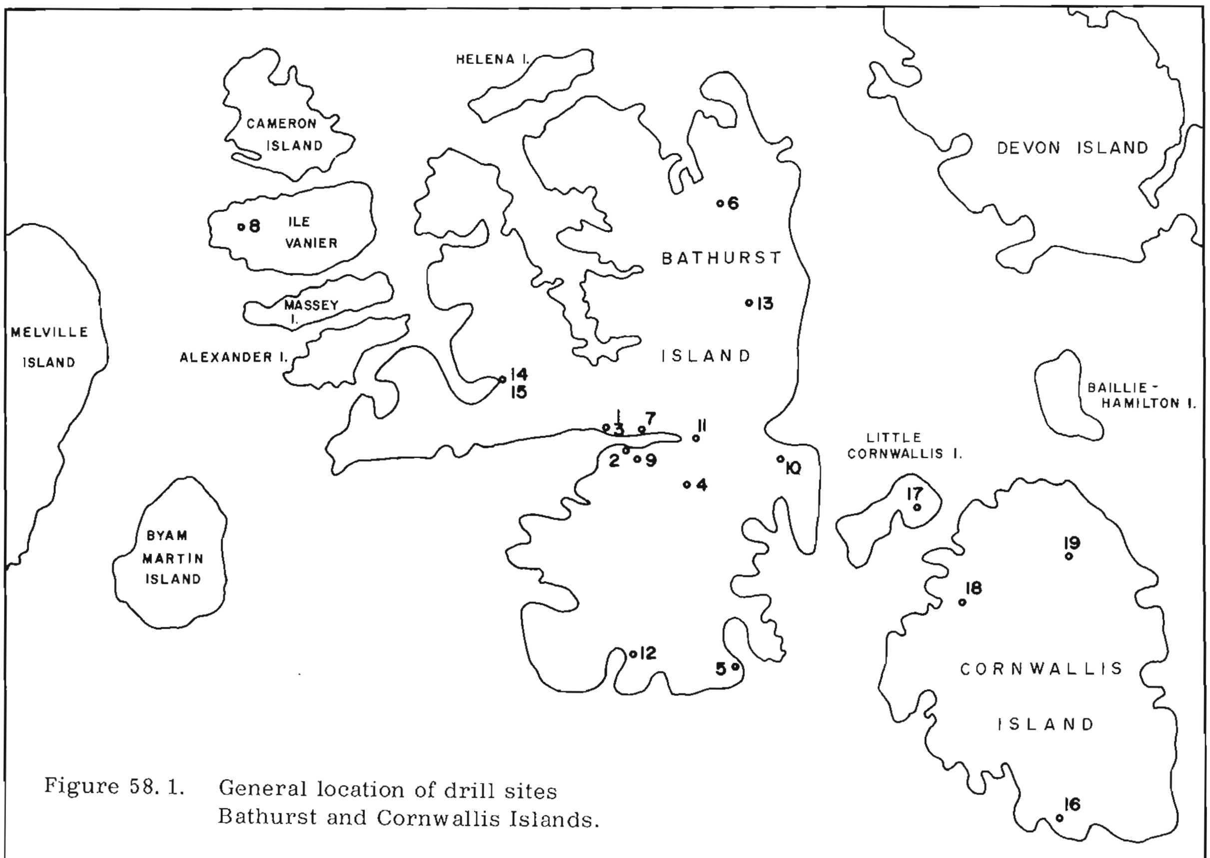

Figure 58. 1. General location of drill sites Bathurst and Cornwallis Islands.

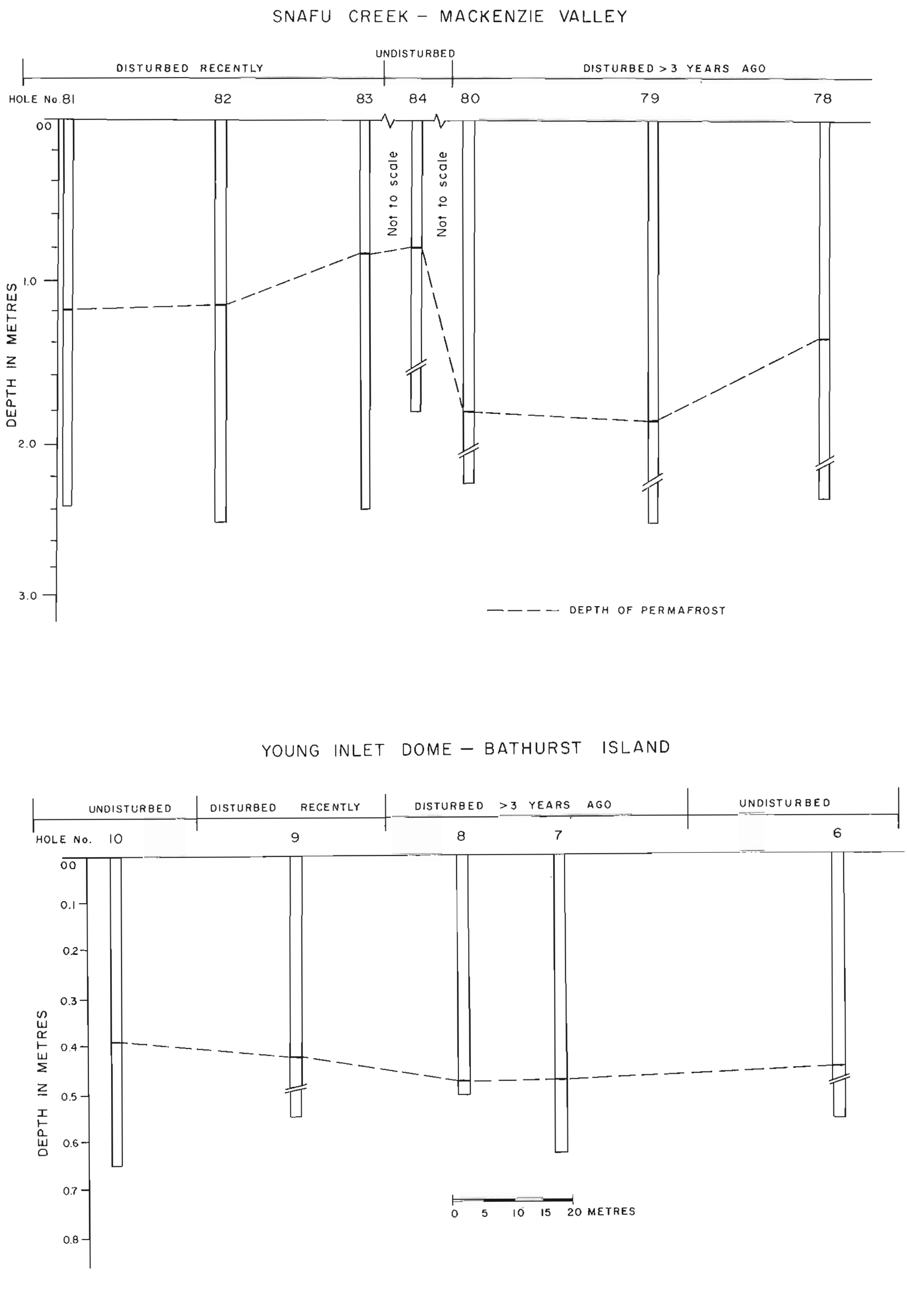

Figure 58. 2. Permafrost degradation due to disturbance.

engineering index properties such as natural moisture content, Atterberg limits, and sand-silt-clay ratio.

Two similar profiles from the Mackenzie Valley and the Arctic Islands are illustrated in Figure 58.2. Both profiles show the original permafrost table in the undisturbed areas and permafrost degradation caused by terrain disturbance inflicted recently and more than three years before investigation. In the Mackenzie Valley there is a significant increase in depth of thaw in the disturbed areas, and rate of permafrost degradation seems to accelerate after three or more years. In the Arctic Islands, however, the process of permafrost degradation due to disturbance seems to be almost negligible and its rate is very slow.

No final conclusions can be drawn at this stage, but preliminary field observations and results of the aerial photography study seem to indicate that:

(1) initial disturbance is more easily inflicted on terrain in the Arctic Islands than in the Mackenzie Valley;

(2) rate of terrain degradation due to man-made and natural disturbance is much slower in the Arctic Islands than in the Mackenzie Valley; and

(3) after minor damage terrain "bounces back" to its original state much slower or not at all in the Arctic Islands, while this process of recuperation is much faster in the Mackenzie Valley.

The difference in terrain behaviour is probably due to lower than annual air and ground temperatures, shorter summers and hence thinner active layer, lack of vegetation, and presence of thin veneer overlying bedrock in the Arctic Islands.

An extensive study of slope stability and soil movement on the slopes along the proposed Gas Arctic Pipeline Corridor was conducted on Bathurst Island. It was found that numerous but minor flow slides occurred mainly in unconsolidated sediments situated below the marine limit and overlying limestones and siltstones of the Bird and Eids formations of middle Devonian age. Where these unconsolidated sediments are situated above marine limit and within the Bird and Eids formations, however, no flow slides occurred. Also where these sediments overlie more competent bedrock of other formations, either above or below the marine limit, no flow slides occurred. This seems to indicate that the higher clay content of marine sediments overlying less competent bedrock makes them more susceptible to sliding even on gentle slopes.

References

Heginbottom, J.A. and Kurfurst, P.J.

1972: Terrain sensitivity evaluation and mapping – Mackenzie Valley Transportation Corridor; *in* Report of Activities, Part A, Geol. Surv. Can., Paper 72-1A, p. 145-146.

1973: Terrain sensitivity evaluation and mapping – Mackenzie Valley Transportation Corridor; *in* Report of Activities, Part A, Geol. Surv. Can., Paper 73-1A, p. 226-229.

Kurfurst, P.J.

1974: Terrain sensitivity evaluation and mapping – Mackenzie Valley Transportation Corridor; *in* Report of Activities, Part A, Geol. Surv. Can., Paper 74-1A, p. 279.

1975: Assessment of terrain performance in the Mackenzie Valley and the Arctic Islands; *in* Report of Activities, Part B, Geol. Surv. Can., Paper 75-1B, p. 97-98.

59. ASSESSMENT OF TERRAIN PERFORMANCE, SOMERSET AND PRINCE OF WALES ISLANDS, DISTRICT OF FRANKLIN

Project 740046

J. Veillette
Terrain Sciences Division

A nine-week field program was carried out in conjunction with surficial geology mapping (Netterville *et al.*, 1976) in the District of Franklin, Northwest Territories during summer 1975, with the purpose of assessing terrain performance on Somerset and Prince of Wales islands. This project is under the general direction of P. J. Kurfurst, and the main objectives of the program are listed by Kurfurst (1976).

Most field work was confined to Somerset Island due to time and weather constraints. Six main sites, five on Somerset Island and one on Prince of Wales Island, were selected for detailed ground based investigation on the basis of initial aerial photograph interpretation or helicopter reconnaissance flights. The sites were chosen to be representative of a wide range of sensitive terrain. Temporary camps were set up at each site for periods of four to six days. Emphasis was put on the detection and sampling of ice-rich deposits. Additional geophysical surveys using seismic refraction, electrical resistivity, and gamma-gamma density techniques were conducted at each site in the latter part of the season by a field crew under the direction of J. A. Hunter, Resource Geophysics and Geochemistry Division. Borehole and other data will be correlated with the geophysical results.

Studies were conducted at a number of other localities in addition to these six main sites.

Field methods of investigation included examination and description of exposed sections, gullies, and naturally disturbed areas. A water jet occasionally was used for washing off poorly exposed sections and trenching. Coring operations, employing a hand-held, gas-powered Stihl 4308 auger, were carried out at 64 sites. These holes ranged in depth from 1 to 4.5 m with an average depth of 2.1 m. A Kango air hammer was used to obtain shallow subsurface data in materials where a high stone content prevented adequate coring by other methods. Some 340 borehole samples and 170 Kango hammer test pit samples were collected for natural water (ice) content determinations and other soil index engineering tests. Within regions 8, 8a, 10, and 10a on Somerset Island (*see* Netterville *et al.*, 1976, Fig. 31.1) the detailed investigations in general were concerned with:

- ice content of marine silts and sands;
- occurrences of massive ground ice;
- retrogressive flow slides, gullying, and other erosional features;
- ice-wedge size and distribution;
- wetlands and peat deposits; and
- vertical water content variations in Eureka Sound Formation deposits.

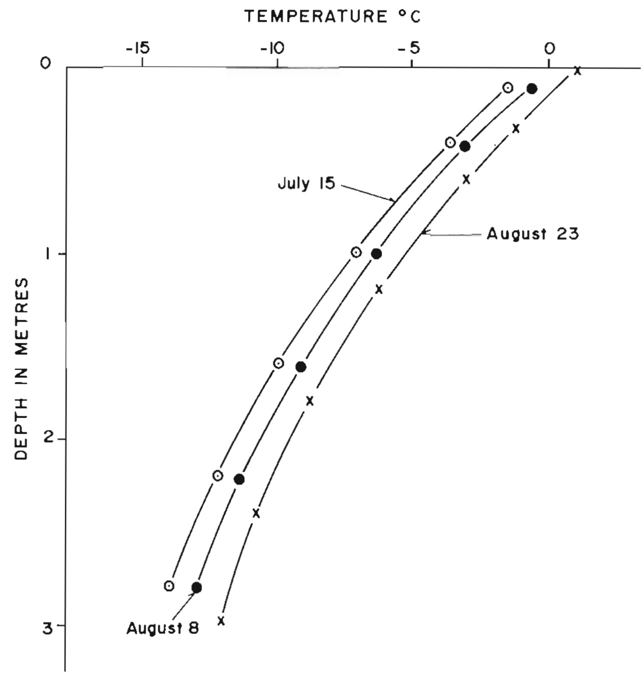

Figure 59.1. Vertical ground temperature profiles, Somerset Island, N.W.T.

In addition, ground temperature profiles to depths of 3 m were recorded at three locations on Somerset Island. Figure 59.1 shows three vertical ground temperature profiles in ice-rich silts. The July 15 and August 8 readings are from the same borehole in the Aston Bay area. The August 23 reading was recorded in the Cunningham Inlet area. Profiles of this type are associated with the texture of the material and assist in the interpretation of geophysical data.

Netterville *et al.* (1976) divided the area into 31 different regions distinguished by differences in texture, relief, slope, and origin of materials. It is outside the scope of this report to comment on each of the regions identified; however, field observations make it possible to group these into the main categories: (1) a relatively stable, predominantly bedrock controlled, low ground-ice content category (Somerset Island regions 1 to 7); although bedrock is exposed in most regions, a veneer of colluvium, glacial drift, or felsenmeer mantles most of these regions; and (2) a more sensitive, predominantly fine textured, and ice-rich category; these regions consist of marine silts and sands (Somerset Island regions 8, 8a, and 8b) and unconsolidated to semi-consolidated Eureka Sound sands (Somerset Island regions 10 and 10a).

On Prince of Wales Island ground investigation was restricted to the Back Bay area (Netterville *et al.*, 1976, Fig. 31.1, region 3a) where one detailed site was located. On Somerset Island four detailed sites were located in deposits of regions 8, 8a, 10, and 10a, and one was in a marine basin near Aston Bay. Kango hammer test pits provided most of the subsurface information for regions 1 to 7 on Somerset Island.

Detailed results based on laboratory analyses of samples, borehole logs description, and other observations will be available in a future report.

References

Netterville, J.A., Dyke, A.S., Thomas, R.D., and Drabinsky, K.A.
1976: Terrain inventory and Quaternary geology, Somerset, Prince of Wales, and adjacent islands; *in* Report of Activities, Part A, Geol. Surv. Can., Paper 76-1A, report 31.

Kurfurst, P.J.
1976: Assessment of terrain performance in the Mackenzie valley and the Arctic Islands; *in* Report of Activities, Part A, Geol. Surv. Can., Paper 76-1A, report 58.

PRELIMINARY RESULTS OF TERRAIN MAPPING AND BASE METAL ANALYSIS OF TILL IN THE RED INDIAN LAKE AND GANDER LAKE MAP-AREAS OF CENTRAL NEWFOUNDLAND

Project 740072

D. R. Grant and C. M. Tucker
Terrain Sciences Division

Surficial geological mapping was carried out along all passable roads in the Red Indian Lake and Gander Lake map-areas of central Newfoundland, with follow-up studies on Burin Peninsula (Grant, 1975c). Observations of the texture and lithology of glacial deposits, particularly till, together with measurements of direction and sequence of ice movement, will be used to interpret the distribution of terrain types delineated by airphoto interpretation. Reconnaissance-style maps at 1: 50 000 scale will be put on open file as they are compiled. This information is designed for application to the search for mineral deposits and to the siting of transportation corridors in an area where bedrock is extensively and thickly covered with glacial debris.

Red Indian Lake map-area

A preliminary reconnaissance airphoto interpretation of the Red Indian Lake map-area has been published (Grant, 1975b). Recent ground checking reveals, however, that the composition of till cover and the sequence of its emplacement is considerably more complex than might be inferred from a cursory examination of the regular pattern of drift lineations diverging from a former ice divide interpreted as trending diagonally

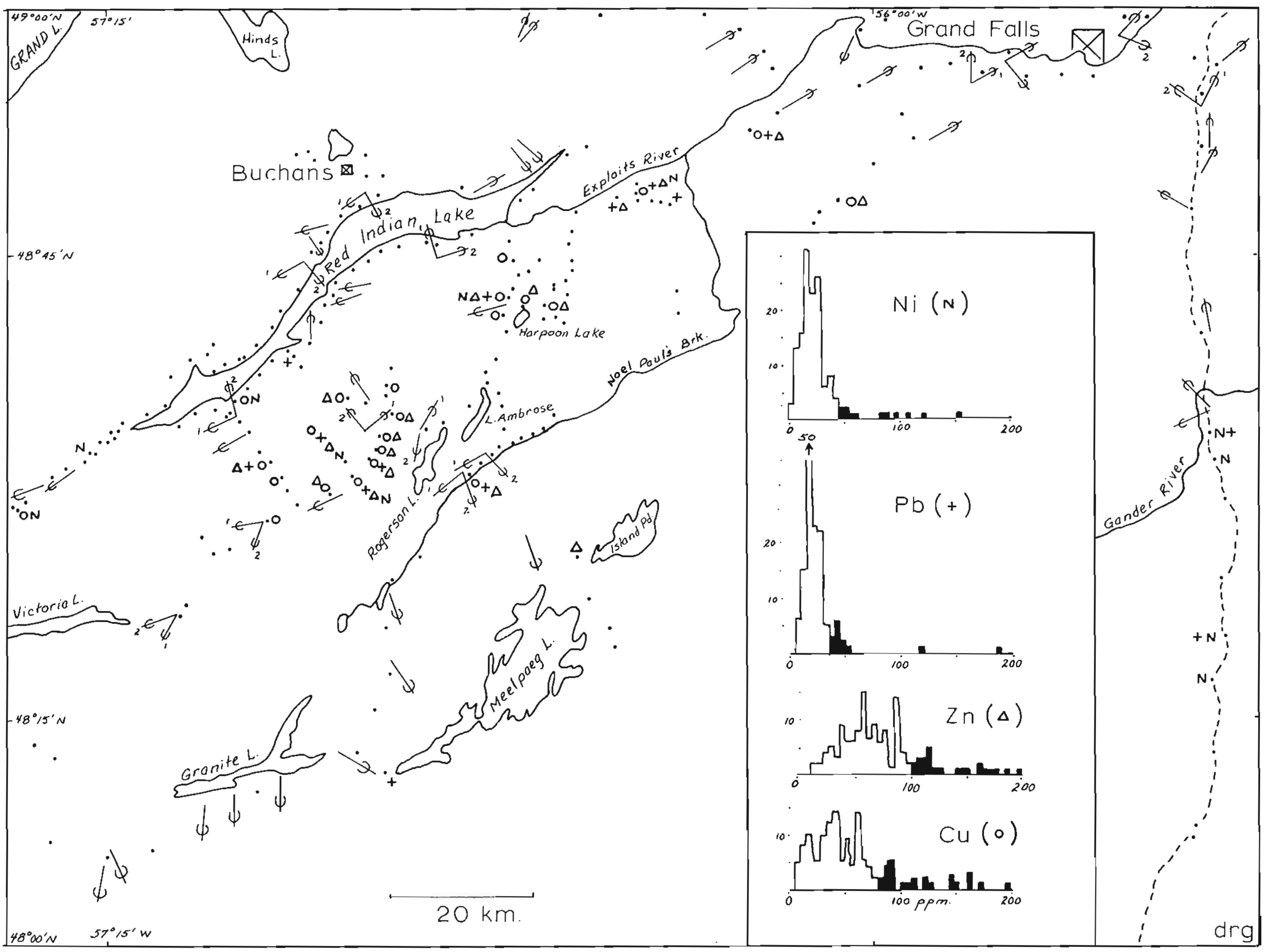

Figure 60. 1. Ice-flow indicators and base-metal anomalies in till, central Newfoundland. (Anomalous values plotted represent the shaded portion of histograms.)

across the area. Notably, most tills appear to be of mixed lithology and varied provenance. While part of the heterogeneity is due to the juxtaposition of different rock types, most of the mixing and variability is now regarded as the result of successive ice flows from very divergent directions. This is borne out by the frequent occurrence of outcrops showing intersecting striated stossed facets oriented in two, commonly perpendicular directions (Fig. 60.1). Although the relative ages could be determined in most cases, the pattern over the area as a whole cannot be resolved into a limited number of distinct regimes. For example in the central part, an early flow moved southwestward and was succeeded by flow either to the northwest or southeast, but the timing of the latter two is unknown. Later movements appear to converge on Red Indian Lake, but this contradicts the conclusion, based on glacial deposits and features, that the lake basin was the locus of a radiating remnantal ice mass (Grant, 1974). Similarly, in the Grand Falls area the dominant flow was northeastward towards the coast, and while this is diametrically opposed to movement in the interior, it is moreover crossed, at right angles, by later flow both to the northwest and to the southeast. One of these later flows may be the effect of a late ice mass in the New Bay Pond area and the other the result of drawdown into Exploits Valley. In summary, the numerous unevenly distributed, somewhat contradictory exposures of directional ice-flow indicators show that several distinct and vigorous pulses of ice-flow affected the area. These are variously attributable to the growth and shifting of local ice sheets, to the inundation by ice sheets from outside the area, and to phases that predate the last glaciation (Grant, 1975a). Obviously these initial findings only vaguely outline general relationships; more ice-flow indicators are needed in the unmapped areas, supported by correlations with till sheets of a given provenance. As a result of this erratic ice-flow behaviour much difficulty should be expected in the search for bedrock sources of geochemical anomalies revealed by activities of the private sector.

Base metal anomalies have been brought to light as a result of the current till sampling program. The minus 200-mesh fraction of all granular materials was analyzed spectroscopically for Cu, Pb, Zn, and Ni under the trace element program of W. W. Shilts (Geol. Surv. Can.). Histograms of the values in parts per million for each element are shown in the inset of Figure 60.1. The polymodal clustering of values towards the zero end of the histogram is taken to represent a composite background attributable to several lithologic terranes. The upper limit of the background range was chosen arbitrarily by inspection, and the remaining anomalous or excessive values (shaded on the histograms) were plotted on the map by means of symbols. The cutoff point for copper was 80 ppm, zinc 100 ppm, lead 35 ppm, and nickel 45 ppm. By this admittedly crude method of sampling and discrimination, some fairly interesting distributions emerge. A pronounced group of higher than average concentrations of all four metals occurs northwest of Rogerson Lake. The area is underlain both by Silurian basic volcanics and by fine grained Ordovician clastics (Williams, 1970). The displacement of the parent till type from its bedrock source is not yet clear from the pebble analyses. A second concentration occurs in the vicinity of Harpoon Lake near a Devonian diorite-granodiorite intrusion in the fine grained clastic material. Other occurrences are situated at the ends of Red Indian Lake. Nickel and lead anomalies occur in tills along the Bay D'Espoir highway. One near the Northwest Gander River crossing overlies and extends down-ice from a mineralized ultrabasic plug (Anderson and Williams, 1970). Another farther south near Berry Hill Pond is underlain by Ordovician and Silurian slates; however, it may be related to a 10-km-wide ultrabasic pluton stiuated 20 km to the northwest up-ice along the trend of a drumlin field. No doubt some of the anomalies are essentially in place near source, but others may originate in transported till. For this reason, it is judged essential that future geochemical exploration work in central Newfoundland carefully analyze till variations and the responsible ice movements when interpreting anomalies.

Gander Lake map-area

A few general remarks may be made regarding surficial features in the Gander Lake map-area as they apply to drift prospecting. Virtually the entire area is till covered except for a rugged granitic area in the southeast corner. Outcrops are found mainly along larger rivers and occur sporadically throughout the interior. The terrain is characterized by broad elliptical undulations, several kilometres across with a relief of 70 to 190 m, which reflect the bedrock surface. These are believed to be the product of repeated glacial breaching of preglacial drainage divides. Differences in bedrock structure and lithology are manifested only vaguely in the topography due to intense glacial scour and complete mantling of the surface with till. This cover is nearly everywhere organized onto glacial lineations which mainly take the form of flutings; crag-and-tail hills and true drumlins are situated nearer the more rugged sloping coastal zone. Common features on the upland near the break in slope to the coast are reversed crag-and-tail hills which are recognized as a ramp of till on the stoss side of a plucked rock pinnacle. The flutings, which are remarkably parallel over long distances, are grouped into three abutting domains that trend northerly, easterly, and southerly from a line of divergence thought to mark an ice divide, which may be traced along an arcuate course from the centre of the west margin to the south margin near Meta Pond. This simple divergent pattern of ice flow may be the result of conditions that prevailed during the last glacial maximum when the island supported an independent ice cap. One might assume that the main course of drift transport is marked by these flutings, but on closer inspection two other ice-moulded till forms that trend at marked angles to the main fabric are evident. In one case these are narrower, shorter, superimposed flutes and crag-and-tail forms, obviously produced by slightly re-oriented flow during recession. Fine striae and small eskers trend parallel to the youngest fluting. Elsewhere the slender fluting seems to be superimposed on much larger elliptical till hills which could be relict from either the advance phase of the last glaciation or a previous glaciation.

Two related aspects may be mentioned. Firstly, the younger weak fluting has a complex of several morainal belts of ice-marginal features associated with it. These have been traced only over a quarter of the area but appear to be distributed along the southern margin of the plateau inland of the coastal break in slope. They curve to the north behind Clarenville and may join with several classic end moraine features discovered between Gambo and Terra Nova ponds. These ice-marginal features are probably younger than the glaciodeltaic complexes formed around the head of Notre Dame Bay ca. 12 000 years ago. Secondly, the pattern of disorganized drainage and the distribution of bedrock hills suggest numerous possible instances of buried valleys, developed in the Gander Lake Group (Jenness, 1963), of which two prominent examples may be cited. The upper reaches of Bay du Nord River and Terra Nova River appear to flow over the largely filled extension of the valley occupied by Gambo Pond and Freshwater Bay. Similarly, North Terra Nova River, Lake St. John, and Kepenkeck Lake probably mark the main course of Terra Nova River valley.

References

Anderson, F.D. and Williams, H.
1970: Gander Lake (west half), Newfoundland; Geol. Surv. Can., Map 1195A.

Grant, D.R.
1974: Prospecting in Newfoundland and the theory of multiple shrinking ice caps; *in* Report of Activities, Part B, Geol. Surv. Can., Paper 74-1B, p. 215-216.

1975a: The contrasting styles of Late-Wisconsinan Laurentide and Appalachian glaciation: New England and the Atlantic Provinces; Geol. Soc. Am. Abstr., v. 7, no. 1, p. 66.

1975b: Surficial geology of Red Indian Lake map-area, Newfoundland – a preliminary interpretation; *in* Report of Activities, Part B, Geol. Surv. Can., Paper 75-1B, p. 111-112.

1975c: Glacial features of the Hermitage-Burin Peninsula areas of Newfoundland; *in* Report of Activities, Part C, Geol. Surv. Can., Paper 75-1C, p. 333-334.

Jenness, S.E.
1963: Terra Nova (Gander Lake, east-half), Newfoundland; Geol. Surv. Can., Map 1129A.

Williams, H.
1970: Red Indian Lake (east-half), Newfoundland; Geol. Surv. Can., Map 1196A.

ICE SEGREGATION AT DEPTH IN PERMAFROST

Project 680047

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The maximum depth to which segregated ice can occur in permafrost is unknown. The uncertainty is of practical concern, because the thaw of excess ice at depth might cause instability problems, as around a hot, producing oil well. The purpose of this note is to examine some of the evidence for ice segregation at depth.

The Evidence from Drilling

Many thousands of shotholes have been drilled in the Canadian Arctic, but most have been less than 50 m in depth. Nevertheless, massive segregated ice has been encountered in an appreciable fraction of those holes that have exceeded 40 m. Unfortunately, coring in deep exploratory wells is rarely carried out in the upper few hundred metres, so no data are available on the presence or absence of ice at such depths.

The permafrost literature of the U.S.S.R. carries numerous references to massive ice at depth. Some of the ice appears to be segregated ice. Baulin *et al.* (1973) discuss ice-containing horizons of great thickness at depths up to 200 m. Vtyurin (1973) reports that test borings have revealed the presence of segregated ice in unconsolidated deposits to a depth of 130 m or more. Romanovskii (1973) states that there are large amounts of injection and segregated ice to depths of 200 m in the northern part of western Siberia. Bulmasov (1963) mentions a frozen core of friable material, copiously impregnated with ice, extracted from a depth of 1100 m. Upon thawing, the core turned into a sludge. Thus the drilling evidence for segregated ice suggests a minimum depth of at least 200 m.

The Evidence from Pingos

Pingos are ice-cored hills that grow either from the freezing of free water at the bottom of the ice core or from the growth of segregated ice (Mackay, 1973). For example, if the ice core of Ibyuk Pingo near Tuktoyaktuk, N.W.T. is assumed to be of pure ice, then the bottom of the core is about 65 m below the top. Since Ibyuk Pingo is still growing, ice obviously can segregate at a minimum depth of 65 m under present western Arctic coastal conditions.

Pore Water Expulsion

The theoretical aspects for ice segregation have been discussed by Williams (1967, 1968) and others. Segregated ice tends to grow in a given soil type when the following equation is satisfied, viz:

$$\sigma_i - u \leqq \frac{2\sigma_{iw}}{r_c} \qquad (1)$$

where σ_i is the pressure of the ice; u the water pressure at the penetrating frost line; σ_{iw} the interfacial tension ice-water; and r_c the radius of the largest continuous pore openings or channels. Where an ice lens of large areal extent is growing, the pressure of the ice (σ_i) can be taken equal to the overburden pressure (σ). The right hand term of equation (1) is a constant for a given soil type. Williams (1967) suggests a range of 0 to 0.075 kg/cm^2 for sands; 0.075 to 2.0 kg/cm^2 for silts to silty clays; and upwards of 2.0 kg/cm^2 for clays. Equation (1) can be rewritten:

$$\sigma - u \leqq C \qquad (2)$$

where σ-u is the effective stress (McRoberts and Morgenstern, 1975), and C is a soil constant.

Numerous laboratory studies (e.g. McRoberts and Morgenstern, 1975; Takashi and Masuda, 1971; Takashi *et al.*, 1974) have shown that if the effective stress exceeds C, pore water tends to be expelled in advance of the freezing plane.

Artesian pressures from expelled pore water are commonly encountered in drillholes that penetrate aggrading permafrost along the western Arctic coast. Although other factors also enter, such as the freezing rate and stress history, the growth of segregated ice with heave, pore ice with heave, and pore ice with water expulsion appears to depend upon whether the effective stress is less than, equal to, or greater than the soil constant (Fig. 61.1).

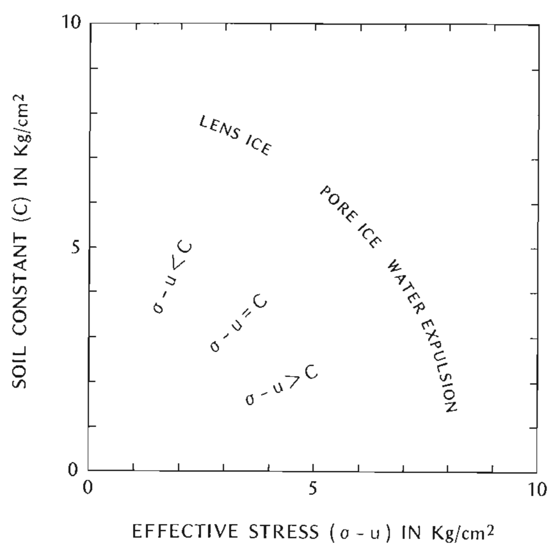

Figure 61.1

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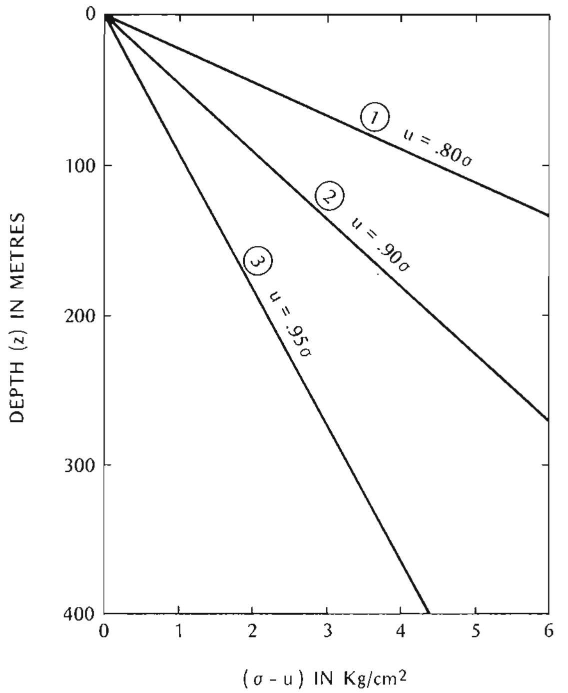

Figure 61.2

Laboratory and field studies show that pore water pressures, generated by water expulsion, may exceed 80 per cent of the overburden pressure. Figure 61.2 shows the effective stress (assuming a mean soil density of 2.2 g/cm^3) for the unfrozen soil beneath aggrading permafrost for different pore water pressures and depths. For example, the effective stress at a depth of 200 m could be less than the soil constant for some clays, if pore water pressures were 95 per cent of the overburden pressure. Under field conditions, where permafrost was aggrading on a regional scale, permafrost would vary in thickness and the soil constant C also would vary greatly, commonly within a few centimetres, in interstratified sands, silts, and clays. Consequently, if adjacent sites with thick and thin permafrost were hydraulically connected by permeable soils, water expulsion from beneath thick permafrost might reduce the effective stress beneath thin permafrost so that conditions favoured ice lensing in fine grained soils at the downward moving frost line, as now occurs in pingo growth.

Conclusion

If aggrading permafrost is considered in isolation with hydrostatic conditions prevailing, minimal ice lensing would be expected at depths of 50 m or more. However, when aggrading permafrost is considered on a regional scale with variations in permafrost thickness, pore water expulsion, and interstratified coarse and fine grained soils, then segregated ice probably could grow to depths of at least 200 m.

References

Baulin, V.V., Dubikov, G.I., and Uvarkin, Yu. T.
1973: The main features of the structure and development of the permafrost of the west Siberian plain (in Russian); Second Int. Conf. Permafrost, U.S.S.R. Acad. Sci., Yakutsk, v. 2, p. 10-17.

Bulmasov, A.P.
1963: Structures and gravity anomalies of cryogenic origin in the Baikal Region (in Russian); Geol. Geofiz. (Akad. Nauk SSSR, Sib. Otd.), v. 2, p. 75-85; trans. by E.R. Hope, Defence Res. Bd. Canada T482R, 1968, p. 1-12.

Mackay, J.R.
1973: The growth of pingos, western Arctic coast, Canada; Can. J. Earth Sci., v. 10, p. 979-1004.

McRoberts, E.C. and Morgenstern, N.R.
1975: Pore water expulsion during freezing; Can. Geotech. J., v. 12, p. 130-141.

Romanovskii, N.N.
1973: Effect of neotectonic movements on the formation of permafrost regions (in Russian); Second Int. Conf. Permafrost, U.S.S.R. Acad. Sci., Yakutsk, v. 3, p. 51-57.

Takashi, T. and Masuda, M.
1971: An experimental study of the effect of loads on frost heaving and soil moisture migration; J. Japanese Soc. Snow Ice, v. 33, p. 1-11.

Takashi, T., Masuda, M., and Yamamoto, H.
1974: Experimental study on the influence of freezing speed upon frost heave ratio of soil under constant effective stress; J. Japanese Soc. Snow Ice, v. 36, p. 1-20.

Vtyurin, B.I.
1973: Patterns of distribution and a quantitative estimate of the ground ice in the U.S.S.R. (in Russian); Second Int. Conf. Permafrost, U.S.S.R. Acad. Sci., Yakutsk, v. 3, p. 12-19.

Williams, P.J.
1967: Properties and behaviour of freezing soils; Norwegian Geotech. Inst., Oslo, Publ. no. 72, 119 p.

1968: Ice distribution in permafrost profiles; Can. J. Earth Sci., v. 5, p. 1381-1386.

GEOMORPHOLOGICAL PROCESS AND TERRAIN DISTURBANCE STUDIES, BANKS ISLAND, DISTRICT OF FRANKLIN

Project 640004

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Terrain Sciences Division

Various geomorphic and terrain disturbance studies were continued at Sachs Harbour and elsewhere. Additional support was provided by the ALUR Program of the Department of Indian and Northern Affairs.

Mass Wasting Processes, Sachs Harbour

Rates of subsurface movement are now available for the six-year period 1969 to 75 at six sites. Technical details concerning the measurements and the results obtained for the three-year period 1969 to 72 have already been presented (French, 1973, 1974a). Data obtained in 1975 relate to nonsorted stripes on low-angled slopes and are directly comparable with earlier results.

The absolute amount of movement recorded at various depths is presented in Table 62. 1. In general, these long-term measurements support earlier conclusions concerning the nature of solifluction movement. Rates of average subsurface movement, however, appear to be significantly less when measured over the six-year period. Average annual movement in the top 30 cm was 1. 05 cm/yr for 1969 to 75 as compared to 1. 56 cm/yr for 1969 to 72. This difference probably does not relate to temporal differences in the rate of operation of mass wasting processes before and after 1972 but rather is the effect of relatively rapid soil movement following the initial installation of the markers and the infilling of the soil pits in 1969.

Rough calculations indicate the volume of material moved by mass wasting processes on Banks Island. Given that the mean value for movement near the surface was 1. 2 cm/yr and assuming that movement in the upper 50 cm is, on average, half of the surface rate, this gives a volumetric downslope movement of approximately 30 $cm^3/cm/yr$. Alternatively, if the maximum thickness of the active layer is assumed to be 75 cm and if average movement in this zone is assumed to be one quarter of the surface rate, then the total volumetric downslope movement is approximately 22 $cm^3/cm/yr$. These values are of the same order of magnitude as those recorded in other permafrost areas (e. g. Young, 1974, Table II).

Gully Erosion and Permafrost Conditions, Sachs Harbour Townsite

During 1975 little additional gullying took place and those gullies which had developed in the spring of 1973 and 1974 (French 1975a) did not significantly increase in size. This was attributed to (a) the early and gradual spring thaw of 1975 which enabled runoff to proceed in a uniform and continual manner and (b) the installation of a number of culverts to channel water

Figure 62. 1.

Thermokarst terrain developed in silty sands and gravels at site of Deminex Orksut 1-44 drillsite, Big River area. Date of the original disturbance was August 10-20, 1973. Photo taken July 20, 1975.

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Table 62.1. Average subsurface movement on low-angled slopes, Sachs Harbour, Banks Island, 1969-1975*

Depth	Site 6	Site 7	Site 8	Site 9	Site 10	Site 11	Total average movement, 1969-1975
3	9.2	8.1	11.5	7.1	2.6	6.3	7.47
8	6.9	6.6	11.1	6.8	2.5	5.8	6.60
15	6.0	6.3	10.3	7.3	0.7	4.5	5.85
30	5.5	5.1	9.5	6.8	1.0	3.6	5.25
Total average movement, 1969-75	6.9	6.5	10.6	7.0	1.7	5.0	6.29
Annual average movement, 1969-75	1.15	1.08	1.77	1.16	0.28	0.84	1.05

*Data are given in cm

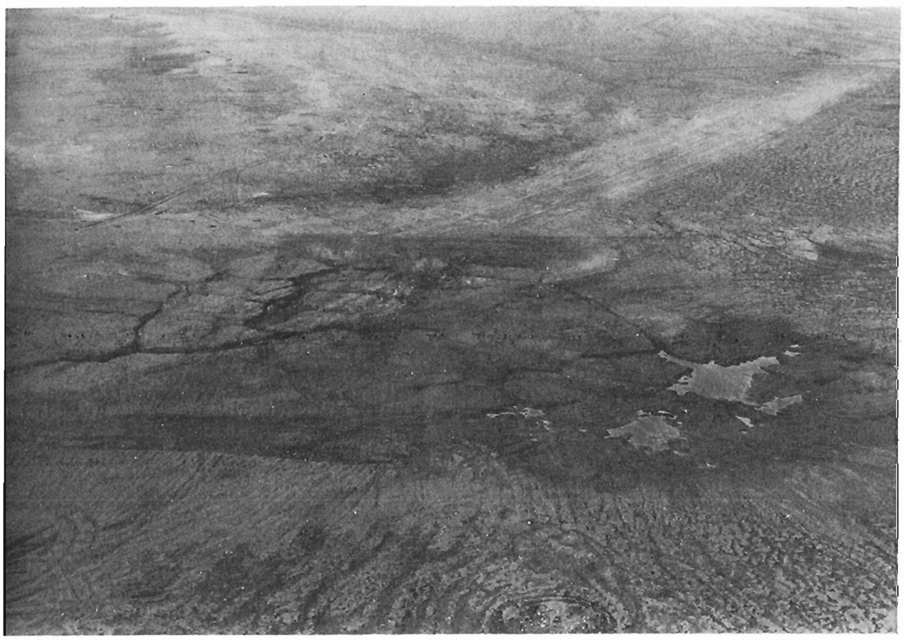

Figure 62.2.

Oblique air view from 200 m of disturbed terrain, Orksut 1-44, Big River area. Standing water bodies, hummocky terrain, and the beginnings of a polygonal system of gullies draining downslope towards left of photo can be seen. The landing strip is at top right. Photo taken July 20, 1975.

into the existing gullies. In addition, efforts have been made by the residents to minimize the amount of off-road vehicle movement and other terrain disturbance.

The permafrost and ground ice conditions of the sediments underlying the village have been examined in a number of small ice cellars dug by residents to serve as food preservers. Several ice wedges of moderate dimensions can be observed; typically they are 0.3 to 0.6 m in width near the surface and extend downwards to depths of 2.0 m or more. There is little or no expression of these ice wedges at the surface. The frozen material is composed of layers of silty sand, fine sand, and gravel. The silty sands and the gravels commonly contain excess ice of between 10 and 25 per cent which is visible as individual ice inclusions and ice coatings around pebbles (Vx/Vc ice; Pihlainen and Johnston, 1963). The fine sands are horizontally bedded and contain little visible or excess ice except where they are penetrated by pure ice lenses up to 10 cm thick or by thin icy layers (Vs ice) which give excess ice values of between 40 and 95 per cent.

The occurrence of these icy layers and ice lenses, the ice-rich nature of the silty sands and gravels, and the presence of the ice wedges suggest that the sediments underlying the townsite are highly susceptible to any surface disturbance. In addition, active layer depths and soil temperatures in a number of undisturbed and disturbed sites in the townsite indicate variations in

Figure 62. 3.

Oblique air view from 200 m of Hercules landing strip, Orksut 1-44, Big River area, showing frost fissure polygons and unvegetated nature of the gravel terrace. At upper right, the landing area extends over wet tundra where subsidence is occurring. Photo taken July 20, 1975.

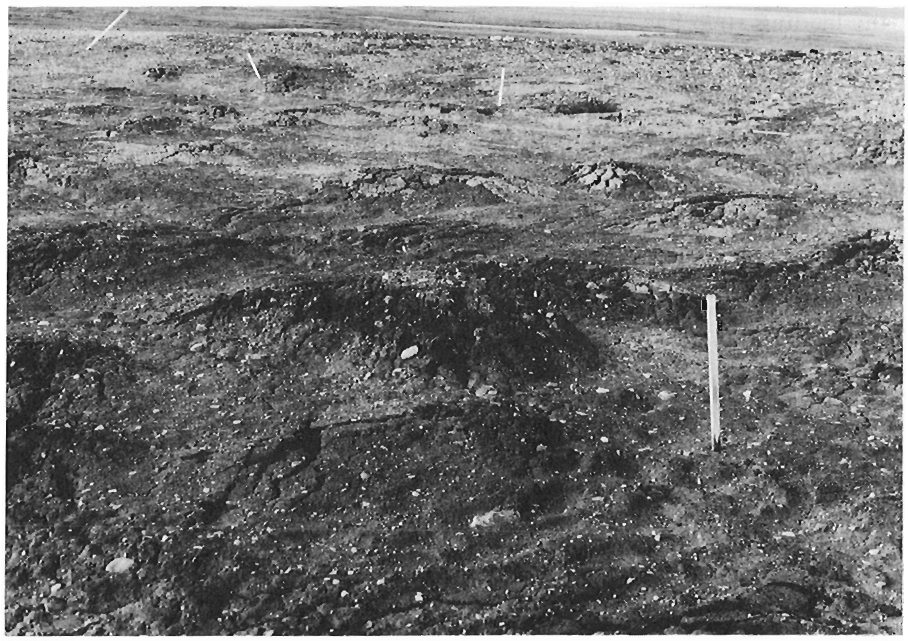

Figure 62. 4

Active subsidence at end of Hercules landing strip, Big River. Photo taken July 20, 1975.

thaw depths of at least 10 to 20 cm. This data emphasizes the protective thermal role of the tundra vegetation, which is present on the lower slopes around the townsite, and the necessity to minimize surface disturbance.

Terrain Disturbance, Big River (72°23'N, 122°42'W)

This location is the site of the Deminex exploratory borehole Orksut 1-44 which was drilled during the 1972-3 winter season. It was first examined in August, 1973 immediately prior to its abandonment (French, 1974b). It was revisited for one day in July 1975 during part of a terrain disturbance study supported by the ALUR Program.

Several important changes had occurred in the two-year period. The gently sloping area which had been bulldozed and disturbed to provide fill for the sumps had begun to subside. Micro relief in excess of 1.0 m in relative height had developed. It consisted of small mounds separated by linear depressions and standing water bodies developed in silty sands and gravels (Fig. 62. 1). The topography was remarkably similar to that which can be observed adjacent to the airstrip at Sachs Harbour (French, 1975b). In addition, a crude polygonal gully system had developed to drain the terrain (Fig. 62. 2), probably reflecting preferential subsidence along underlying ice wedges. The short time period over which these changes have occurred emphasizes the rapidity at which thermokarst processes commence operation. Future visits are planned to document the progress of thermokarst development.

The two sites used as landing strips also were revisited. A few minor changes were noticed on the strip adjacent to the campsite. Soil temperatures at 10 cm depth indicated the airstrip to be approximately 0.5°C warmer than the adjacent hummocky Dryas terrain and the active layer to be over 100 cm thick on the airstrip and only 84 cm thick on the undisturbed terrain. No obvious subsidence of gullying had occurred. The Hercules landing strip had been located approximately 3 km away on a low gravel terrace near Big River. This terrace possesses a well developed network of frost fissures and is virtually devoid of vegetation (Fig. 62.3). At one end, however, attempts had been made to expand the strip into a wet tundra depression. The vegetation had been disrupted, and by late summer 1973 there had been signs of subsidence. At that time markers were relevelled in along two lines across the landing strip in that vicinity. In 1975 the markers were relevelled and an average per annum subsidence of between 2 and 5 cm was found to have occurred (Fig. 62.4). By contrast, the remainder of the landing strip has shown little or no subsidence, even above the lines of the frost fissures. It would appear that the absence of vegetation on the terrace resulted in minimal changes to the thermal regime of the underlying sediments when the landing strip was prepared and in use.

References

French, H.M.
1973: Geomorphological processes and terrain sensitivity, Banks Island, District of Franklin; *in* Report of Activities, Part A, Geol. Surv. Can., Paper 73-1A, p. 220-223.

1974a: Mass wasting at Sachs Harbour, Banks Island, N.W.T., Canada; Arct. Alp. Res., v. 6, p. 71-78.

1974b: Geomorphological processes and terrain sensitivity, Banks Island, District of Franklin; *in* Report of Activities, Part A, Geol. Surv. Can., Paper 74-1A, p. 263-266.

1975a: Pingo investigations and terrain disturbance studies, Banks Island, District of Franklin; *in* Report of Activities, Part A, Geol. Surv. Can., Paper 75-1A, p. 459-464.

1975b: Man-induced thermokarst, Sachs Harbour airstrip, Banks Island, Northwest Territories; Can. J. Earth Sci., v. 12, p. 132-144.

Pihlainen, J.A. and Johnston, G.H.
1963: Guide to a field description of permafrost; Can. Nat. Res. Coun., Assoc. Comm. Soil Snow Mech., Tech. Mem. 79., NRC Publ. 7576.

Young, A.
1974: The rate of slope retreat; Inst. Br. Geogr., Spec. Publ. no. 7, p. 65-78.

ROLE OF LAHARIC BRECCIA IN GENESIS OF VOLCANOGENIC MASSIVE SULPHIDE DEPOSITS

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In outlining the spatially significant parameters of regional geology associated with volcanogenic massive sulphide deposits, Sangster (1972) noted that volcaniclastic breccia, commonly termed 'mill rock', is the dominant and recurrent rock type found in the stratigraphic unit which encloses the sulphide body. Although Sangster indicated that this rock is commonly stratigraphically above the ore, examinations of massive sulphide regions at Sturgeon Lake, Flin Flon, and Slave Province indicate that a somewhat similar facies occurs in the footwall sequence of many deposits. Detailed investigation of a footwall breccia unit near the Mattabi deposit, supplemented by preliminary examinations of the Flin Flon, Snow Lake, and Slave Province deposits, indicates that this rock has a spatial and probable genetic importance to volcanogenic deposits, and may be of some use in exploration for orebodies of this type. The observations and model established at Mattabi will be further tested as part of a continuing metallogenic study of the western Canadian Shield.

Volcaniclastic breccias include a wide range of volcanic-related, coarse clastic rocks. The term 'breccia' connotates size only, and Fisher (1961) includes only clasts of greater than 256 mm in this class. Fisher (1960, 1961) divides volcaniclastic rocks into three groups: (1) autoclastic, (2) pyroclastic (direct products of volcanic activity; includes various size classes such as autoclastic flow breccia and pyroclastic breccia, agglomerate, lapillistone and tuff), and (3) epiclastic rocks (include epiclastic volcanic breccia, volcanic conglomerate, sandstone, siltstone and claystone). These rocks are "derived by weathering and erosion of lithified or solidified volcanic rocks" (Fisher, 1961, p. 1413), and are clearly formed by weathering processes. As part of his epiclastic group, Fisher (1960) included laharic breccias. The latter are formed from "mudflow carrying, dispersing, and depositing coarse- and fine-grained particles and/or admixed non-volcanic material" (Fisher, 1960, p. 128).

Archean volcanic sequences include virtually all types of volcaniclastic rocks as outlined by Fisher, including significant quantities of laharic sediment.

Possible Significance of Footwall Rocks

The footwall rocks beneath volcanogenic massive sulphide deposits are important because in addition to containing the alteration zone which marks the 'hot-spring' path, they may also be the source of the metal. The source concept permits two possibilities: (a) that the metals are directly derived by fractionation, forming part of the vapour phase associated with the felsic volcanic fractional melt, or (b) that the metals are derived by leaching of volcanic materials below the deposit.

The first hypothesis would require large quantities of juvenile water in the melt, a fact which both experimental evidence and observation (Eggler, 1972; Burnham, 1967) seem to render unlikely. As Burnham (1967) pointed out, a melt must be saturated with water before a separate hydrous fluid phase will appear. Eggler (1972) showed that the Paracutin lavas had approximately 2.2 per cent water and were undersaturated. As examples of modern exhalitive systems, Craig (1966, 1969) has shown that the metal-rich water from the Red Sea and Salton Sea geothermal systems is largely of meteoric origin. Thus a meteoric rather than juvenile origin for the water appears most probable.

The first hypothesis also requires that metals are concentrated in the felsic fractionation product. If this

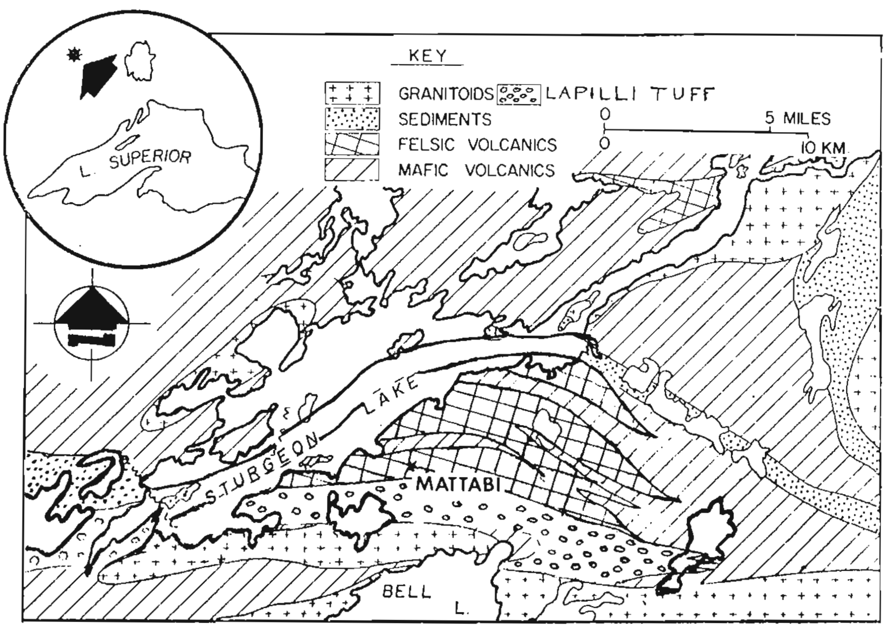

Figure 63.1.

General geology, Mattabi Area.

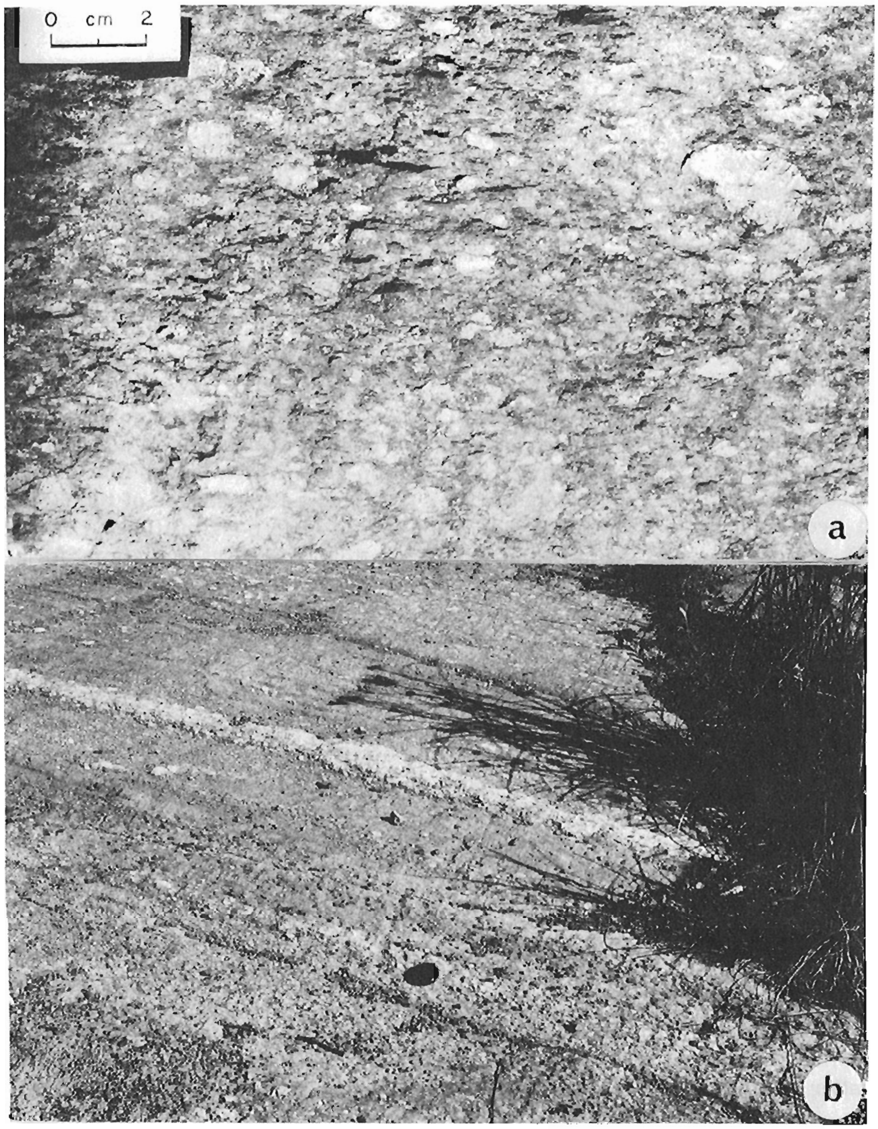

Figure 63. 2

(a) Footwall lapilli tuff; note felsic fragments (light) in basaltic matrix (dark).

(b) Footwall lapilli tuff, 3 miles west of Mattabi; note bedding of fragments.

were true, the felsic fraction (either extrusive or hypabyssal intrusive) should be similarly enriched in base metals, yet examination of trace element data (Goodwin, 1972 and Table 63. 1) indicates that metal is more abundant in the basaltic phase than the felsic phase. Furthermore, data on chemically similar island arc suites indicate that partial melting rather than fractional crystallization is the dominant process in producing the observed volcanic suites (Carmichael *et al.*, 1974).

The second hypothesis, leaching of metal from rocks underlying the deposit, requires — (a) an adequate 'plumbing system', including reasonable permeability and some driving force causing metal-bearing solutions to move up and out of the rock as a geothermal brine, and (b) an adequate metal source readily available to a leaching system.

In meeting the first criteria (a), only the physical nature of the source rock is important. Obviously, massive rocks have very low permeability, and transmission of fluids would be confined to fractures and interflow surfaces. An increase in porosity and permeability can be attained in vesicular rocks; however, vesicular or highly pumiceous units are less common than breccias within the footwall sequence of massive sulphide deposits. Thus volcaniclastic breccias are of major importance, as they have textural inhomogeneity and high permeability due to the poor packing and angularity of the fragments.

Evaluation of the second criteria (b), is best done through examination of some examples. A model can be developed on the basis of detailed work at the Mattabi deposit; this model is being further tested through examination of other deposits in Churchill, Slave, and Superior provinces.

Mattabi Footwall Rocks

Mattabi is in most respects a typical massive sulphide deposit, with two stacked massive sphalerite-pyrite orebodies conformably lying in a sequence of

Table 63.1

	Cu	Zn
Felsic volcanic (greater than 65% SiO_2)[1]	25	68
Mafic volcanic (less than 65% SiO_2)[1]	90	91
Footwall breccia ('lapilli tuff')[1]	36	104
Beidleman trondhjemitic sill[2]	9	35

Notes: [1]Samples taken away from obviously altered (Na-depleted) zone.

[2]Samples taken away from metal-enriched 'porphyry copper' zone.

[3]Analyst: J.M. Franklin.

[4]Method: Atomic absorption at Lakehead University Laboratory.

felsic pyroclastic rocks (Fig. 63.1). The chalcopyrite stringer zone, under the massive ore zones, cuts across footwall rhyolite breccia and extends down into a 'lapilli tuff' unit. A more complete description of the geology and alteration chemistry is available elsewhere (Franklin *et al.*, 1973, 1975).

The footwall rhyolite breccia in the Mattabi mine area is composed of densely-packed, angular to subangular pyroclastic breccia- and agglomerate-sized fragments set in a coarse, tuff-sized matrix which comprises no more than 10 per cent of the rock. Away from the deposit, the fragments are agglomerate-sized and more rounded, with up to 75 per cent matrix. The unit is 250 feet thick.

Below the footwall rhyolite is a 5000-foot thick, laterally extensive unit of 'lapilli tuff' (Franklin, 1975). This unit is composed of felsic and mafic lapilli set in a dark green, mafic matrix (Fig. 63.2a). The lapilli locally form up to 30 per cent of the rock, but on the average form only about 6 per cent. Near the stratigraphic top of the unit, they form lenticular beds (Fig. 63.2b) but below, they are randomly distributed. The fragments are dominantly felsic below and west of Mattabi, but are of mixed lithology to the east. The fragments are subangular to rounded; near the top of the unit some are pumiceous. The fragments are equant, and some appear broken. The matrix is dark green, fine grained, massive, and composed of chlorite (20 to 60 per cent), carbonate (up to 30 per cent), biotite (up to 20 per cent), sericite (up to 20 per cent), chloritoid (up to 20 per cent), and quartz and feldspar. Quartz

Table 63.2

Selected Analyses of Footwall 'Lapilli tuff'

	SiO_2	Al_2O_3	TiO_2	Fe_2O_3	MgO	MnO	CaO	K_2O	Na_2O	P_2O_5	CO_2	H_2O
SL 21 B	52.80	14.40	1.34	16.00	4.69	0.52	0.34	1.48	0.22	0.26	5.87	3.16
FR SL 39	55.22	16.42	1.05	9.95	6.23	0.12	8.29	0.01	3.51	0.21	0.40	0.90
FR SL 14	56.16	16.94	0.80	6.94	3.22	0.17	4.96	1.80	1.14	0.20	1.32	1.52
C 43	86.64	7.62	0.17	1.70	1.42	0.02	0.04	1.78	0.12	0.02	0.00	1.17
9 W82 N	72.95	10.56	0.47	2.95	2.33	0.09	2.75	1.17	3.67	0.14	4.00	0.81

SL 21 B — 350 feet south of Mattabi, in alteration pipe.
FR SL 39 — 2 miles west of Mattabi.
FR SL 14 — 1½ miles east of Mattabi.
C 43 — Felsic fragment, 300 feet south of Mattabi, in alteration pipe.
9 W82 N — Felsic fragment, 1 mile west of Mattabi.
Analyst: K.H. Poulsen, Lakehead University.
All elements XRF except MnO (A.A.), Na_2O (flame), P_2O_5 (colorimetric), CO_2, H_2O (gas detector)

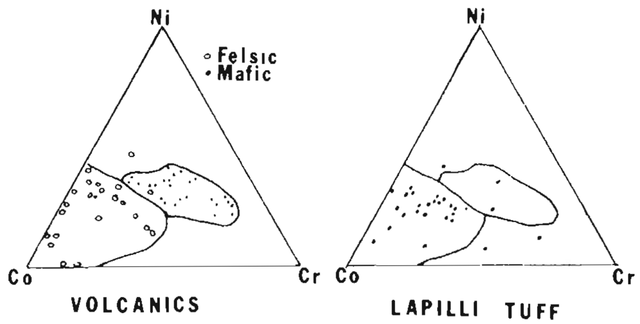

Figure 63. 3.

(a) Ni-Co-Cr in volcanic rocks, South Sturgeion Lake volcanic pile; open circles = felsic rocks (greater than 65%SiO_2), closed circles = mafic rocks (less than 65% SiO_2).

(b) Ni-Co-Cr in footwall lapilli tuff; fields from Figure 63. 3a.

grains are angular and may be polycrystalline. Feldspar occurs in lithic clasts and as broken chips. The matrix is a rather homogeneous 'paste', resembling chloritic greywacke.

Compositionally (Table 63. 2), most of the rocks are clearly mafic in bulk composition. The felsic lapilli content causes an increase in silica to over 60 per cent in a few samples, but generally the rocks are compositionally close to Goodwin's (1972) average andesite. The fragments are clearly more rhyolitic (Table 63. 2). The mafic matrix in this rock is either related to a mafic volcanic parent, or formed as a result of metasomatic alteration of a more felsic ash. Although Mg metasomatism from sea water has been documented (Bischoff and Dickson, 1975), an examination of the TiO_2 content of the rock precludes this origin. The average TiO_2 content of 40 samples of footwall lapilli tuff is 0. 85 per cent, similar to the average TiO_2 content (0. 84 per cent) of 25 samples of mafic rock (less than 65 per cent SiO_2) taken from the upper volcanic cycles, and to Goodwin's (1972) average andesite (0. 89 per cent). In that titanium is an immobile element and therefore is useful as a primary petrogenetic indicator in altered rocks (Cann, 1970), it must be accepted that the matrix of the footwall lapilli tuff is derived from a basaltic parent. Thus, in terms of major element chemistry, the fragments are generally rhyolitic in composition, but the matrix is basaltic.

Volcanic products of two greatly different compositions (rhyolite and basalt) are produced simultaneously from the same vent only under very limited, unusual conditions, such as liquid immiscibility to produce a variolitic lava (Gélinas, 1975) or by removal of one rock type from the sides of the vent and incorporation into a later flow. In either case, distinctive textures are produced, and the latter case is confined to a localized area. The footwall 'lapilli tuff' at Mattabi extends for over 14 km and is over 1500 m thick. No variolitic texture is evident, and with such a large volume and lack of igneous texture in the matrix, some process other than a purely volcanic phenomenon must have operated to produce the rock.

A further interesting comparison may be made by examining the relations of chrome, nickel, and cobalt. Nickel and chrome are usually strongly fractionated between felsic and mafic phases; cobalt is less well fractionated. A ternary plot (Fig. 63. 3a) of these elements in felsic (greater than 65 per cent SiO_2) and mafic (less than 65 per cent SiO_2) volcanics from throughout the pile shows that the two fields can be easily distinguished, with mafic rocks clearly enriched in Ni and Cr relative to Co. A similar plot for the footwall lapilli tuff (Fig. 63. 3b) illustrates that these essentially mafic rocks are clearly anomalous in their Ni-Co-Cr ratios, with a distribution pattern essentially similar to felsic rocks. In that their titania is immobile (Cann, 1970) and its content indicates a mafic parent, some leaching process must have affected these rocks to remove Ni and Cr. Although the exact nature of this process is unclear, comparison of these rocks with limited data on chloritic sediments associated with iron formation at the top of the South Sturgeon Lake volcanic pile indicates that these sediments have a similarly depleted Ni and Cr content. Thus these elements may have been removed during either the pre-transportation weathering, or the transportation of the mafic detritus.

The zinc and copper contents of the footwall lapilli tuff (Table 63. 1) indicate a very high zinc content, but a copper content considerably lower than a typical mafic volcanic. Thus the base metal content of the rock has been modified, again possibly immediately before or during a transportation event.

In summary, the texture, distribution, bimodal composition, and minor element composition of this unit all indicate a sedimentary origin, yet the fragments are clearly derived from a volcanic source. The unit most closely resembles a lahar, as described by Fisher (1960), and is probably a submarine lahar. A lahar such as this is important to massive sulphide genesis, because (a) it was a water-charged sediment, and therefore contained abundant medium to carry the metal and eventually form the exhalative fluid, (b) its brecciated, poorly sorted, and coarse clastic texture provides a high permeability relative to most volcanic rocks, and (c) its mafic component has a high metal content.

Model

Epiclastic laharic breccias with a significant mafic component are spatially and possibly genetically related to massive sulphide deposits and the fact that rhyolite flows and pyroclastic (hydro- and steam explosion) breccias are also common rocks in the immediate footwall must be considered in any genetic model.

Within Archean volcanic sequences, rhyolitic volcanism is volumetrically small. For example, in the Abitibi belt, Goodwin and Ridler (1970) indicate that only 7. 3 per cent of the volcanic rocks are felsic. In most areas, felsic volcanism is confined to a few small centres; felsic ash may be transported from these centres and deposited over a wide area, but the centres of flow activity and coarse breccia formation are generally less than one kilometre in diameter.

The spatial relation of felsic volcanic centres to massive sulphide deposits indicates that felsic volcanism has acted in some way to concentrate the exhalative activity which produced the deposit. A process which includes both localized felsic volcanism and laharic breccia in generating massive sulphide deposits may be envisaged as follows:

(a) emplacement of a seawater-charged laharic breccia, by rapid slumping off of adjacent volcanic pile, mixing of lithotypes, and possible introduction of minor pyroclastic component.
(b) initiation of local felsic volcanism, causing localized heating of interstitial water. This action would: 1) alter the chemistry of the water, including a drop in pH, thus enabling the water to leach metals from the mafic portion of the breccia, 2) provide a local point of anomalous heat flow, thereby causing convective flow of the interpore water and enhancing the leaching activity, and 3) provide a local exhalative centre through which the heated, metal-rich waters could escape.

The metal-leaching ability of heated (200°C), pressurized (500 bars) seawater on basalt (BCR1) has been experimentally demonstrated by Bischoff and Dickson (1975). They show that heavy metals are dissolved during the seawater-basalt interaction over a geologically very short time (198 days), in amounts adequate to easily provide the metal for a massive sulphide deposit Recalculation of their data shows that approximately 9 per cent of the copper was removed from the sample, principally from the glass phase of the basalt. Removal of a similar amount of metal from the Na-depletion zone under Mattabi would provide the amount of copper contained in the Mattabi deposit and its alteration zone.

Examples and Application

(a) Mattabi: The 'footwall lapilli tuff' is probably a laharic breccia and provides adequate metal source, in that it had favourable permeability, a suitably large leachable surface area of fragments, and high metal content of the matrix. Localized felsic volcanism provided the 'engine' to cause both the leaching and expulsion of metal-rich fluid. The large alteration zone outlined by Na-depletion, forming a hemisphere under the deposit, probably represents the metal source volume.

(b) Flin Flon: Andesite breccia forms the dominant footwall rock (Koo and Mossman, 1975), and is composed of angular to rounded blocks of amygdaloidal to massive andesite, set in a fine grained mafic matrix (Fig. 63.4a). The unit is cut by the 'quartz porphyry' (Stockwell, 1960), a series of hypabyssal sills and dykes which possibly represent feeders to the quartz-porphyry rhyolite which forms stratiform units above the deposit. The mixture of textural types within the breccia indicates that it has been transported, probably in a fluid medium. The angularity of the fragments would enhance the permeability and, furthermore, the intrusive quartz-porphyry could act as a heat engine.

Elsewhere in the Flin Flon area, similar laharic breccias are present near the West Arm deposit (on islands in lake to east of mine) and near the Schist Lake and Mandy deposits. Detailed stratigraphic studies have not yet been undertaken, so that the role and position of these breccias relative to the deposits is not established.

(c) Snow Lake: Although detailed stratigraphic relations around the Snow Lake deposits have not yet been determined, bimodal (mafic matrix-felsic fragment) epiclastic breccias are present at Chisel Lake (near powderhouse) and Stall Lake (footwall of deposit along railroad tracks). Anderson Lake deposit is underlain by mafic pyroclastic rocks, including some mafic breccias. In all cases, the breccias appear localized in the region of the deposits. Also, in all cases, felsic volcanism is evident in the immediate region.

(d) Slave Province: Bimodally composed epiclastic breccias are present in the stratigraphic footwall of the High Lake (Kennco) (Fig. 63.4b) and Hackett River Bathurst Norsemines) deposits (Fig. 63.4c). The rocks of the latter area are highly metamorphosed; consequently, the fragmental nature is only easily discerned where emphasized by the bimodal composition. At the Agricola Lake (Yava Syndicate) occurrence, a thin unit of bimodal epiclastic rock is present in the footwall (west) side of the deposit.

The 10-zone deposit of Texasgulf Ltd. near Takijuk Lake does contain minor epiclastic breccia, but also has as its footwall (?) rock a unit of very vesicular andesite (Fig. 63.4d). This rock could provide the important characteristics of permeability (due to high vesicularity) and metal-rich source rock (mafic composition) and thus act in a similar manner to an epiclastic breccia.

Summary

Epiclastic breccias with significant mafic component, either as matrix or fragment, are genetically important to massive sulphide deposits as source rocks. Felsic volcanism appears equally important in many areas, as the driving mechanism for exhalative activity. It is important to note that epiclastic breccias and felsic volcanism are absent in the areas of many small deposits (e.g. – Coronation, Birch Lake, Flexar copper deposits), and some deposits occur within sedimentary (?) sequences, with no immediately evident source (e.g. – Osborne Lake, Sherridon). It also is noteworthy that any metal-rich and water-charged source rock, such as shale, could, provided some mechanism of release is provided, act in a similar manner to that suggested for epiclastic breccia.

The pertinent factors which may be useful to exploration are:

1. Epiclastic, water-emplaced breccias, with significant mafic component (bimodal composition), and high metal content, are possibly important source rocks to massive sulphides. A high TiO_2 content of the matrix indicating true mafic parentage, together with

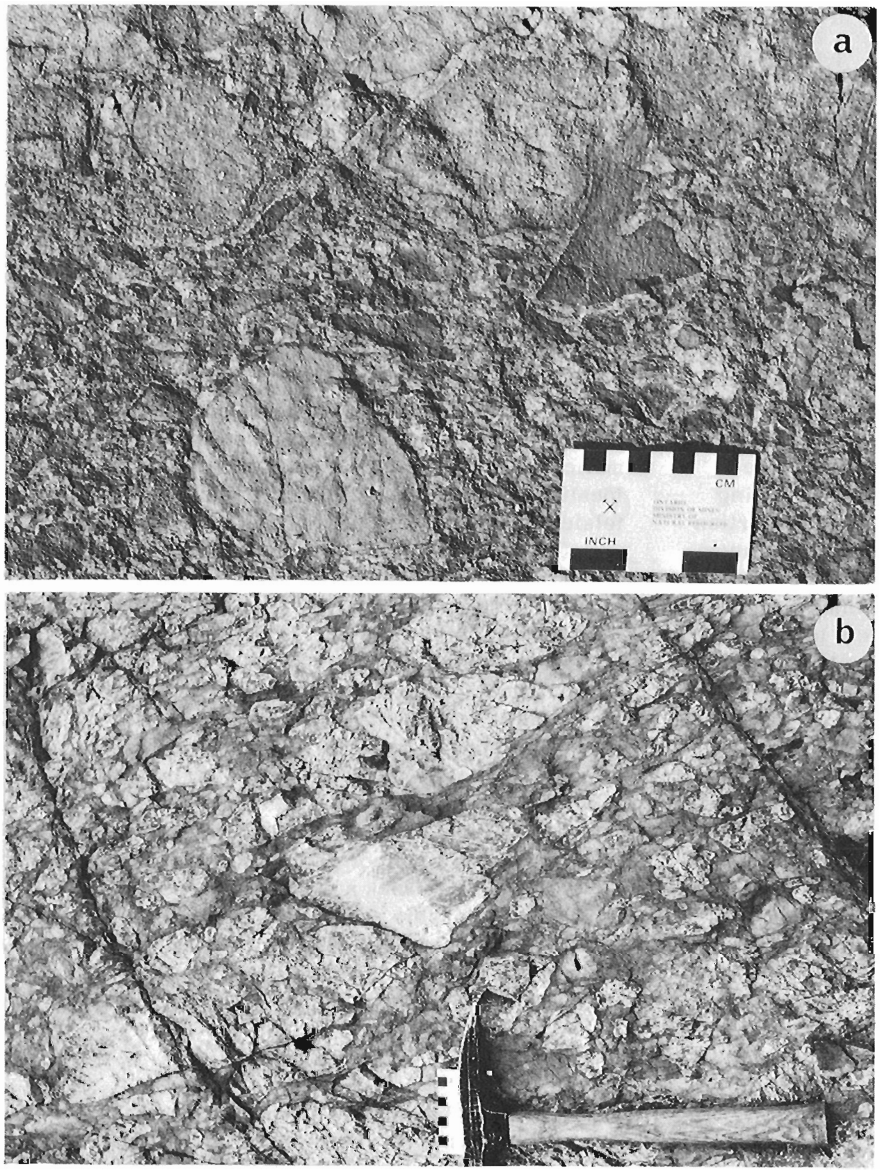

Figure 63.4.

(a) Andesite breccia, footwall of Main Mine, Hudson Bay Mining and Smelting Co., Limited, Flin Flon, Manitoba.

(b) Laharic breccia, High Lake deposit, N.W.T.

appropriate texture and minor element composition, all are useful aids in recognizing the rock as being a transported, volcanic derived sediment.

2. A mobilizing force such as localized felsic volcanism should be present. Identification of felsic volcanic centres, characterized by massive flows and angular pyroclastic breccia, should aid in narrowing the area of potential exhalative activity.

Acknowledgments

Dr. H.K. Sakrison, of New Insco Mines Ltd., has independently developed a similar model based on years of work in the Noranda area. Stimulating discussions with him have aided considerably in formulating the ideas presented here.

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References

Bischoff, J.L. and Dickson, F.W.
1975: Seawater-basalt interaction at 200°C and 500 bars: implication for origin of seafloor heavy metal deposits and regulation of seawater chemistry; Earth. Plan. Sci. Lett., v. 25, no. 3, p. 385-397.

Burnham, C. Wayne
1967: Hydrothermal fluids at the magmatic stage; *in* Chapt. 2 of Geochemistry of hydrothermal ore deposits, ed. H.L. Barnes; Holt, Rinehart, and Winston Ltd., p. 34-76.

Carmichael, I.S., Turner, F.J., and Verhoogen, J.
1974: Igneous petrology; McGraw-Hill, Toronto, 739 p.

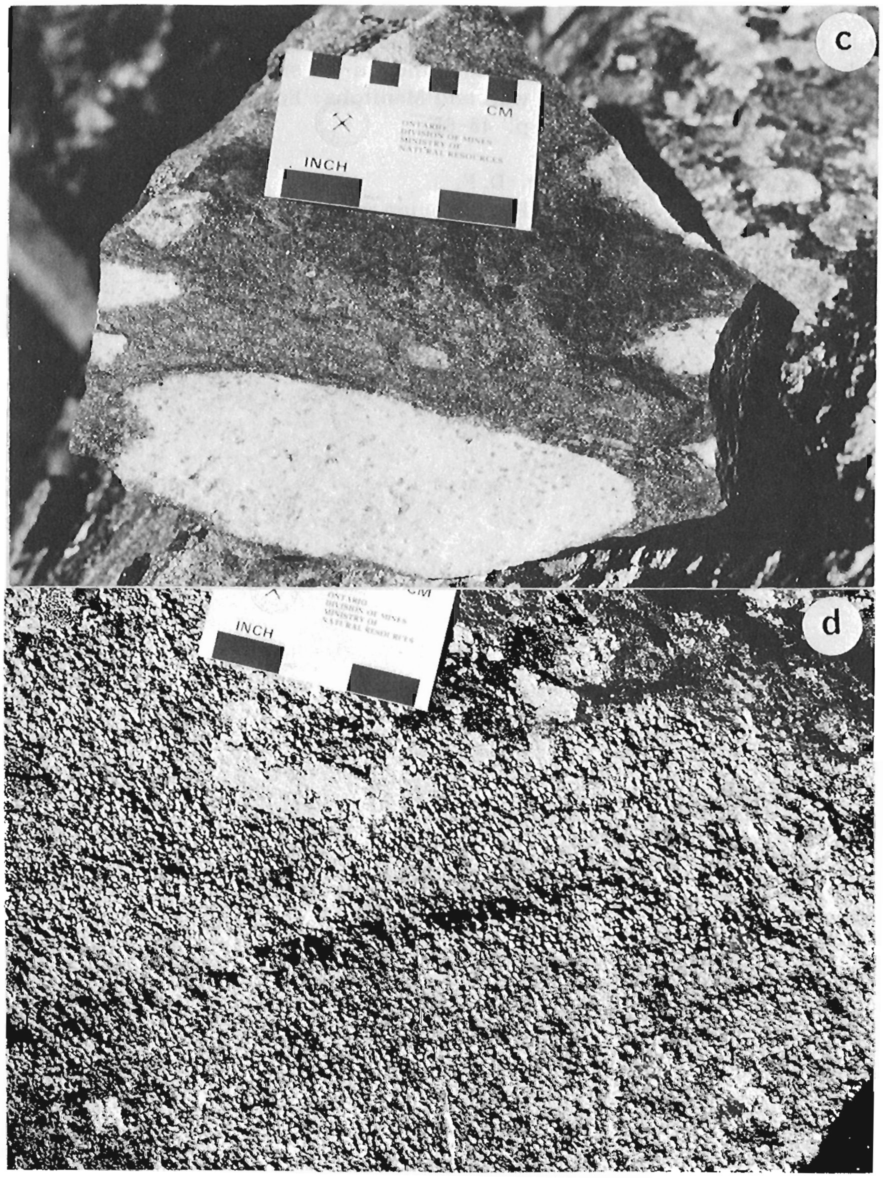

Figure 63. 4.

(c) Bimodal breccia, Bathurst Norsemines Ltd., Hackett River, N. W. T.

(d) Highly amygdular mafic lava, Texasgulf Ltd., no. 10 zone, Takijuk Lake, N. W. T.

Craig, H.
1966: Isotopic composition and origin of the Red Sea and Salton Sea brines; Science, v. 154, p. 1444.

1969: Geochemistry and origin of the red-sea brines; *in* Hot brines and recent heavy metal deposits in the Red Sea, eds. Degens & Ross; Spring Verlag, New York, p. 225-242.

Eggler, D. H.
1972: Water saturated and undersaturated melting relations in a Paricutin andesite and an estimate of water content in the natural magma; Contrib. Min. Petrol., v. 34, no. 4, p. 261-271.

Fisher, R. V.
1960a: Criteria for recognition of laharic breccias, southern Cascade Mtns., Washington; Geol. Soc. Am. Bull., v. 71, p. 127-132.

Fisher, R. V. (cont'd.)
1960b: Classification of volcanic breccias; Geol. Soc. Am. Bull., v. 66, p. 973-982.

1961: Proposed classification of volcaniclastic sediments and rocks; Geol. Soc. Am. Bull., v. 72, p. 1409-1414.

Franklin, J. M., Gibb, W., Tamman, A., and Kerr, B.
1973: The Mattabi deposit; Can. Inst. Min. Met., Geol. Div. Ann. Field trip, Guidebook, 19 p.

Franklin, J. M., Kasarda, J., and Poulsen, K. H.
1975: Petrology and chemistry of the alteration zone of the Mattabi massive sulphide deposit; Econ. Geol., v. 70, no. 1, p. 63-79.

Gélinas, L.
1975: Evidence for liquid immiscibility in basalts; *in* Guidebook, Field trip, volcanology and sedimentology, Rouyn-Noranda area; Que. Dept. Nat. Resour., p. 16-19.

Goodwin, A.M.
1972: The Superior Province; *in* Variations in tectonic styles in Canada, eds. R. Price and R. Douglas; Geol. Assoc. Can., Spec. Paper 11, p. 527-624.

Goodwin, A.M. and Ridler, R.H.
1970: The Abitibi Orogenic belt; *in* Symposium on basins and geosynclines of the Canadian Shield; Geol. Surv. Can., Paper 70-40, p. 1-24.

Koo, J. and Mossman, D.J.
1975: Origin and metamorphism of the Flin Flon stratabound Cu-Zn sulfide deposit, Saskatchewan and Manitoba; Econ. Geol., v. 70, no. 1, p. 48-62.

Sangster, D.F.
1972: Precambrian volcanogenic massive sulphide deposits in Canada: A review; Geol. Surv. Can., Paper 72-22.

Project 650056

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1. Carbonate- versus volcanic-hosted lead-zinc deposits: A review of the period 1970-1975

The discovery of the Robb Lake deposit in the fall of 1969 reactivated interest in carbonate-hosted lead-zinc deposits, following the initial flurry of activity in the Pine Point area from 1964 to 1967. In particular, the Yukon and Northwest Territories, because of several factors (Sangster, 1975; Brock, 1975), received a great deal of attention in addition to the Arctic Islands and the Appalachian Region. Several spectacular discoveries in carbonate rocks during this period prompted speculation that deposits of this type will be the major sources of lead and zinc of the near future, and that this could result in a major shift in the zinc-lead producing centres of Canada.

In 1973, 77 per cent of Canada's zinc, and 60.5 per cent of her lead came from volcanic-hosted deposits. This compared with 15.5 per cent zinc and 20.5 per cent lead from carbonate-hosted deposits and illustrates the traditional dominance of the volcanic source for both zinc and lead in Canada.

Table 64.2 lists only those discoveries (of both deposit-types) which were made during the sample period in question (1970-1975) and their tonnages and grades as reported to date. In addition, of course, many other significant, but as yet untested, discoveries of both types have also been recorded but, for comparison purposes, only those on which sufficient work has been completed to allow tonnage grade estimates, are recorded in Table 64.2. From this, it can be seen that the rate of discovery (as measured in terms of number of new deposits) of the volcanic type has far exceeded that of the carbonate type. One of many reasons for this, of course, is that the volcanic type normally has a mineralogical

TABLE 64.1

Carbonate- and volcanic-hosted lead-zinc deposits beginning production, 1970-1975

Carbonate-hosted

Name	Tonnage (million tons)	Grade, % Cu	Pb	Zn	Reference
Nfld. Zinc, Nfld.	4.5			8.8	Cndn. Mines Hnbk., 1974-75

Volcanic-hosted

Name	Tonnage (million tons)	Grade, % Cu	Pb	Zn	References
Centennial, Man.	1.4	2.06		2.6	Metal Sourcebook, 6/5/74
White Lake, Man.	.33	2.6		5.4	CIM[1] Bull. - July/1972
Millenbach, Que.	2.075	3.5		3.6	CIM Bull. - Nov./1973
Sturgeon Lake, Ont.	2.11	2.98	1.47	10.64	N.M. 31/10/74
Mattabi, Ont.	12.87	.91	.84	7.6	N.M. 22/3/73
Ruttan, Man.	51	1.47		1.61	N.M. 22/3/73
Ming, Nfld.	1.426	2.57		.56	Cndn. Mines Hnbk., 1974-75

[1] Canadian Institute of Mining and Metallurgy.

TABLE 64.2

Carbonate- and volcanic-hosted lead-zinc deposits discovered in Canada, 1970-1975

Carbonate-hosted

Name	Tonnage (million tons)	Grade, % Cu	Pb	Zn	References
Robb Lake, B.C.	6.1		7.3	*	N.M.[1] 30/1/75
Goz Creek, Y.T.	1.5		10	*	N.M. 30/1/75
Bear-Twit, N.W.T.	6 to 8		6 to 8	*	Brock, 1975
Arvik, N.W.T.	25		4.3	14.1	N.M. March/1975
Gays River, N.S.	12		5.9 to 7	*	N.M. May/1975

*denotes combined Pb + Zn

Volcanic-hosted

Name	Tonnage (million tons)	Grade, % Cu	Pb	Zn	References
Sturgeon Lake, Ont.	2.11	2.98	1.47	10.64	N.M. 31/10/74
Lyon Lake, Ont.	3.096	1.15	.6	6.2	N.M. 31/10/74
New Creek Zone, Ont.	.809	1.66	.76	8.84	N.M. 22/3/75
AEX (Grum), Y.T.	30	-------not reported---			N.M. 1/5/75
Reed Lake, Man.	1	2		4	N.M. 6/12/73
Frotet Lake, Que.	1.46	1.73		2.96	N.M. 13/2/75
New Bay Pond, Nfld.	20(approx.)	.5		2	
Patino, Que.	.625	4.5		10.8	N.M. 29/5/75
Iso (Magusi), Que.	4.11	1.2		3.55	N.M. 24/10/74
Westarm, Man.	.71	4.6		.6	H.B.M. & S.[2] Inf. to stockholders, 1973
Ming, Nfld.	1.426	2.57		.56	Cndn. Mines Hnbk., 1974-75
Ming Extension, Nfld.	.713	2.92		.65	
Detour River, Que.	35.4	.39		2.3	N.M. June/1975
East Cleaver, N.W.T.	8	.51	1.01	4.04	N.M. 28/3/74
Izok Lake, N.W.T.	.4	4		4.8	TGS[3] 1974 Ann. Rept.

[1]Northern Miner.

[2]Hudson Bay Mining and Smelting Co. Ltd.

[3]Texas Gulf Sulphur Co. Ltd.

composition such that it is readily detected by airborne geophysical methods, whereas the carbonate type does not, and requires much slower exploration methods. Offsetting this, however, is the much greater surface area of carbonate rock exposed by the rugged Cordilleran terrane, contrasted with the more subdued Precambrian topography where most of the volcanic discoveries were made.

Measured in terms of new producers during the same period, Table 64.1 illustrates that more volcanic lead-zinc is being placed on the market than carbonate lead-zinc.

The plethora of small zinc-lead occurrences in carbonate rocks of all ages in the Mackenzie Mountains, in the Carboniferous of Nova Scotia, and in the Ordovician of the Arctic Islands, for example, must be viewed as demonstrating that small sphalerite occurrences can be regarded as a normal feature of carbonates, just as chalcopyrite is of volcanics. Heyl (1968), for example, has pointed out this same characteristic for the classical Mississippi Valley lead-zinc district of the United States. When this is taken into consideration, the relatively few newly-proven significant deposits (Tables 64.2 and 64.1), relative to the literally thousands of new occurrences, must be regarded as a normal consequence of intense exploration in carbonate rocks, and may not necessarily be proportional to the potential of those rocks to contain exploitable deposits.

The data presented in Tables 64.2 and 64.1, however, do point out that zinc in both deposit-types is being discovered at a much greater rate than is lead. This is, in part, because most of the 1970-1975 new volcanic discoveries are of the typically Pb-poor Archean type and the carbonate-hosted lead-zinc deposits which were discovered have characteristically high Zn:Pb ratios. Canada's major lead producers, other than Pine Point, are volcanogenic deposits of Paleozoic age (e.g. - Brunswick No. 12, Faro) or shale-hosted deposits (e.g. - Sullivan).

In summary, in spite of several discoveries of the carbonate type in the period from 1970 to 1975, "volcanic zinc" was being discovered at a greater rate. The reverse, however, was the case for lead.

References

Brock, J.S.
1975: Mining: Yukon's first industry; Western Miner, Feb., p. 63-66.

Heyl, A.V.
1968: Minor epigenetic, diagenetic, and syngenetic sulfide, fluorite, and barite occurrences in the Central United States; Econ. Geol., v. 63, p. 585-594.

Sangster, D.F.
1975: The significance of smithsonite in Northern Cordillera; *in* Report of Activities, Part A, Geol. Surv. Can., Paper 75-1A, p. 235-236.

2. The Mackenzie Valley lead-zinc district, Canada

Until relatively recently, most Canadian carbonate-hosted lead zinc deposits, with the exception of the Kootenay Arc in British Columbia (Fyles, 1970) and Pine Point, N.W.T. (Skall, 1975) districts, occurred scattered singly or in very small groups across the country, and lacked demonstrable regional, stratigraphic continuity. Thus, for example, for many years, the Monarch-Kicking Horse (B.C.), Newfoundland Zinc (Newfoundland), Nastapoka Islands (N.W.T.), Bruce Peninsula (Ont.), Gays River (N.S.), and Strathcona Sound (Nanisivik) (N.W.T.) deposits (other than the exceptions noted above) were the major carbonate-hosted lead-zinc deposits in Canada. These were so scattered, in both time and space, that no cohesive, identifiable districts, comparable to those of the United States Mississippi Valley, Irish Central Plain, Polish Silesia, English Pennines, or Austrian Alps emerged.

Recently, however, literally hundreds of new discoveries, many of them significant in the economic sense, have been made in the northern Canadian Cordillera. Furthermore, these deposits are capable of being grouped, both geographically and stratigraphically (Taylor *et al.*, 1975; Brock, 1975; Smith, 1974; Dawson, 1975).

Accepting the distinction made by Heyl *et al.* (1974) that many of the major United States carbonate-hosted lead-zinc districts lie within the drainage of the Mississippi River (i.e. - within the Mississippi Valley), it is the purpose of this note to demonstrate that several Canadian carbonate-hosted lead-zinc sub-districts and deposits lie within a comparable district, herein referred to as the Mackenzie Valley district.

The Mackenzie Valley, the second largest watershed in North America (12th in the world), occupies an area of 1 841 000 km^2 (711 000 sq. mi.) and touches on three provinces and two territories (Fig. 64.1). It contains two of the largest lakes (Slave and Great Bear) in Canada, outside of the Great Lakes, and such major tributaries as the Peel, Liard, Peace, and Athabasca rivers. Geologically, portions of the Cordilleran Orogen, Interior Platform, Arctic Coastal Plain, and Canadian Shield regions are contained within it.

Although it is the purpose of this report to draw attention to the carbonate-hosted lead-zinc deposits of the Cordilleran Orogen and Interior Platform regions (Fig. 64.1), geologists familiar with the district will recognize that the Canadian Shield portion of the Mackenzie Valley district also contains the important uranium deposits of Rabbit Lake (Sask.), Cluff Lake (Sask.), Uranium City (Sask.), and Port Radium (N.W.T.), the gold deposits of Yellowknife, and the silver deposits of Echo Bay. Furthermore, the Cordilleran and Interior Platform regions encompass significant deposits of tungsten (Cantung, Mactung), copper

Figure 64. 1. Proposed Mackenzie Valley lead-zinc district, Canada, and tentative sub-districts. Examples of carbonate-hosted lead-zinc deposits as follows:

1. Flunk
2. Cab
3. Ab-Dab
4. Ping
5. Goz Creek (Barrier Reef)
6. Tara
7. Tom
8. Gayna
9. Tic
10. Rio
11. Bear-Twit
12. Art-Ekwi
13. Arn-Tee
14. Ice-Emily
15. Snobird
16. McBean
17. Cadillac
18. Matt Berry
19. Hundere
20. Quartz Lake
21. Silvertip
22. McDame
23. Redfern Lake
24. Mt. Helen
25. Robb Lake
26. Perkins
27. Pine Point
28. Wrigley
29. Og
30. Mel

TABLE 64.3

Grade-tonnage data of selected carbonate-hosted lead-zinc deposits or sub-districts, Mackenzie Valley district

Sub-district	Deposit	Tonnage (million tons)	Grade, % Pb	Grade, % Zn	References
** Pine Point		65	3	7	
Robb Lake		6.1	7.3*		N.M.[1] 30/1/75
Mackenzie Mtns.	Goz Creek	1.5	10 *		N.M. 30/1/75
Mackenzie Mtns.	Bear-Twit	7(approx.)	7 *		Brock, 1975
South Nahanni	Prairie Creek	2.1	12.5	15.5	N.M. 15/5/75

* denotes combined Pb + Zn

** production plus published reserves; data include 1.4 million tons of direct shipping ore at 19.3% Pb and 26.7% Zn

[1]Northern Miner.

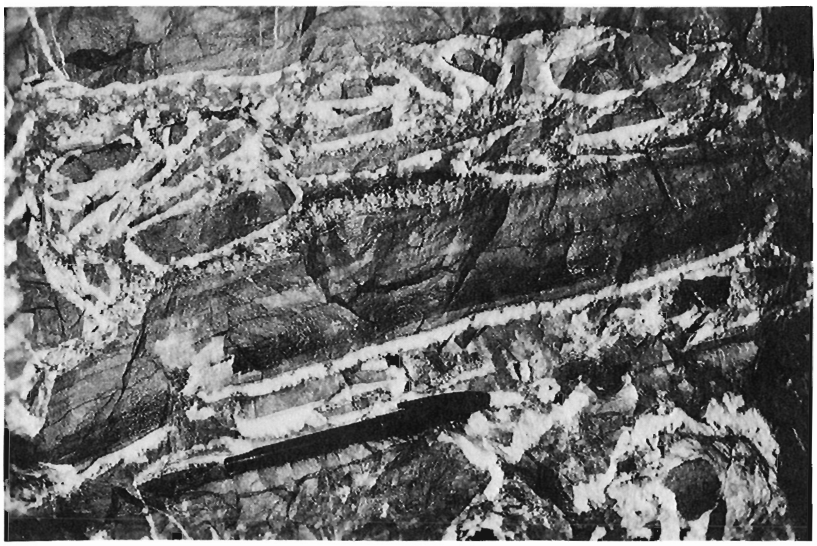

Figure 64.2.

Concordant collapse breccia cemented with white sparry dolomite and sphalerite; Jefferson City Mine, Tennessee. (GSC-20229B)

(Redstone River), and iron (Snake River), as well as the oil, gas, and sulphur fields of the Peace River and Norman Wells districts of Alberta, British Columbia, and the Northwest Territories. The Athabasca tar sands are also included in the Mackenzie Valley district.

To show their relationship to the proposed Mackenzie Valley district approximately two dozen carbonate-hosted lead-zinc deposits are plotted on Figure 64.1 (including the Pine Point district, which in itself contains over 40 known deposits in an area 50 by 17 km (Skall, 1975)). Most of the deposits have been discovered only within the past five years, so division into sub-districts can only be tentative. Nevertheless, following the lead of Smith (1974, Fig. 2) and Dawson (1975, Fig. 1), the Mackenzie Valley district might be said to contain four reasonably distinguishable sub-districts (Fig. 64.1), namely: North Mackenzie Mountains, South Nahanni, Robb Lake, and Pine Point. Published grade-tonnage figures of several deposits and sub-districts are presented in Table 64.3.

Of the deposits given in Table 64.3, Pine Point has been in continuous production since 1965, Prairie Creek is in an advanced development stage with over 14 000 feet of underground work completed, and the remainder are in an early development stage with drilling still in progress.

In addition to the deposits in these proposed sub-districts, the Mackenzie Valley district also contains literally hundreds of scattered showings and deposits of lead-zinc, barite, and fluorite in carbonates, copper in shales, etc., in both epigenetic and syngenetic forms, a feature in common with the famous Mississippi Valley district of the United States, as documented by Heyl (1968).

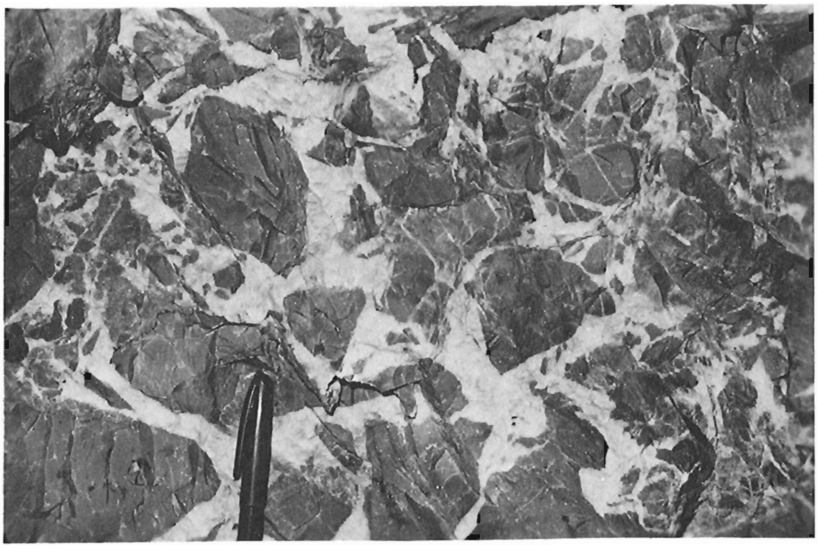

Figure 64.3.

Collapse breccia with white sparry dolomite cement; Jefferson City Mine, Tennessee. (GSC-202229D).

As in the United States Mississippi Valley district (except for the huge Tri-State field), most carbonate-hosted deposits in the Mackenzie Valley district, especially the more economically significant ones, are found in rocks of lower Paleozoic (and upper Proterozoic) to mid-Paleozoic age. However, topography (as expressed by high relief) and climate (which inhibits prolific vegetation) have combined, in the Richardson and Mackenzie mountains, to produce exposures of the favourable rocks unmatched in the Mississippi Valley district. This is compensated in part, however, by the easier access and mining techniques possible in the flat-lying United States district. The (relatively) flat-lying Interior Platform portion of the Mackenzie Valley, on the other hand, contains Pine Point, Canada's most important known carbonate-hosted lead-zinc ore field. Although the depth of cover rocks has hampered the westward extension of the Pine Point ore field, it is likely that a long-term grid-drilling program will extend the sub-district, just as similar programs have done in the central Tennessee, upper Mississippi Valley, and southeast Missouri sub-districts of the Mississippi Valley district.

This comparison of the Mackenzie and Mississippi Valley districts does not, however, imply that all carbonate-hosted lead-zinc deposits in the former are of the Mississippi Valley-type as defined by (for example) Heyl *et al.* (1974). Recent field observations by the senior author have led him to suggest that deposits in the Mackenzie Valley district are of two main types (Sangster, 1975), namely: 1) Stratabound deposits emplaced post-lithification of host rocks and which appear to be controlled by pre-ore, post-host structures (i.e. - typical Mississippi Valley-type); 2) Deposits which appear to be generally low grade, stratiform, and synsedimentary relative to their host rocks. Many of these are modified by post-ore structures resulting in higher-grade discordant bodies (i.e. - Apline type - see, for example, Maucher and Schneider, 1967; Schneider, 1964; Schulz, 1964). Further amplification of this two-fold division of carbonate-hosted lead-zinc deposits in the Mackenzie Valley district will be presented following the 1976 field season.

Even in this short review, it should be abundantly clear that the Mackenzie Valley district may eventually emerge as Canada's answer to other internationally-known carbonate-hosted lead-zinc districts such as those in the United States, Ireland, Poland, central Europe, and North Africa. Certainly when the other commodities mentioned previously in the text are taken into consideration, the Mackenzie Valley district emerges as a major Canadian mineral resource district challenging that of the St. Lawrence Valley in potential economic importance.

References

Brock, J.S.
1975: Mining: Yukon's first industry; Western Miner, Feb., p. 63-66.

Dawson, K.M.
1975: Carbonate-hosted zinc-lead deposits of the Northern Canadian Cordillera; *in* Report of Activities, Part A, Geol. Surv. Can., Paper 75-1A, p. 239-241.

Fyles, J.T.
1970: Geological setting of the lead-zinc deposits in the Kootenay Lake and Salmo areas of British Columbia; *in* Lead-zinc deposits in the Kootenay Arc, ed. A.E. Weissenborn; Wash. Dep. Nat. Resour., Bull. 61, p. 45-53.

Heyl, A.V.
1968: Minor epigenetic, diagenetic, and syngenetic sulfide, fluorite, and barite occurrences in the Central United States; Econ. Geol., v. 63, p. 585-594.

Heyl, A.V., Landis, G.P., and Zartman, R.E.
1974: Isotopic evidence for the origin of Mississippi Valley-type mineral deposits: A review; Econ. Geol., v. 69, p. 992-1006.

Maucher, A. and Schneider, H.J.
1967: The Alpine lead-zinc ores; *in* Genesis of stratiform lead-zinc-barite-fluorite deposits, ed. J.S. Brown; Econ. Geol. Monogr. 3, p. 71-81.

Sangster, D.F.
1975: Canadian carbonate-hosted lead-zinc deposits: A summary; Geol. Soc. Amer., Geol. Assoc. Can., Mineral. Assoc. Can., Abstract with Programs, v. 7, no. 6, p. 848.

Schneider, H.J.
1964: Facies differentiation and controlling factors for the depositional lead-zinc concentration in the Ladinian geosyncline of the eastern Alps; *in* Developments in sedimentology, v. 2, Sedimentology and ore genesis, ed. G.C. Amstutz; Elsevier Pub. Co., Amsterdam, p. 29-45, 184 p.

Schulz, O.
1964: Lead-zinc deposits in the calcareous Alps as an example of submarine-hydrothermal formation of mineral deposits; *in* Developments in sedimentology, v. 2, Sedimentology and ore genesis, ed. G.C. Amstutz; Elsevier Pub. Co., Amsterdam, p. 48-52, 184 p.

Skall, H.
1975: The paleoenvironment of the Pine Point lead-zinc district; Econ. Geol., v. 70, p. 22-47.

Smith, C.L.
1974: Newly discovered Mississippi Valley-type Zn-Pb province, Northwest Territories, Canada; Paper presented to the Soc. Min. Eng. Fall Meeting, Acapulco, Mexico, Preprint 74-I-306.

Taylor, G.C., Macqueen, R.W., and Thompson, R.I.
1975: Facies changes, breccias, and mineralization in Devonian rocks of Rocky Mountains, northeastern British Columbia; *in* Report of Activities, Part A, Geol. Surv. Can., Paper 75-1A, p. 577-585.

3. A brief comparison of selected U.S. and Canadian carbonate-hosted lead-zinc deposits

During the 1975 field season the senior author made a two-week trip to five classical U.S. carbonate-hosted lead-zinc districts, namely: Austinville, VA.; eastern Tennessee; upper Mississippi Valley (Wisconsin-Illinois); Viburnum Lead Belt (southeast Missouri); and Tri-State (Kansas-Oklahoma Missouri). This was done in the company of A.V. Heyl and H. Wedow, Jr. of the U.S. Geological Survey, who have devoted a considerable part of their careers to the study of the upper Mississippi Valley and eastern Tennessee districts, respectively. The group also included three Polish geologists actively working in the famous Silesia-Cracow lead-zinc district of Poland: Mr. T. Galkiewicz (Senior Mine Geologist), Dr. F. Ekiert (Geological Institute), and Dr. J. Pawlowska (Geological Institute).

The several mine visits, surface exposures, and informal discussions which took place during the trip allowed the writer to compare the geology of Canadian carbonate-hosted lead-zinc deposits with the classical, "type deposits" of the United States and Poland.

The purpose of this report is not to repeat in detail the senior author's notes on the geology of the deposits and comparisons with Canadian ones, but rather to mention a few impressions which may be of interest to Canadian geologists.

With the exception of the Austinville-Idaho district of Virginia, the genetic model espoused by geologists familiar with the United States districts was a variation of the paleoaquifer theme (involving solution of selected strata by circulating groundwater and resultant collapse of overlying beds) which has been developed so well in the eastern Tennessee district (Wedow, 1971). In the upper Mississippi Valley district, solution thinning of certain strata has long been advocated (Heyl *et al.*, 1959; Heyl, 1968), but the form of the orebodies is difficult to explain by any normal process (*see*, for example, Plates 10 and 11 in Heyl *et al.*, 1959). At the Buick mine in the Viburnum Lead Belt, the mine geologist advocated solution of gypsum layers, developed in a back-reef evaporite facies, to produce the spectacular mineralized breccias in this particular mine. The solution process would have had to be extremely efficient because no evaporite minerals have yet been found anywhere in the southeast Missouri Lead Belt.

With regard to the famous collapse breccias and associated zinc deposits of eastern Tennessee, one of us (DFS) noted three features of interest: 1) the unconformity to which the breccia structures were related was as much as 1000 feet <u>above</u> the mineralized breccias; 2) the unconformity was in reality a disconformity and, when seen locally in outcrop, it was extremely innocuous and could easily be missed by the casual observer; 3) forms of the orebodies in the Mascot-Jefferson City district, particularly in the eastern part, as exemplified by the New Jersey Zinc Company's Jefferson City Mine, were not as discordant as we had expected solution collapse breccia structures to be (Fig. 64.2).

Most of the ore (80 per cent was the figure quoted) is derived from collapse zones that are "stratiform in that their major dimensions are planar and are controlled by stratigraphy" (Crawford and Hoagland, 1968, p. 250). These essentially stratiform manto orebodies "are largely confined to a limestone bed, about 20 feet thick, between two layers of fine grained primary dolomite.

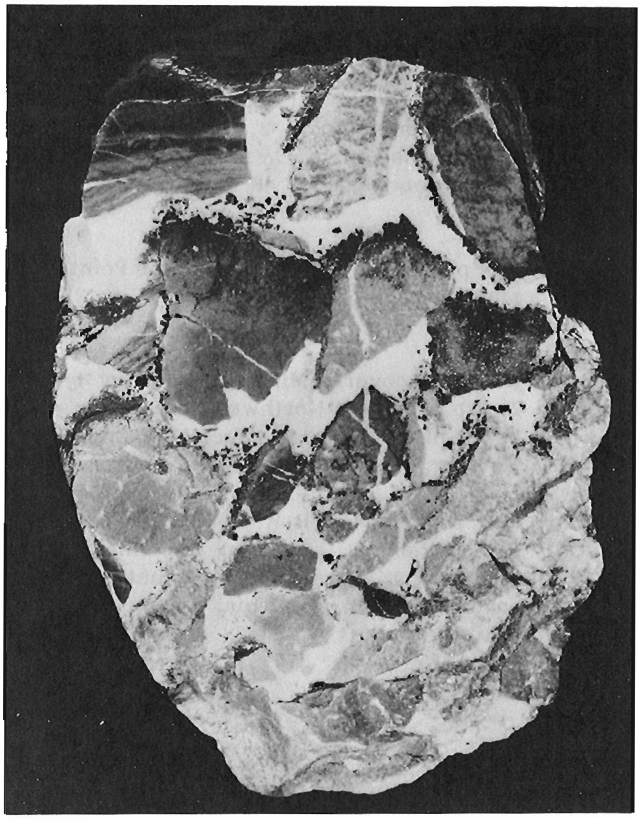

Figure 64.4. Robb Lake breccia cemented with white sparry dolomite and dark sphalerite. Specimen is approximately seven inches (18 cm) in length and is oriented right-way-up. (GSC-202229E)

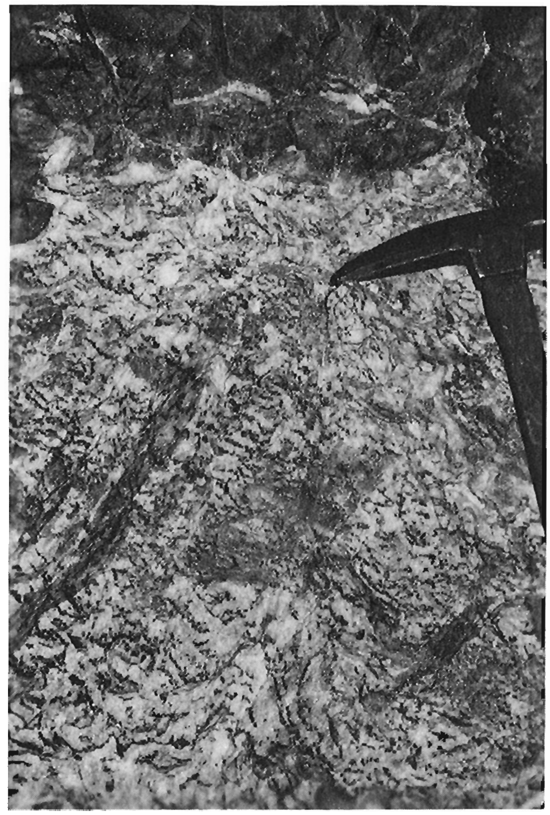

Figure 64.5. Ore-bearing secondary "met" dolomite in sharp contact with microcrystalline Cambrian dolomite; Austinville mine, Idaho. (GSC-202229C)

The mantos extend for many hundreds, or even thousands, of feet in a network of runs and crossings" (ibid; emphasis by the present authors). The planar, concordant nature of the angular collapse breccia, cemented with white sparry dolomite and light-coloured sphalerite struck the writers as extremely similar to breccias of the Robb Lake zinc deposits of northeastern British Columbia (Figs. 64.3 and 64.4). The Robb Lake breccias were originally described as due to collapse (Sangster, 1973) but this concept has been mildly challenged recently by Taylor *et al.* (1974, 1975) In response, the writers point to those features which are similar to those of eastern Tennessee i.e. to the broadly conformable nature of the breccia zones and the similarity in breccia textures on a small scale (including a lack of "fines" in the breccia in both districts). Also, as pictured by Crawford *et al.* (1969, Fig. 4), the "bedded-type ore bodies" (i.e. - conformable breccia zones), although best developed in the so-called "U" bed, also occur to a more limited extent in the overlying "R" and "S" beds. That is to say, mineralization can occur in concordant breccias at two or more stratigraphic horizons, a characteristic feature of the Robb Lake deposits as well (Sangster, 1972; Taylor *et al.*, 1975). In the Jefferson City Mine, the vertical, possibly structurally-controlled "break-through-" type orebodies are narrow (a few hundred feet at most) relative to the flat, widespread distribution of the "conformable" ore. In summary, all the features of the Robb Lake breccias noted by Taylor *et al.* (1974) were noted in the Jefferson City "conformable" ore zones, namely: "The lack of fine matrix "trash" associated with the cement; general conformable nature of the breccias and a geometry consistent with a very large lateral dimension relative to the vertical dimension; lack of internal organization of fragments such as a preferred tectonic fabric or evidence of gravitational settling; lack of involvement of overlying and underlying beds (contact between brecciated and unbrecciated dolomite is often sharp and may follow bedding or cut obliquely to it); and the apparent lack of channelway or cavern development (i.e. - no steep-sided ovoid structures were observed)." If the two districts are, in fact, similar in origin, the ratio of "conformable-" to "breakthrough-" type breccias in the eastern Tennessee district may help to explain the "apparent lack of channelway or cavern development" noted at Robb Lake (Taylor *et al.*, 1974).

The senior writer found the deposits at Austinville-Ivanhoe, VA. (Brown and Weinberg, 1968), which occur in Cambrian dolomite, to be most similar to those of Newfoundland Zinc in spite of the fact the latter occur at the same stratigraphic position and occur below the same disconformity as the deposits of eastern Tennessee,

Figure 64. 6.

Ore-bearing secondary "pseudobreccia" dolomite in sharp contact with microcrystalline Ordovician dolomite; Newfoundland Zinc mine, Newfoundland. (GSC-202229A)

with which they have been compared (Collins and Smith, 1973). The Newfoundland deposits are similar to those at Austinville in that, in both areas, mineralization occurs almost solely in a white, sparry, secondary dolomite, containing irregular patches of the original dark grey, primary dolomite. This secondary dolomite, which is essentially concordant in both districts, is referred to as "metamorphic dolomite" or "met" in Austinville, and "pseudobreccia" at Newfoundland Zinc. Nevertheless, the similarity between the two is striking (Figs. 64. 5 and 64. 6). In our opinion, Newfoundland Zinc is dissimilar to eastern Tennessee in that, although collapse breccias have been recognized at the former (Collins and Smith, 1972, 1973), to date no significant mineralization has been found in these features. In fact, relationship between the collapse breccias and mineralization, beyond spatial proximity, has yet to be demonstrated. Until this relationship is established, a genetic connection between the unconformity and mineralization must be regarded as an open question. Matrix in the collapse breccias at Newfoundland Zinc is largely dark, fine grained, locally-derived dolomite (Collins and Smith, 1973), whereas in eastern Tennessee it is largely secondary dolomite and sphalerite.

A useful byproduct of the tour of United States lead zinc districts was the opportunity provided for informal discussion with the Polish delegation regarding the famous Silesia-Cracow district in the Triassic carbonates of Poland. During these discussions, the great similarity between the Polish deposits and those in eastern Tennessee was emphasized. In particular, the occurrence of ore in dolomite breccias where the latter are underlain by limestone was selected as a feature common to both districts. The Polish breccias, long a subject of controversy, have recently been quite convincingly re-interpreted as collapse breccias by Sass-Gustkiewicz (1974). Readers who have access to the literature will notice the striking similarity between the reconstructed mine section presented by Sass-Gustkiewicz (1974, Fig. 5) and the "high-domal ore structure" pictured by McCormick *et al.* (1971, Fig. 5) for the eastern Tennessee district.

References

Brown, W.H. and Weinberg, E.L.
1968: Geology of the Austinville-Ivanhoe district, Virginia; *in* Ore deposits in the United States, ed. J.D. Ridge; v. I, A.I.M.E., N.Y., p. 169-186, 991 p.

Collins, J.A. and Smith, L.
1972: Sphalerite as related to the tectonic movements, deposition, diagenesis, and karstification of a carbonate platform; 24th Int. Geol. Cong., Montreal, Sect. 6, p. 208-215.

1973: Lithostratigraphic controls of some Ordovician sphalerite; *in* Ores in sediments, ed. G.C. Amstutz and A.J. Bernard; Springer-Verlag, Berlin, p. 79-91, 350 p.

Crawford, J., Fulweiler, R.E., and Miller, H.W.
1969: Mine geology of the New Jersey Zinc Company's Jefferson City Mine; *in* Papers on the stratigraphy and mine geology of the Kingsport and Mascot Formations (Lower Ordovician) of East Tennessee; Tenn. Dep. of Conservation, Div. Geol., Rep. Invest. 23, p. 64-75.

Crawford, J. and Hoagland, A.D.
1968: The Mascot-Jefferson City zinc district, Tennessee; *in* Ore deposits in the United States, ed. J.D. Ridge; v. I, A.I.M.E., N.Y., p. 242-256, 991 p.

Heyl, A.V.
1968: The Upper Mississippi Valley base-metal district; *in* Ore deposits in the United States, ed. J.D. Ridge; v. I, A.I.M.E., N.Y., p. 431-459, 991 p.

Heyl, A. V., Agnew, A. F., Lyons, E. J., and Behre, C. H.
1959: The geology of the Upper Mississippi Valley zinc-lead district; U. S. Geol. Surv., Prof. Paper 309.

McCormick, J. E., Evans, L. L., Palmer, R. A., and Rasnick, F. D.
1971: Environment of the zinc deposits of the Mascot-Jefferson City district, Tennessee; Econ. Geol., v. 66, p. 757-762.

Sangster, D. F.
1972: Robb Lake, B. C.; *in* Report of Activities, Part A, Geol. Surv. Can., Paper 73-1A, p. 129.

Sass-Gustkiewicz, M.
1974: Collapse breccias in the ore-bearing dolomite of the Olkusz mine (Cracow-Silesian ore-district); Ann. Soc. Geol. de Pologne, v. XLIV, p. 217-226.

Taylor, G. C., Macqueen, R. W., and Thompson, R. I.
1975: Facies changes, breccias, and mineralization in Devonian rocks of Rocky Mountains, northeastern British Columbia; *in* Report of Activities, Part A, Geol. Surv. Can., Paper 75-1A, p. 577-585.

Taylor, G. C., Thompson, R. I., and Macqueen, R. W.
1974: Geological details building up about zinc-lead breccias in northeast B. C.; Northern Miner, Nov. 28, p. 60-61.

Wedow, H., Jr. (Chairman)
1971: A paleoaquifer and its relation to economic mineral deposits: The Lower Ordovician Kingsport Formation and Mascot dolomite; Econ. Geol., v. 66, p. 695-810.

65. OCCURRENCE OF SELENIDE AND SULPHIDE MINERALS IN BORNITE ORE OF THE KIDD CREEK MASSIVE SULPHIDE DEPOSIT, TIMMINS, ONTARIO

Projects 680060, 620308

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An elongate bornite-rich ore zone on the 1200 level of the Kidd Creek mine is completely flanked by chalcopyrite-rich ore. In the contact zone between bornite-rich and chalcopyrite-rich ores, the mineralogy is particularly complex and this contact zone is characterized by the presence of macroscopic amounts of tennantite, carrollite and, in places, clausthalite. The bornite ore contains many rare minerals, some new minerals, and common sulphides with unusual compositions. The presence of selenides, tin sulphides, cobalt-bearing minerals, and rare indium minerals is especially noteworthy.

Ore from the bornite zone forms a very small part of the total ore to be mined from the 1200 level of the Kidd Creek mine. The bulk of the ore consists of normal copper and copper-zinc massive sulphide types described by Walker and Mannard (1974).

About 160 polished sections and 50 polished thin sections have been studied optically, and about 80 of these have been surveyed with a MAC electron microprobe equipped with an energy dispersive spectrometer. Use of the latter system is an efficient method to obtain qualitative and semi-quantitative results on minerals that are unknown or are difficult to recognize optically, and to select significant grains for X-ray diffraction and for quantitative analysis by conventional methods.

Minerals that have been identified in and near the bornite body are listed in Table 65.3. The quantitative analyses of a number of these minerals are listed in Table 65.1. Several of these analyses were obtained from grains less than 50 μm in maximum dimension, and any contribution of the major elements in surrounding minerals to the reported minor element concentrations has not yet been evaluated in detail. This is the first reported Canadian locality for the minerals colusite, bohdanowiczite ($AgBiSe_2$), tungstenite, and cadmoselite. Bismuth and bismuthinite have been identified in more disseminated ore lying east of the bornite zone. Quartz, albite, calcite, dolomite, sericite, chlorite, and tourmaline are important gangue constituents in the ore.

Some mineral phases that have been encountered in the ore are not yet positively identified, and will be the object of further study. Some of these phases are listed in Table 65.3. In addition to these minerals, the ore contains compositional variants of a number of established, often common, mineral species. These varietal phases include selenian chalcopyrite, bornite, digenite, mawsonite and stannoidite, and cobaltian pyrite. In addition, silver-bearing chalcopyrite and bornite which appear to be optically homogeneous have been analyzed (analyses 4, 5, 6, 7, Table 65.1) but those with higher silver contents are not stoichiometric and may be sub-microscopic mixtures.

Selenium in the bornite ore forms a number of selenide minerals and also occurs as a minor constituent in other minerals. Naumannite, eucairite, clausthalite, klockmannite, and bohdanowiczite are the most commonly encountered selenide minerals, whereas cadmoselite and a Ag-Au selenide, possibly fischesserite (Ag_3AuSe_2), are very rare. Selenium is also a significant constituent in unknown Cu-S-Se, Cu-W-Sn-S-Se and Cu-In-S-Se phases, and in an unknown Pb-Bi-Se S phase which has a rim of variable composition. To date the maximum selenium content that has been determined for bornite is 2.6 per cent (analysis 3), for chalcopyrite 0.5 per cent (analysis 1), for mawsonite 1.3 per cent (analysis 18), and for stannoidite 1.55 per cent (analysis 20).

The tin-bearing minerals stannoidite (analysis 20) and mawsonite (analysis 18) are widely distributed in minor amounts in the bornite ore. Colusite (analysis 22) and kesterite (analysis 21) are less widely distributed and an unknown Cu-W-Sn-S-Se phase has been observed as a single grain. A mawsonite-like phase with a lower tin content than normal (analysis 19), and several "colusite" phases, one with a greenish-grey colour and higher tin content than normal, are also present. Only a few grains of cassiterite have been encountered, although elsewhere in the massive sulphide ore this is the principal tin mineral, and tin is recovered at the mine as cassiterite (Mulligan, 1975).

Cobalt occurs in the ore as carrollite (Fig. 65.1), cobaltite, cattierite, and unknown cobalt sulphide, Cu-Co-As-S, and Cu-Co-Fe-As-S phases. In addition, up to 6.5 per cent cobalt (analysis 24) has been determined in pyrite having a distinctly lower reflectance (7, Table 65.2 and Fig. 65.1) than normal pyrite. Cattierite (analysis 25) is apparently a rare mineral in the ore. The unknown cobalt sulphide occurs as a light yellow phase closely associated with carrollite (Fig. 65.1), from which it has apparently exsolved. This mineral has a reflectance (9, Table 65.2) above that of cobalt pentlandite and the composition (analysis 26) is closer to that of the uncertain mineral jaipurite (CoS) than to cobalt pentlandite (Co_9S_8). The unknown Cu-Co-Fe-As-S phase occurs as flesh-coloured rims (analysis 30) around some tennantite grains, as brownish equant inclusions in carrollite grains, and as violet-tinted discrete grains in chalcopyrite near large carrollite grains. The apparent colours may be largely due to the association in which the grains are viewed, as no difference in reflectance was measured. Cobaltite in the ore most commonly occurs as scattered equant grains in gangue. In one specimen cobaltite was found

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Table 65. 1

Representative analyses (wt. %) of minerals occurring within the bornite zone

	1	2	3	4	5	6	7	8	9	10
Cu	34.4	63.3	61.1	34.5	34.4	32.0	61.2	75.3	63.6	68.4
Fe	30.2	11.2	11.4	30.4	30.3	28.7	10.2	1.1	0.7	0.2
Ag	–	0.0	–	0.1	0.8	5.4	1.2	0.05	0.2	–
S	34.4	25.2	25.4	34.2	34.2	32.7	24.6	20.9	17.6	30.0
Se	0.5	0.3	2.6	–	–	–	1.8	2.8	17.8	0.9
Total	99.5	100.0	100.5	99.2	99.7	98.8	99.0	100.15	99.0	99.5

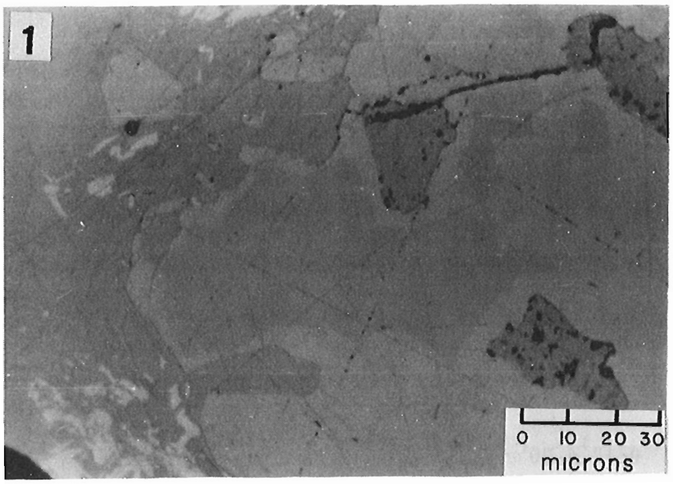
	11	12	13	14	15	16	17
Pb	–	–	67.9	0.0	19.1	8.6	12.9
Bi	–	–	3.7	44.5	52.8	69.2	66.2
Ag	44.4	73.5	1.05	22.5	0.3	0.1	0.1
Cu	24.7	0.0	0.04	0.0	3.7	0.1	0.1
Fe	0.02	0.03	0.02	0.0	0.05	0.1	0.1
Se	31.4	27.6	26.8	32.5	15.3	18.8	18.5
S	0.07	0.02	1.3	1.2	10.2	2.5	2.7
Total	100.59	101.15	100.81	100.7	101.45	99.4	100.6

	18	19	20	21	22
Cu	42.3	42.4	38.0	27.6	47.8
Fe	13.1	13.7	8.6	1.1	0.3
Zn	0.0	0.0	4.3	13.2	0.7
V	–	–	–	–	3.3
Sn	14.5	7.7	18.7	28.4	9.0
As	0.15	4.5	0.5	0.2	7.3
Se	1.3	1.2	1.55	0.4	2.4
S	28.5	29.8	27.9	28.9	27.9
Total	99.85	99.30	99.55	99.80	98.7

	23	24	25	26	27	28	29	30	31
Cu	–	–	0.6	0.8	12.0	43.7	45.7	35.1	1.4
Fe	41.9	40.4	15.7	0.7	0.6	0.1	4.4	6.6	15.8
Zn	–	–	–	–	–	8.1	5.1	0.1	–
Ni	–	–	0.4	1.6	2.1	–	–	2.5	16.7
Co	4.7	6.5	29.0	60.2	40.0	0.0	0.0	13.0	1.7
Ag	–	–	–	–	–	–	–	0.2	–
As	0.1	0.3	6.1	0.1	0.15	19.7	16.2	17.5	44.8
Se	–	–	–	2.6	3.8	–	–	–	–
S	53.5	53.4	47.3	32.3	41.2	27.8	28.0	23.6	19.4
Total	100.2	100.6	99.1	98.3	99.85	99.4	99.4	98.6	99.8

Table 65.1 (opposite)

1. Selenian chalcopyrite, $Cu_{1.00}Fe_{1.00}(S_{1.99}Se_{0.01})_{2.00}$, S320-2
2. Selenian bornite, $Cu_{5.03}Fe_{1.02}(S_{3.98}Se_{0.02})_{4.00}$, K1010-225.5
3. Selenian bornite, $Cu_{4.66}Fe_{0.99}(S_{3.84}Se_{0.16})_{4.00}$, S320-2
4. Chalcopyrite, $Cu_{1.02}Fe_{1.02}Ag_{0.002}S_{2.00}$, TQ75-1(1)
5. Chalcopyrite, $Cu_{1.02}Fe_{1.02}Ag_{0.014}S_{2.00}$, TQ75-1(1)
6. Argentian chalcopyrite, $Cu_{0.99}Fe_{1.01}Ag_{0.10}S_{2.00}$, TQ75-1(1)
7. Argentian selenian bornite, $Cu_{4.88}Fe_{0.92}Ag_{0.06}(S_{3.88}Se_{0.12})_{4.00}$, TQ74-679(6)
8. Selenian digenite, $Cu_{1.72}Fe_{0.03}Ag_{0.001}(S_{0.95}Se_{0.05})_{1.00}$, K1010-225.5
9. Unknown, $Cu_{1.29}Fe_{0.02}Ag_{0.002}(S_{0.71}Se_{0.29})_{1.00}$, K1010-225.5
10. Covelline, $Cu_{1.14}Fe_{0.004}(S_{0.99}Se_{0.01})_{1.00}$, S320-2
11. Eucairite, $Ag_{1.03}Cu_{0.97}Fe_{0.001}(Se_{0.995}S_{0.005})_{1.00}$, TQ74-561(6)
12. Naumannite, $Ag_{1.95}Fe_{0.001}(Se_{0.998}S_{0.002})_{1.00}$, TQ74-561(6)
13. Clausthalite, $Pb_{0.86}Bi_{0.05}Ag_{0.03}Cu_{0.002}Fe_{0.001}(Se_{0.89}S_{0.11})_{1.00}$, K1010-231(1)
14. Bohdanowiczite, $Bi_{0.95}Ag_{0.93}(Se_{1.83}S_{0.17})_{2.00}$, K1010-231(1)
15. Unknown $(Bi_{2.47}Pb_{0.90}Cu_{0.57}Ag_{0.03}Fe_{0.01})_{3.98}(S_{3.11}Se_{1.89})_{5.00}$, K1010-238(2)
16. Rim on No. 15, $Bi_{4.24}Pb_{0.53}Cu_{0.02}Fe_{0.02}Ag_{0.01}Se_{3.05}S_{1.00}$, K1010-238(2)
17. Rim on No. 15, $Bi_{3.76}Pb_{0.74}Cu_{0.02}Fe_{0.02}Ag_{0.01}Se_{2.78}S_{1.00}$, K1010-238(2)
18. Mawsonite, $Cu_{5.88}Fe_{2.07}As_{0.02}Sn_{1.08}Se_{0.15}S_{7.85}$, TQ74-681(2)
19. Low-Sn mawsonite, $Cu_{5.65}Fe_{2.08}As_{0.51}Sn_{0.55}Se_{0.13}S_{7.87}$, S196
20. Stannoidite, $Cu_{8.07}Fe_{2.08}Zn_{0.89}As_{0.09}Sn_{2.13}Se_{0.26}S_{11.74}$, TQ74-681(2)
21. Kesterite, $Cu_{1.92}Fe_{0.09}Zn_{0.89}As_{0.01}Sn_{1.06}Se_{0.02}S_{3.98}$, K863-148
22. Colusite, $Cu_{3.34}Fe_{0.02}Zn_{0.05}As_{0.43}Sn_{0.34}V_{0.29}Se_{0.14}S_{3.86}$, S196
23. Cobaltian pyrite, $Fe_{0.90}Co_{0.10}As_{0.002}S_{1.998}$, TQ74-560(4)
24. Cobaltian pyrite, $Fe_{0.87}Co_{0.13}As_{0.005}S_{1.995}$, S196-1
25. Ferroan cattierite, $Co_{0.63}Fe_{0.36}Cu_{0.01}Ni_{0.01}As_{0.10}S_{1.90}$, TQ74-561(2)
26. Unknown $(Co_{0.98}Ni_{0.03}Fe_{0.01}Cu_{0.01})_{1.03}(As_{0.001}Se_{0.032}S_{0.967})_{1.000}$, S195-7
27. Carrollite, $Co_{2.03}Cu_{0.57}Ni_{0.11}Fe_{0.03}As_{0.006}Se_{0.144}S_{3.85}$, S195-7
28. Tennantite, $Cu_{10.31}Fe_{0.03}Zn_{1.86}As_{3.94}S_{13.00}$, S300
29. Fe-tennantite, $Cu_{10.71}Fe_{1.17}Zn_{1.16}As_{3.22}S_{13.00}$, S300
30. Unknown, $Cu_{9.76}Fe_{2.09}Zn_{0.03}Ni_{0.75}Co_{3.90}Ag_{0.03}As_{4.13}S_{13.00}$, K863-124.5
31. Unknown, $Ni_{0.47}Fe_{0.47}Co_{0.05}Cu_{0.04}As_{0.99}S_{1.00}$, K55-97-1793A(2)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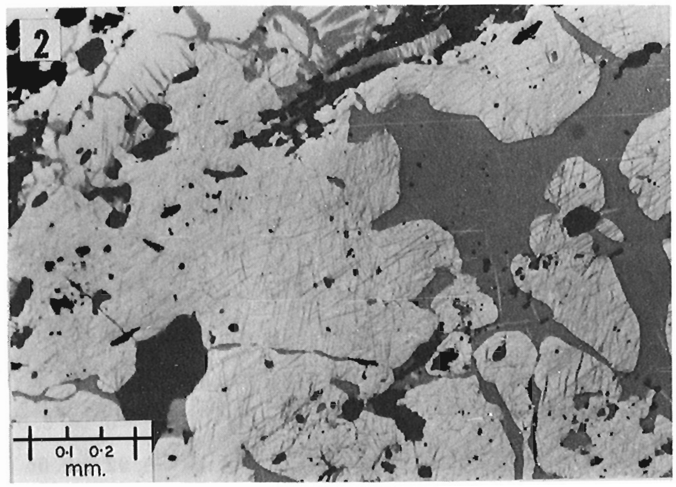
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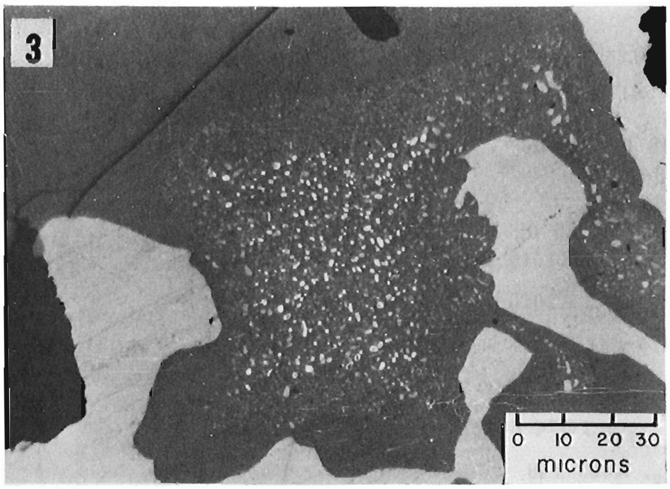
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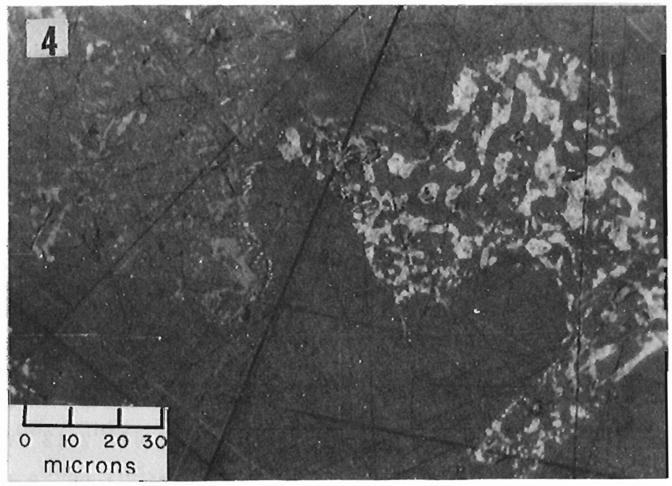
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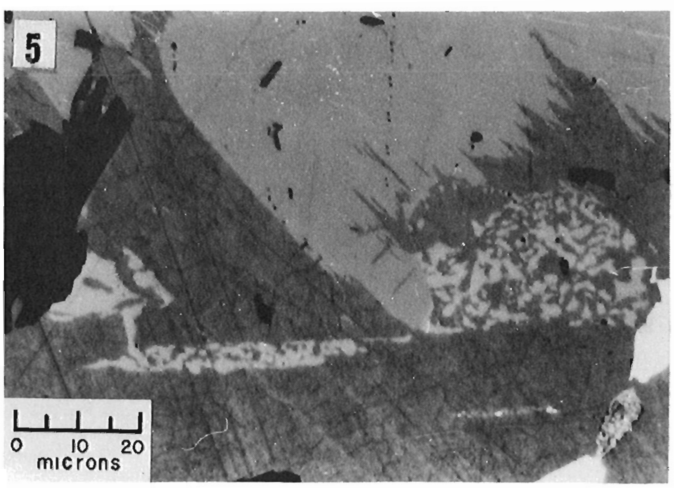
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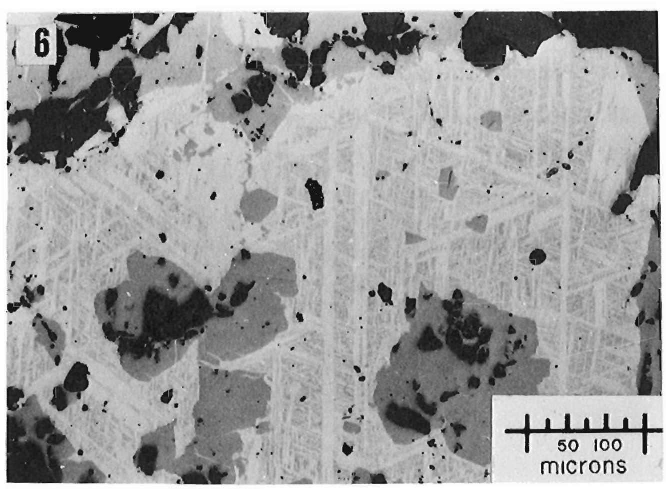
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Table 65.2

Reflectance measurements of selected minerals

Wavelength	1	2	3	4	5	6	7	8	9	10	11
470	53.2	55.9	51.2	52.9	23.2	23.7	43.6	43.3	55.1	49.8	52.0
546	48.9	50.7	49.8	51.5	22.0	22.4	46.2	43.2	55.5	49.6	50.4
589	47.7	50.0	50.1	51.5	21.6	21.7	47.6	43.7	57.9	49.7	50.7
650	46.2	47.8	50.1	51.6	21.4	21.5(5)	49.0	44.4	60.8	50.2	51.1

Wavelength	12	13	14	15	16	17	18	19	20	21	22
470	33.6	31.0	36.4	24.6	30.3	26.2	20.0	23.1	36.2	36.3	23.5
546	35.3	32.6	40.0	26.8	29.7	25.2	23.0	26.0	33.7	33.3	23.4
589	36.7	34.4	41.6	28.1	28.7	25.1	24.9	27.9	32.5	31.8	23.6
650	38.4	36.2	44.0	29.7	27.2	26.1	27.8	32.5	30.7	31.2	23.7

1, 2. Clausthalite; 3, 4. two orientations on one grain of bohdanowiczite; 5, 6. roquesite; 7. cobaltian pyrite; 8. carrollite; 9. unknown cobalt sulphide; 10, 11. unknown Co-Ni-As-S mineral; 12, 13. maximum and minimum of 8 measurements on unknown Cu-Co-Fe-As-S phase; 14. unknown creamy Cu-Co-As-S phase; 15. colusite; 16. tennantite; 17. enargite; 18, 19. two orientations on one grain of mawsonite; 20, 21. naumannite; 22. unknown Cu-W-Sn-S-Se.

Figure 65.1. Pyrite (smooth, light grey) including darker patches of cobaltian pyrite. White phase is an unknown cobalt sulphide associated with carrollite, which is medium grey with open cleavage and not clearly distinguished from chalcopyrite as at upper left. Dark grey pitted grains are naumannite. Specimen S195(7). Photo GSC 165904.

Figure 65.2. Carrollite with slightly open cleavage. Bornite (dark grey) and chalcopyrite (smooth white) are interstitial. Specimen K863-114.5. Photo GSC 165851.

Figure 65.3. Area of exsolved native silver and naumannite in bornite. Large light grey areas are naumannite. Specimen TQ74-679(6). Photo GSC 165847.

Figure 65.4. Eucairite (light grey) and digenite (medium grey) intergrown with bornite. Specimen K863-93. Photo GSC 165832.

Figure 65.5. Roquesite (medium grey, top) with a partly flame-like contact with bornite. Unknown light grey Cu-In-S-Se phase in myrmekitic intergrowth with bornite. White is chalcopyrite. Specimen K863-93. Photo GSC 165842.

Figure 65.6. Bornite (medium grey) and large areas consisting of a fine lamellar intergrowth of bornite and digenite and larger chalcocite blades. Specimen TQ74-560(1). Photo GSC 165924.

to be finely intergrown with bohdanowiczite and clausthalite.

Silver occurs in the bornite ore in naumannite, eucairite, bohdanowiczite, native silver, electrum, fischesserite (?), and acanthite. The first three of these minerals are undoubtedly the most important silver-bearing phases, whereas the last three are very rare. Another rare Cu-Fe-Ag-As-S phase (possibly argentian tennantite) also contains silver. Silver is also present in chalcopyrite and bornite either due to submicroscopic mixtures or complex substitution. Chalcopyrite containing a moderate amount of silver (analysis 5) rapidly tarnishes brown, whereas chalcopyrite with a higher content of silver (analysis 6) tarnishes iridescent blue to purple. The most silver-rich bornite encountered (analysis 7) has a medium blue colour. Naumannite in the ore often occurs in association with eucairite, clausthalite and bohdanowiczite. Naumannite and bornite together sometimes form a net texture around grains of chalcopyrite, and in a few specimens naumannite occupies a grain junction position in chalcopyrite that has apparently been annealed. In certain grains of bornite, or certain areas within grains, very fine blebs of native silver and naumannite have been exsolved (Fig. 65.3). These textures suggest considerable solid solubility of silver and selenium in chalcopyrite and bornite at elevated temperatures.

Geochemically the ore of the bornite zone is of interest because of its enrichment in copper, selenium, cobalt, arsenic and silver, relative to the normal massive sulphide ore. Vanadium is present in the bornite ore in small amounts of colusite and related phases, and indium in small amounts of roquesite and an unknown Cu-In-S-Se phase. It is not known whether these elements are enriched relative to the normal ore. Very

Table 65.3

Minerals identified from within and near the bornite zone, Kidd Creek mine, Timmins

	Theoretical Formula	Confirmed by X-ray	See analysis no.	See reflectance no.
Major minerals				
Bornite	Cu_5FeS_4	x		
Chalcopyrite	$CuFeS_2$	x		
Ubiquitous, minor in quantity				
Tennantite	$Cu_{12}As_4S_{13}$	x	28, 29	16
Carrollite	$CuCo_2S_4$	x	27	8
Naumannite	As_2Se	x	12	20, 21
Eucairite	AgCuSe	x	11	
Cobaltite	CoAsS	x		
Mawsonite	$Cu_6Fe_2SnS_8$	x	18	18, 19
Stannoidite	$Cu_8(Fe, Zn)_3Sn_2S_{12}$	x	20	
Sphalerite	ZnS			
Rare, generally minor				
Clausthalite	PbSe	x	13	1, 2
Pyrite	FeS_2	x		
Roquesite	$CuInS_2$	x		5, 6
Chalcocite	Cu_2S			
Digenite	Cu_9S_5	x	8	
Silver	Ag			
Electrum	(Au, Ag)			
Klockmannite	CuSe			
Rutile	TiO_2	x		
Cassiterite	SnO_2			
Enargite	Cu_3AsS_4	x		17
Colusite	$Cu_3(As, Sn, V, Fe, Sb)S_4$		22	15
Bohdanowiczite	$AgBiSe_2$		14	3, 4
Unknown Cu-Co-Fe-As-S			30	12, 13
Unknown Cu-Co-As-S				14
Unknown Co-Ni-As-S				10, 11
Very rare				
Arsenopyrite	FeAsS			
Gold	Au			
Unknown cobalt sulphide			26	9
Tungstenite	WS_2			
Cadmoselite	CdSe			
Nickeline	NiAs			
Galena	PbS			
Cattierite	$(Co, Fe)S_2$	x	25	
Kesterite	Cu_2ZnSnS_4	x	21	
Stannite	Cu_2FeSnS_4			
Covelline	CuS	x	10	
Bismuth	Bi			
Bismuthinite	Bi_2S_3			
Acanthite	Ag_2S			
Unknown Cu-W-Sn-S-Se				
Unknown Cu-Fe-Ag-As-S				
Unknown Cu-In-S-Se				
Unknown Ni-Fe-As-S			31	
Unknown Pb-Bi-Se-S			15	
Unknown Cu-S-Se			9	
Unknown Ag-Au-Se (fischesserite?)				

locally gold has been enriched in association with selenium, arsenic and cobalt.

Textures in the ore are very complex. Striking myrmekitic intergrowths of bornite with eucairite (Fig. 65.4), an unknown Cu-In-S-Se phase (Fig. 65.5), and an unknown Cu-S-Se phase are present. Chalcocite, digenite and the unknown Cu-S-Se phase form good lamellar intergrowths with bornite (Fig. 65.6). Exsolution processes are no doubt responsible for these textures. Textures other than these suggest that bornite, in addition to the above minerals, may have exsolved chalcopyrite, naumannite, native silver, bohdanowiczite and colusite. Naumannite and native silver are sometimes found as small blebs in specific areas in bornite (Fig. 65.3). Characteristic blades and flames of chalcopyrite are present in bornite. Bornite and naumannite have apparently exsolved from chalcopyrite. Tennantite in some cases includes small blebs of naumannite, silver, enargite and chalcopyrite that may be products of exsolution. Minerals that have possibly been exsolved from carrollite include bornite, chalcopyrite, cobaltian pyrite, an unknown cobalt sulphide, and an unknown Cu-Co-Fe-As-S phase. Stannoidite may have exsolved small amounts of bornite and mawsonite.

Overgrowths and reaction rims are also quite common textural features in the ore. Rims of a Cu-Co-Fe-As-S phase on grains of tennantite are common. An unknown Co-Ni-As-S phase, and rarely a little nickeline, are sometimes associated with these reaction rims. Also, tennantite (analysis 28) often has a darker, more iron-rich, border phase (analysis 29), and in other cases is rimmed by enargite. An unknown Pb-Bi-Se-S phase has a partial rim containing the same elements but of variable composition, and some colusite grains have partial rims of a greenish-grey more tin-rich phase. Rarely a Ni-Fe-As-S phase (analysis 31) occurs as overgrowths on grains of cobaltite. Mawsonite in chalcopyrite-rich specimens is commonly screened from the surrounding chalcopyrite by a narrow border of bornite, presumably a reaction product.

This brief report illustrates many of the complexities of the bornite ore at the Kidd Creek mine. Further study will be done on several aspects of the ore including the identification of many of the unknown phases, the spatial distribution of minerals, and the compositional variability of a number of the minerals.

References

Mulligan, R.
1975: Geology of Canadian tin occurrences; Geol. Surv. Can., Econ. Geol. Rept. No. 28, 155 p.

Walker, R.R. and Mannard, G.W.
1974: Geology of the Kidd Creek mine, a progress report; Can. Inst. Min. Met. Bull., Dec. 1974, p. 41-57.

GOLD CONTENT OF GREYWACKE AND SLATE OF THE GOLDENVILLE FORMATION, NOVA SCOTIA, AS DETERMINED BY NEUTRON ACTIVATION ANALYSIS

Project 680060

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Introduction

Sediments of the Halifax and Goldenville formations, predominantly slate and greywacke respectively, are host rocks for most of the gold showings and deposits in Nova Scotia. The sediments are of flysch type and form a very monotonous, characteristically unfossiliferous, thick sequence that is of Early Paleozoic, possibly Ordovician, age. The sediments have been folded into a series of long anticlines and synclines, generally upright and open, with gradual changes in the direction of plunge. Gold has commonly been found in saddle reefs and strike veins, rarely in cross veins, associated with domal portions of anticlines. A lateral secretion hypothesis for these veins has been seriously considered by many geologists.

The sedimentology of the Halifax and Goldenville formations, which together comprise the Meguma Group, has been described by Schenk (1970). Gold-quartz veins are generally confined to areas lying in the zone of greenschist facies of regional metamorphism (Taylor and Schiller, 1966).

This study was conceived as an undergraduate thesis project for the junior author. He recorded the megascopic characteristics of the individual samples, and did much of the preparation work on these samples prior to their submission for neutron activation analysis.

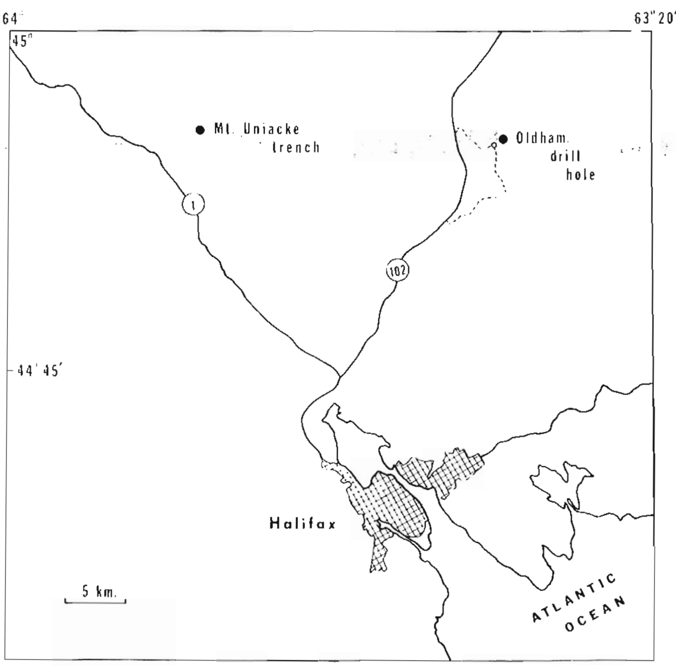

Figure 66.1. Index map showing the locations of the Oldham drillhole and Mount Uniacke trench.

Table 66.1

Comparison of duplicate neutron activation gold analyses

Sample	ppb Au	ppb Au
Oldham hole, 352 (greywacke)	1.4	1.8
Oldham hole, 353 (greywacke)	2.9	0.8
Oldham hole, 202 (slate)	37	30
Mt. Uniacke trench, 223 (greywacke)	0.5	1.7

Sampling and Analytical Procedure

During the 1974 field season, greywacke and slate from the Goldenville Formation were sampled for the determination of background gold content, and as a possible test of the lateral secretion theory. Neutron activation analyses for gold were done early in 1975 on samples from a drillhole at Oldham and from a trench at Mount Uniacke (Fig. 66.1).

The hole at Oldham was drilled to a depth of 2000 feet (609.6 metres) by the Nova Scotia Department of Mines in 1963 and penetrated mainly greywacke, with narrow slate interbeds. This hole was located on or near the axis of the Oldham anticline about 4500 feet (1.4 km) northeast of Oldham and apparently several hundred feet west of the nearest known gold-bearing vein in the district. Core samples of the greywacke were collected at about 50-foot (15 m) intervals and slate interbeds were sampled when they were greater than 2 feet (60 cm) thick and more than 25 feet (7.6 m) apart. At Mount Uniacke 15 samples were taken of the country rock which consisted of interbedded slate and greywacke. These samples, 9 of slate and 6 of greywacke, were taken at irregular intervals along a distance of 90 feet (27.4 m) in a trench running perpendicular to a mineralized quartz vein previously mined for gold.

Samples from the Oldham drillhole consisted of split portions of 4 to 6 inches of core, whereas hand specimens were taken from the Mount Uniacke trench. All samples were passed through a Chipmunk crusher and, after each sample, the jaws were cleaned with a steel wire brush and blown clean with forced air. The samples were then ground in a Braun pulverizer. After each sample of slate was ground, a visible residue on the grinding wheels was removed by grinding a handful of pure crushed quartz. After each sample was ground all contact surfaces were blown free of residual powder using forced air. Final reduction of 15 grams

Table 66.2

Neutron activation analyses of gold for samples from the Oldham drillhole

Sample No.	Footage in hole	Greywacke	Slate	Pyrite	Arsenopyrite	Comments	ppb Au
193	7	x				Minor muscovite	2.7
194	51	x		x		Minor muscovite, carbonate veinlet	<0.2
195	64		x	x		Dark grey	0.9
196	90	x		x		Some muscovite	1.2
197	99		x	x	?	Dark grey to black	1.8
198	140	x					0.9
199	164		x	x	x	Dark grey	7.6
200	208	x				Minor muscovite, carbonate on fractures	0.4
201	254	x					0.4
202	277		x	x	?	Dark grey to black, carbonate films	37
203	314	x				Minor muscovite	<0.2
204	316		x			Black, carbonate films	0.4
205	374	x		x			0.2
306	425	x		x		Minor muscovite	0.3
307	433		x	x			1.3
308	490	x				Carbonate on fractures	0.4
309	525	x		x	?	Sheared, muscovite present	1.6
310	577		x			Minor muscovite, carbonate veinlet	<0.2
311	609		x	x			0.5
312	646	x				Minor muscovite, carbonate veinlets	<0.2
313	675	x				Minor muscovite	<0.2
314	704		x			Some muscovite	1.6
315	724	x		x		Minor muscovite	0.6
316	750		x			Grey to black, carbonate veinlets	2.1
317	774	x				Altered, thin slate interbeds	1.4
318	814		x	x	x	Dark grey to black, asp. in carb. vein	161
319	844	x				Minor muscovite	0.7
320	896	x	x			Muscovite	<0.2
321	899		x	x		Dark grey to black, some carbonate	0.4
322	933	x			x	Slate interbed, carbonate veinlets	<0.5

Table 66.2 (cont'd.)

Sample No.	Footage in hole	Greywacke	Slate	Pyrite	Arsenopyrite	Comments	ppb Au
323	987		x			Grey to black, carbonate veinlets	0.5
324	995	x					<0.3
325	1043	x		x		Some muscovite	0.3
326	1099		x	x		Dark grey, some white carbonate	0.4
327	1109	x					0.3
328	1159	x				Some muscovite	0.5
329	1190		x	x	x	Sheared, grey to black, minor muscovite	<0.5
330	1208	x				Very fine grained	0.7
331	1251		x			Sheared, black	3.2
332	1261	x				Carbonate on fractures	0.5
333	1301		x	x	x	Sheared	0.8
334	1324	x				Minor muscovite	<0.3
335	1377	x		x		Minor muscovite	0.2
336	1428	x				Minor biotite	<0.8
337	1469		x	x		Grey to black, sheared minor biotite	0.4
338	1484	x				Some muscovite	0.2
339	1508		x	x		Dark grey, sheared	0.5
341	1529		x	x		Dark grey	<0.5
342	1585	x				Some muscovite	<0.9
343	1628	x		x		Some muscovite	0.7
344	1636		x	x	x	Dark grey	<0.6
345	1679	x				Minor biotite	2.7
346	1692		x			Black, sheared	1.2
347	1729	x		?	?	Some biotite	<0.8
348	1776	x				Some muscovite	<0.3
349	1787	x				Very fine grained (siltstone)	1.2
350	1829	x				Some muscovite, carbonate on fractures	<0.5
351	1873	x				Some muscovite	0.4
352	1877	x		x		Very fine grained, minor muscovite	1.4
353	1932	x				Minor muscovite	2.9
354	1999	x				Very fine grained	<0.4

Table 66. 3

Neutron activation analyses of gold for samples from the Mount Uniacke trench

Sample No.	Distance from vein (ft.)	Type of sample	ppb Au
210	at contact	slate	4. 4
211	2	5-inch greywacke interbed	3. 9
212	2½	slate	0. 7
213	5	slate	4. 0
214	11	greywacke	23
215	20	greywacke	<0. 4
216	26	slate	283
217	35	slate	140
218	40	greywacke	120
219	50	8-inch slaty interbed	6. 1
220	57	slate	
221	70	8-inch greywacke interbed	1. 2
222	70½	1. 5-foot slate interbed	0. 2
223	80	greywacke, little sulphide	0. 5
224	90	slate	55

of each sample to minus 250-mesh was accomplished by ball-milling. The samples were analyzed by Atomic Energy of Canada Ltd., Commercial Products Division, Ottawa, under the supervision of Dr. V. Armstrong. Approximately one gram of sample was taken for each analysis and the samples were leached with aqua regia for two hours. Irradiation for the neutron activation analysis was by the Slowpoke reactor of Atomic Energy of Canada Ltd.

Results

Gold was determined by neutron activation analysis in a total of 80 samples of slate and greywacke. Sixty-one of these analyses (39 greywacke and 22 slate) were on samples from the Oldham drillhole, 15 analyses were on samples from the Mount Uniacke trench, and 4 were duplicates. The results are presented in Tables 66. 2 and 66. 3 and as line graphs in Figures 66. 2 and 66. 4. The results of duplicate analyses of 4 samples are presented in Table 66. 1. The results for two of the samples are quite similar, but for the other two samples differ by more than a factor of three. Seeland (1973) has considered the problem of reliability of gold analyses for medium grained (0. 37 mm) sandstone and found that, to assure 95 per cent certainty that the true gold content varies no more than 50 per cent from the analytical results obtained (assuming that the gold is present as detrital flakes in hydraulic equilibrium with the sand grains), a sample size of 120 kilograms would be required. This sample size is not practical and greater uncertainty must be accepted. In the case of a much finer grained sedimentary rock, such as slate, a much smaller sample could presumably be representative. Inhomogeneity in gold content, whether because of the presence of detrital gold grains, gold precipitated on sulphides or other factors, probably accounts for the very different gold values for two of the duplicates.

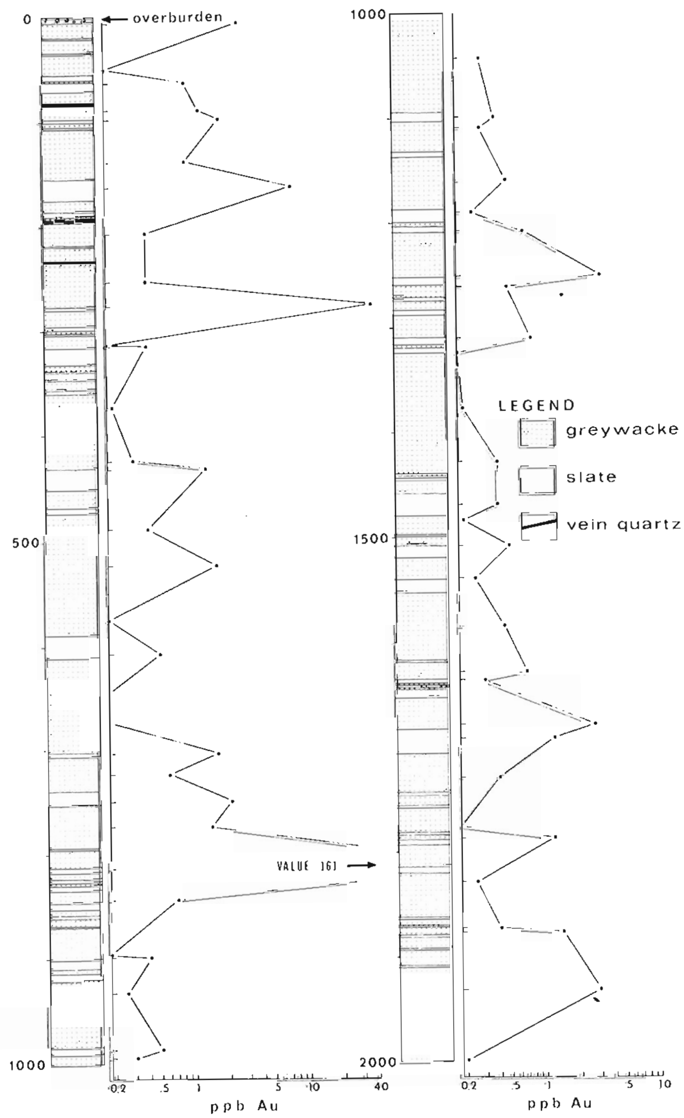

Figure 66. 2. Lithologic log of the 2000 foot (609. 6 m) Oldham drillhole with a line profile of the determined gold contents. Slate beds greater than 1 foot (30. 5 cm) thick are shown. Gold values below the detection limit are plotted at half the maximum value, except that values reported as <0. 2 and <0. 3 ppb are plotted on the baseline (0. 18 ppb Au).

Figure 66. 2 presents the gold values for the Oldham drill core in relation to a graphic log of the hole. This plot suggests that slate is enriched in gold relative to greywacke, and that there may be a cyclic variation of gold content in the hole. In particular, it appears that gold content is higher in those sections of the hole where interbedding of slate and greywacke is most frequent. It seems probable that the elevated gold values in these intimately interbedded sequences are to be attributed to a somewhat different depositional environment, possibly a shift to more reducing conditions which would favour deposition of sulphides and possibly other chemical precipitates.

A logarithmic probability plot of all the data for the Oldham hole has at least three inflection points, which could indicate four or more populations of gold values within the complete data set (Sinclair, 1974). This means that at least four factors have controlled gold distribution in these sediments, or possibly that a single concentrating mechanism has operated with varying intensity in a cyclic manner. Figure 66. 3 is a logarithmic probability plot showing the results of subdividing the data into individual slate and greywacke populations. These plots are simpler in form than that for the combined data, suggesting that different factors are responsible for concentration of a significant part of the gold in the two sediment types.

Truncation of the data set below about 0. 4 ppb Au, the approximate detection limit, means that the logarithmic probability curves are not reliable below 40 cumulative per cent. Likewise the portions of the curves above 98 cumulative per cent are not to be seriously considered as these are controlled by a very small number of samples, although in the case of the Oldham greywacke three samples with very similar gold contents are involved. A further consideration is that the number of samples analyzed, especially in the cases of Oldham slate and the Mount Uniacke samples, are undoubtedly not sufficient to define the logarithmic probability curves in detail.

Gold contents for the Oldham greywacke appear to represent the gradational mixing of two populations of values. Because data on the gold content of rock-forming minerals indicate that only quartz and feldspar are consistently very low (Crocket, 1974), the extremely low gold values of the greywacke are probably due to the coarse detrital component (i. e. quartz and feldspar) in these rocks. The second factor accounting for the total gold content of the greywackes is perhaps their finer, probably argillaceous and more gold-rich matrix component.

The Selati Formation of the northeastern Transvaal, Africa, is an Early Proterozoic (Aphebian) formation about 800 m thick that consists of a basal section (130 m), predominantly of fine grained argillaceous quartzite, overlain by shales and mudstones. Gold values for samples from the Selati formations and some overlying beds have been presented by Minnitt *et al.* (1973). The logarithmic probability curve for these gold values, presented in Figure 66. 3 for comparative purposes, is of interest because it exhibits a very sharp break completely at variance with what one might expect (Sinclair, 1974) from the superposition or parallel operation of different mechanisms of gold concentration. This type of curve suggests that the highest gold contents in these rocks are due to a process operating, not in addition to, but mutually exclusive of, the processes responsible for gold in the remainder of the samples. The logarithmic probability curve for the Oldham slate may contain a similar, although less well defined, break. The factor accounting for the higher gold values in these populations cannot be definitely identified but in the case of the Oldham slate, at least, could be the sulphide (pyrite and arsenopyrite) content of the rock. Unfortunately arsenic and sulphur analyses have not yet been carried out on these samples. The remainder of the logarithmic probability curve for the Oldham slate can probably be attributed to the coarse detrital and argillaceous factors tentatively identified for the greywackes, although the gold values for the slates are higher than for the greywackes.

In the greywacke population about 36 per cent of the values are below the detection limit, which varies from 0. 2 to 0. 8 ppb Au. When samples below the detection limit are arbitrarily taken as half the specified upper limit (e. g., values quoted as <0. 5 ppb Au are taken to be 0. 25 ppb) the arithmetic mean obtained is 0. 66 ppb Au. When the two highest values (2. 9 and

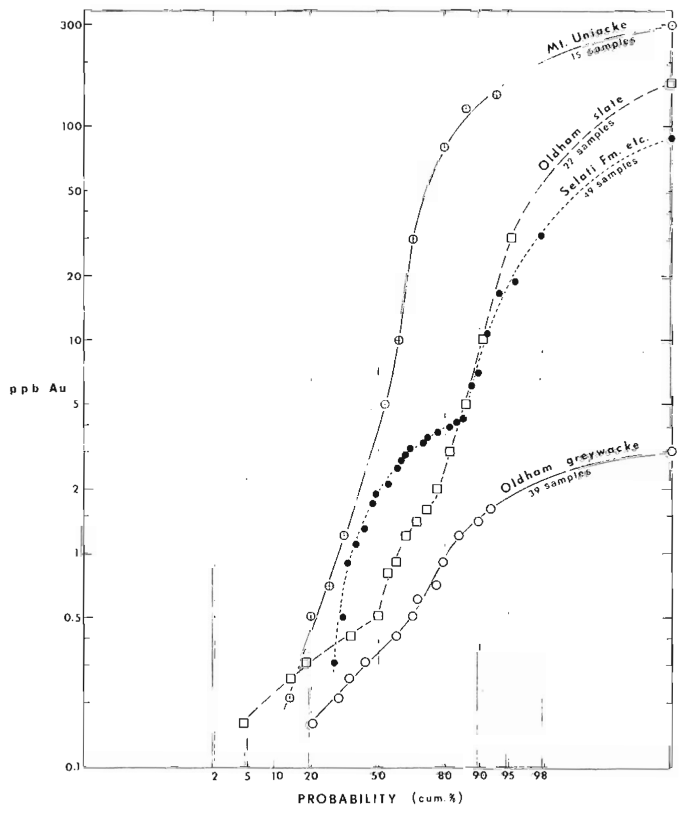

Figure 66. 3. Logarithmic probability plot of the gold values for the Oldham and Mount Uniacke samples. Gold values below the detection limit were arbitrarily assumed to be half the specified maximum value.

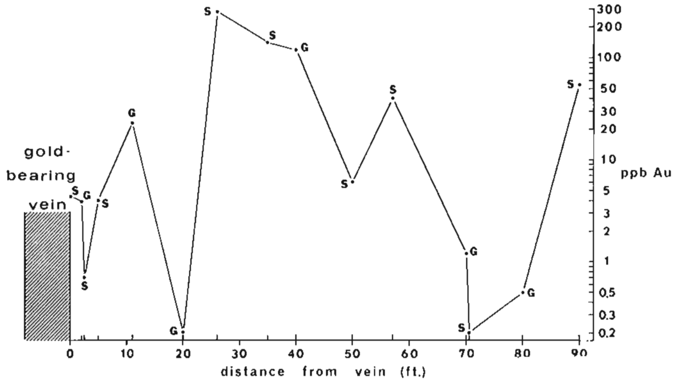

Figure 66. 4. Gold values for slate (S) and greywacke (G) from the Mount Uniacke trench shown in relation to distance from the gold-bearing quartz vein.

2. 7 ppb) are eliminated the arithmetic mean becomes 0. 54 ppb Au. The best arithmetic mean for the greywacke probably lies between these values. When plotted as a histogram the data show a positive skewness that may reflect an underlying lognormal distribution.

In the case of data for the Oldham slate only 18 per cent of the gold values lie below the detection limit. An approximate arithmetic mean, with the values below the detection limit assumed to be half the limiting maximum value, is 10. 1 ppb Au. When the two highest values (161 and 37 ppb Au) are eliminated the calculated arithmetic mean becomes 1. 23 ppb, which suggests that the two highest values unduly influence the former figure and that it is not a good measure of the geochemical abundance of gold in the slate. Specifically, the highest gold value (sample 318, Table 66. 2) can clearly be attributed to the presence of arsenopyrite in a carbonate vein (i. e. , gold has been introduced into this specimen). These data, however, clearly indicate that the slate is considerably enriched in gold relative to the greywacke, and that it in fact contains at least twice as much.

To obtain the best mean or geochemical abundance figures for the gold contents of the greywacke and slate, that is for data sets that are distributed lognormally, or approximately so, and that are truncated or censored (lowest values lie below the detection limit of the analytical method employed), special statistical techniques can be employed (Miesch, 1967; Sichel, 1966). When these techniques were applied in the present study the detection limit, which is variable probably as a result of the specific conditions of irradiation of each batch of samples, was arbitrarily taken to be 0. 3 ppb Au. The data were first log transformed. The maximum likelihood values for the means, that is values that closely approximate geometric means, were then obtained using the methods of Cohen (1959, 1961) and Sichel (1952), as reviewed and presented by Miesch (1967). The confidence intervals about the means were obtained using the equations and tables of Sichel (1966). The maximum likelihood value for the mean gold content of greywacke is 0. 62 ppb Au, and the true mean lies between 0. 50 and 0. 87 ppb Au at the 90 per cent confidence level. For the slate, using all the data, the best mean is 4. 6 ppb Au, and the true mean lies between 2. 3 and 16. 8 ppb Au at the 90 per cent confidence level. However, when the highest gold value for the slate is eliminated, the best mean for the remaining 21 samples is 2. 0 ppb Au and the true mean lies between 1. 2 and 5. 0 ppb Au at the 90 per cent confidence limit. Because of the great influence of the one very high value and the fact that the logarithmic probability curve suggests the presence of two distinct populations of gold values in the slate, these figures for mean gold content of the Oldham slate are not considered particularly reliable.

Gold values for slate and greywacke from the trench at Mount Uniacke are presented in Table 66. 3 and in Figures 66. 3 and 66. 4. The sedimentary rocks at this location, adjacent to a productive gold-bearing vein, are obviously enriched relative to the slate and greywacke in the Oldham drillhole. A simple arithmetic mean for the trench samples is 45. 3 ppb Au. A mean of 69. 5 ppb Au is obtained when a maximum likelihood estimation technique (Sichel, 1966) is used, although the 90 per cent confidence limits (24. 3 to 1650 ppb Au) are extremely broad. The profile plotted in Figure 66. 4, when a few low values are ignored (especially that for a greywacke sample at 20 feet (6 m) from the vein), suggests a rise to a maximum value at about 25 feet (7. 6 m) from the vein followed by a decline at greater distance. However, the final sample at a distance of 90 feet (27. 4 m) is not in agreement with this trend, and in view of the limited number of samples the trend should probably be considered a fortuitous result. The pattern of gold contents in the country rock is not considered indicative of a particular genetic process, although introduction of the gold during mineralization, rather than lateral secretion from the immediate host rocks, would seem to be favoured. Fakhry (1974) found the gold contents (0. 01 to 8 ppm) of altered granodiorite country rock adjacent to the Fawakhir gold vein, Egypt, to be closely correlated with the carbonate contents of the rock. However, although possibly present, carbonatization of the country rock at Mount Uniacke has not been recognized.

Discussion of Results

Neutron activation analyses for gold in clastic rocks are relatively limited in number, and other analytical techniques such as atomic absorption give a poor idea of average gold content and of the overall distribution of values due to truncation at the lower values because detection limits are too high (generally about 20 ppb for the atomic absorption method). For example, Seeland (1973) in a study concerned primarily with clastic sediments in the Great Lakes region, United States, found that only 36 of 777 samples had a sufficient gold content to be measurable by an atomic absorption method. Likewise, of 77 samples of Eocene sandstone and clays from the Ione Formation of California, gold was detected by atomic absorption analysis in less than one-third (Wollenberg and Dodge, 1973). Some of the published neutron activation determinations of gold in clastic

Table 66.4

Neutron activation analyses of gold in clastic rocks

Reference	Locality	Number of samples	Rock type	Gold (ppb) Average	Range
Jones (1969, p. 22)	nine localities	24	Sandstone	7.5	0.6-41
DeGrazia and Haskin (1964, p. 562)	five separate localities	5	Sandstone, quartzite	6	2.3-11.6
Crocket *et al.* (1973)	Canadian Arctic	9	Recent nearshore clay silt	2.7	1.6- 4.2
Crocket *et al.* (1973)	East Pacific Rise	14	Recent deep sea clayrich sediments	3.1	1.3- 7.6
Crocket *et al.* (1973)	Indian, Antarctic and Pacific Oceans	7	Recent deep sea lutite etc.	2.0	0.6- 4.2
Baedecker (1967)	Wisconsin, Kentucky, Ontario, Australia	5	Sandstone	3.6	0.3-12
Shcherbakov (1967)	Altai and Chukotka, U.S.S.R.	27	Sandstone	2.1	0.6- 6.2
Shcherbakov (1967)	Altai, U.S.S.R.	2	Siltstone	5.9	5.5- 6.2
Shcherbakov (1967)	Kuznetsk Ala-Tau and Chukotka, U.S.S.R.	23	Shale	2.3	0.66- 8.6
Moiseenko *et al.* (1971)	Primurya region, U.S.S.R.	14	Sandstone and silty argillite	1.7	0.8-4
DeGrazia and Haskin (1964)	Kansas and South Dakota, U.S.A.	2	Shale	6.0	4.7- 7.2
Minnitt *et al.* (1973)	Northeastern Transvaal, Africa	49	Shales, mudstones, argillaceous quartzites	2.5	0 -31.2

rocks are summarized in Table 66.4. These data have largely been reproduced from a compilation by Crocket (1974).

The analyses of slate and greywacke from the Oldham drillhole have indicated a very low gold content in these rocks, especially the greywacke, although the data are generally comparable to the range of values and the lower averages presented in Table 66.4. In particular, data by Crocket *et al.* (1973) for deep sea lutite, by Shcherbakov (1967) for sandstone and shale, by Moiseenko *et al.* (1971) for sandstone and silty argillite, and by Minnitt *et al.* (1973) for shales, mudstones and argillaceous quartzites are comparable to the present data set.

The fact that the Oldham slate is significantly richer in gold than associated greywacke is interesting in view of the finding of Minnitt *et al.* (1973) that in sedimentary rocks of the Transvaal Supergroup gold was enriched in fine argillaceous quartzites, by a factor of around 10, in comparison with mudstones and shales. In the case of the Oldham sedimentary rocks, gold, in addition to that contained in the detrital silicate components, may have been contributed by absorption or chemical precipitation following transport in solution, whereas much of the gold in the argillaceous quartzites of the Transvaal Supergroup may be detrital native metal.

The highest gold contents of Oldham slate have been tentatively attributed to precipitation of gold with or on pyrite and arsenopyrite in the rock. However, there is no strong evidence that this is the case and the gold could be related to other chemically precipitated components in the rock, absorbed by carbonaceous material, or be present as a fine detrital constituent. However, the very low gold content of the Oldham greywacke can reasonably be attributed entirely to its detrital silicate constituents, so that the presence of fine detrital gold in either the greywacke or slate is considered unlikely.

A mean gold content of 0.62 ppb Au (between 0.50 and 0.87 ppb Au at the 90 per cent confidence level) indicated by maximum likelihood calculations for greywacke in the Oldham drillhole is probably reliable. This is in spite of the fact that 14 of 39 samples had gold contents below the detection limit. Because of a smaller number of samples, greater variability of values, and an indication of more than one population of values (bimodality) the calculated mean of 2.0 ppb Au (between 1.2 and 5.0 ppb Au at the 90 per cent confidence level) for the Oldham slate is much less certain. The higher gold values in the Mount Uniacke rocks are analogous to the gold contents found by Minnitt *et al.* (1973) in the vicinity of the old gold workings at The

Crags (Transvaal, Africa), values that were about nine times as great as in equivalent sedimentary rocks well away from productive areas.

On the basis of neutron activation analyses reported here for gold in slate and greywacke of the Meguma Group in Nova Scotia the following suggestions can be made:

1. In evaluation of a lateral secretion hypothesis for the gold-bearing veins, slate must be considered a better potential source of gold than greywacke. However, in the case of the Mount Uniacke veins this study does not support a process of lateral secretion from the adjacent country rock.

2. Areas in which slate and greywacke are most intimately interbedded may have the greatest potential.

3. Areas yielding results comparable to the gold values obtained for samples from the Mount Uniacke trench may be in the vicinity of significant gold-bearing veins. The possible application of this empirical guide is not dependent on the genetic process by which the veins have formed.

References

Baedecker, P.A.
1967: The distribution of gold and iridium in meteoritic and terrestrial materials; Univ. Kentucky, Ph.D. dissertation, 110 p.

Boyle, R.W.
1961: The geology, geochemistry and origin of the gold deposits of the Yellowknife district; Geol. Surv. Can., Mem. 310, 193 p.

Cohen, A.C., Jr.
1959: Simplified estimators for the normal distribution when samples are singly censored or truncated; Technometrics, v. 1, no. 3, p. 217-237.

1961: Tables for maximum likelihood estimates; singly truncated and singly censored samples; Technometrics, v. 3, no. 4, p. 535-541.

Crocket, J.H.
1974: Gold; Section 79, Handbook of Geochemistry, v. 11/4, Pt. B-O, 79 p.

Crocket, J.H., Macdougall, J.D., and Harris, R.C.
1973: Gold, palladium and iridium in marine sediments; Geochim. Cosmochim. Acta, v. 37, p. 2547-2556.

DeGrazia, A.R. and Haskin, L.
1964: On the gold content of rocks; Geochim. Cosmochim. Acta, v. 28, p. 559-564.

Fakhry, A.A.
1974: Geochemical studies at the Fawakhir gold mine, Eastern Desert, Egypt, 2. Infrared absorption spectroscopy as a means of characterizing the primary dispersion halos of gold deposits; J. Geochem. Explor., v. 3, no. 3, p. 265-275.

Jones, R.S.
1969: Gold in igneous, sedimentary and metamorphic rocks; U.S. Geol. Surv. Circ., no. 610, 27 p.

Miesch, A.T.
1967: Methods of computation for estimating geochemical abundance; U.S. Geol. Surv., Prof. Paper 574-B, 15 p.

Minnitt, R.C.A., Button, A., and Kable, E.J.D.
1973: The gold content of pre-Malmani argillaceous sediments in the Transvaal Supergroup, northeastern Transvaal; Economic Geology Research Unit, Univ. of the Witwatersrand, Johannesburg, Infor. Circ. 82, 11 p.

Moiseenko, V.G., Shcheka, S.H., Fatlyanov, I.I., and Ivanov, V.S.
1971: Geokhimicheskie osobennosti rapredeleniya zolota v porodakh Tikhookeanskogo poyasa (Geochemical aspects of the distribution of gold in rocks of the Pacific Ocean belt); Moscow, Nauka.

Schenk, P.E.
1970: Regional variation of the flysch-like Meguma Group (Lower Paleozoic) of Nova Scotia, compared to recent sedimentation off the Scotian Shelf; p. 127-153 *in* Flysch sedimentology in North America: Geol. Assoc. Can., Spec. Paper No. 7.

Seeland, D.A.
1973: A geochemical reconnaissance for gold in the sedimentary rocks of the Great Lakes region, Minnesota to New York; U.S. Geol. Surv. Bull. 1305, 16 p.

Shcherbakov, Yu. G.
1967: Raspredeleniye i usloviya kontsentratsii zolota v rudnykh provintsiyakh (Distribution and concentration conditions of gold in ore provinces); Moscow, Nauka.

Sichel, H.S.
1952: New methods in the statistical evaluation of mine sampling data; Inst. Min. Met. Trans., v. 61, p. 261-288.

1966: The estimation of means and associated confidence limits for small samples from lognormal populations; p. 106-123 *in* Symposium on mathematical statistics and computer applications in ore valuation; Afr. Inst. Min. Met., Johannesburg.

Sinclair, A.J.
1974: Selection of threshold values in geochemical data using probability graphs; J. Geochem. Explor., v. 3, no. 2, p. 129-149.

Taylor, F.C. and Schiller, E.A.
1966: Metamorphism of the Meguma Group of Nova Scotia; Can. J. Earth Sci., v. 3, no. 7, p. 959-974.

Wollenberg, H.A. and Dodge, F.C.W.
1973: Radioelement and trace-element content of the Ione Formation, Central California; U.S. Geol. Surv. Bull. 1382-B, 17 p.

Project 750060

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Paleozoic massive sulphide deposits in the Appalachian orogen

Attention was concentrated on stratabound, massive sulphide deposits of the volcanogenic type in the Appalachian orogen of New Brunswick and Quebec over a five-week period in July and August. The overwhelming majority of the deposits visited are in New Brunswick, with all but one in the Newcastle-Bathurst mining district.

The host rocks to the New Brunswick deposits are (?) middle Ordovician volcanic and sedimentary rocks of the Tetagouche Group (Skinner, 1956). Special attention was paid to the following aspects of the geology of the ores and their host rocks:

A) The lithological-stratigraphical location of the ore deposits within the Tetagouche Group;

B) The metal and mineral associations present in the sulphide deposits;

C) The metamorphic effects exhibited by the ores.

The following lithological units are widely recognized in the Tetagouche Group rocks (after Skinner, 1956, 1974):

Felsic metavolcanic rocks: (tuffs, flows, quartz-sericite schist, biotite-chlorite schist; minor slate and mafic volcanic rocks).

Augen schist: Quartz and feldspar-phyric schist (crystal tuffs, ashflow tuffs, ash fall tuffs), quartz-sericite schist, quartz-chlorite (biotite) schist; minor mafic metavolcanic rocks and metasedimentary rocks.

Metasedimentary rocks: a) phyllitic slate, argillite, greywacke, siliceous argillite. b) feldspathic schist, minor slate.

Mafic metavolcanic rocks: a) spilite, greenstone, minor trachyte and metasedimentary rocks. b) greenstone, mafic metatuffaceous rocks, red and grey slate and phyllite.

A survey of 36 individual ore occurrences in the Bathurst Newcastle area shows that 19 are in metasedimentary rocks, mainly in slate, phyllite or argillite; 14 are within rocks mainly described as metarhyolites, a term which includes quartz sericite schist, while 3 deposits occur at the contact between metasedimentary and felsic metavolcanic rocks.

The majority of the deposits occurring in a metasedimentary lithology, showed rocks described as metarhyolites or "augen schist" (crystal tuffs) either above or below the immediate wall rocks. In only five cases were metabasaltic (mafic) rocks reported in contact with the immediate wall rocks. The largest body of sulphides so far known in the Bathurst-Newcastle mining district, the Brunswick Mining and Smelting No. 12 deposit, is in a sequence of alternating metasedimentary and metavolcanic formation, the ore being at the contact between two metasedimentary units (Luff, 1975).

Metabasaltic rocks were reported near about half the deposits occurring in a felsic metavolcanic lithology. However, in the majority of the deposits, the felsic metavolcanics are closely associated with a dominantly metasedimentary lithology. It may not be without significance that the very important Brunswick Mining and Smelting No. 6 deposit appears to be in a totally metavolcanic lithology.

This preliminary review of the lithological and stratigraphical relations of the Bathurst-Newcastle stratabound sulphide mineralization has not so far pointed up any systematic or preferential location for the occurrence of orebodies. The general association: metasediments and/or felsic metavolcanics, already well-known, is predictably the predominant one. The relative paucity of metavolcanics of mafic or intermediate character near the deposits is also re-emphasized. Such associations are in contrast with those pertaining in other parts of the fragmented Lower Paleozoic fold belt of the North Atlantic Region, as for example the central and northern Norwegian Caledonides, where mafic volcanics are much more in evidence.

The relative metal abundances of the Bathurst-Newcastle stratabound sulphide ores, both in numbers of occurrences and contained tonnages, based on weight percentages, is zinc, lesser lead, lesser or minor copper. In about 15 per cent of the occurrences, mostly minor in tonnage, zinc remains the dominant base metal, but lead and copper change order of importance or are present in almost equal amounts. In 20 per cent of the occurrences (again mostly small bodies of ore) copper predominates, zinc and lead being of minor importance. Examination of available assay data shows that in many cases single sulphide deposits may be composed of units exhibiting two or even all three of the above metal associations. Commonly this fact is disguised in the grade figures reported for the whole deposit, and it is not always possible to distinguish between the different component units.

The more prominent deposits with more than one metal association are Brunswick Nos. 6 and 12, Heath Steele, Chester Mines and Half Mile Lake.

Consideration of the iron sulphide mineral associated with the different base metal associations noted above revals interesting correlations. The most obvious of these is the division into dominantly pyrite-bearing and dominantly pyrrhotite-bearing ore types. Among other things it is clear that the predominantly zinc- and lead-bearing ores carry pyrite as the sole or major iron sulphide, and the pyrrhotite-rich sulphide bodies are commonly characterized by a high chalcopyrite content and few or no lead and zinc sulphides. (However, some, mostly minor, bodies of pyrite-dominated copper ores occur and not all pyrrhotite-dominated sulphide bodies are necessarily base metal bearing. In addition, there are bodies of so-called "barren" pyrite carrying uneconomic Zn, Pb, and Cu values.)

Stanton (1959) showed some years ago that this division into pyritic and pyrrhotitic ore types is generally a feature of stratabound massive sulphide ores from Paleozoic fold mountain belts in eastern North America (New Brunswick especially), the Eastern Highland Belt of Australia and the Avoca district of Ireland. Following this, Vokes (1962, 1963) discussed the features of the two ore types in the Caledonian fold mountain belt of Scandinavia.

In New Brunswick the writer observed good examples of the juxtaposition of pyritic and pyrrhotitic ore types at the Chester Mines deposit, the Heath Steele ACD orebody and to a lesser extent in the area of Key Anacon Mine.

At Chester Mines, a near-surface body of massive, fine grained pyritic Zn-Pb(Cu) ore measuring 410 feet by 560 feet by 40 feet thick, has a "tail" of banded, disseminated pyrite-pyrrhotite-chalcopyrite mineralization, with a width of up to 400 feet and a total thickness of up to 200 feet plunging at 20° in a northwesterly direction for approximately 5000 feet.

A section of the banded "tail" at No. 1 pump station in the spiral incline at Chester Mine shows coarse grained pyrrhotite-chalcopyrite ore with a typical deformed, roughly foliated ("durchbewegt") texture. Contorted inclusions of the chloritic country rock, here and there in the form of detached fold cores, occur in the sulphide matrix. The ore band is some 5 feet in thickness and occurs in a monoclinal flexure zone, or perhaps a fault zone, which dragged down to the west an otherwise concordant band of sulphides. The deformational texture observed in the ore seems to be the result of movement within the flexure or fault zone.

Copper-rich chalcopyrite-pyrrhotite ore in the footwall zone of the Heath Steele B orebody is well-documented, *inter alia,* by Gates (1972). The present writer had the opportunity of examining a section through the rather more deformed ADC body at this mine (8400 level, cross-cut N through orebody). The metal (ore-type) zoning is not as distinct here as in the B orebody and the interrelation of the pyritic and pyrrhotitic ore types is rather complex. Typical "durchbewegt" or intensely deformed chalcopyrite-pyrrhotite ore can be seen in spatial association with highly folded, interbanded pyritic ore and chlorite schist in much the same manner as at Chester Mines. Elsewhere, a chalcopyrite-poor, pyrrhotite-rich ore type occurs as dyke-like or other cross-cutting features intersecting ore of the pyritic type and enclosing innumerable rounded to angular fragments of the latter. The fragments range in size from a few millimetres to perhaps one decimetre.

On the west bank of the Nepisiguit River, just upstream from the bridge on the Allardville-Bathurst Mines Road, is a small sulphide prospect which prompted investigations of the area that led to the discovery of the Key Anacon orebody. The spoil heap from the main pit of this prospect is typical chalcopyrite-pyrrhotite ore. The sulphide zone, traced southwards towards the river changes character to disseminated and massive pyritic ore, lying concordantly in the quartz-sericite schist or felsic metavolcanics, which also host the Key Anacon ore. Because of poor exposure, the exact relationship between the two ore types at this locality is uncertain.

The occurrence of chalcopyrite-bearing pyrrhotitic ore has also been noted in more recent literature on the New Brunswick ores. Rutledge (1972, p. 65) mentions a "tabular keel of pyrrhotite-pyrite-chalcopyrite (which) extends down-plunge from the thick footwall pyrite (zone) into an envelope of 'footwall rocks' at the Brunswick No. 6 deposit." The same authority (Rutledge, 1972) also mentions that the thick pyrite mass in the deeper levels of Brunswick No. 12 mine encloses a pyrrhotite zone near the footwall side of the deposit. However, "chalcopyrite is relatively insignificant in the No. 12 pyrrhotite zone". Rutledge also notes that the footwall schists in the vicinity of the pyrrhotite zones are strongly chloritized and heavily veined with pyrite, pyrrhotite and variable amounts of chalcopyrite.

Luff (1975) recently interpreted the pyrrhotitic zone at the Brunswick No. 12 orebody as a "vent through which the ore elements ascended to the surface to be chemically precipitated above the footwall rocks in an irregular shallow basin" (*see* Luff's Fig. 10).

The present writer (Vokes, 1962, 1963) previously discussed possible genetic relationships of the pyritic and pyrrhotitic ore types without coming to any definite conclusions. It would seem that the following facts are relevant to such a discussion and must be explained in any genetical hypothesis which may be adopted.

1) The pyrrhotitic ore type can occur either as individual, independent orebodies of the generally stratabound massive sulphide type or as constituent parts of orebodies of the same general type having a dominant pyritic composition.

2) In orebodies of the composite types, the pyrrhotite-rich components can occur as either:

- a) cross-cutting, commonly dyke-like features, in many places showing a brecciated texture, the clasts apparently consisting of the same pyritic type ore which is being cross-cut (e. g. Heath Steele ACD; Brunswick ores; Joma, Norway) or
- b) apparently conformable portions of composite pyritic-pyrrhotitic orebodies showing no evidence of cross-cutting or brecciation. Commonly in such cases the pyrrhotitic portions (generally quite copper-rich) are at or near the footwall contacts of the composite orebodies. Chalcopyrite and pyrrhotite may or may not be present as 'stringer' or disseminated ore in the footwall rocks adjacent to the massive bodies, e. g. Heath Steele B; Bleikvassli, Norway.

3) The pyrrhotitic-type orebodies, especially those of the chalcopyrite-rich type, are almost invariably highly deformed internally, showing fabrics of the "Durchbewegung-type" (Vokes 1963, 1968, 1969, 1973) which indicate deformation occurred subsequent to the deposition of the sulphides and involved both these, their immediate wall rocks and interbanded silicate layers.

Taking the above-listed features of the pyrrhotitic ore types into consideration, it would seem that they may have formed by one or both of two main processes.

A) Primary processes. These have led to the deposition of primary pyrrhotite-dominated sulphide bodies, presumably by "ore forming solutions" deficient in sulphur with respect to iron compared with those depositing the heavily pyritic, stratiform, layered ores.

The cross-cutting and brecciated pyrrhotitic bodies ((2a) above) are patently epigenetic with respect to the pyritic body which appears to have been consolidated before the pyrrhotite body was introduced. This makes it difficult to envisage the pyrrhotitic portions of such composite ores as being related to feeder vents through which the components of the layered or banded pyritic ores ascended to the lava surface as suggested by Luff (1975) for the Brunswick No. 12 ore. Rather they would seem to represent some form of later mineralizing activity. This may have been related to the same general volcanic episode which was responsible for the initial, pyritic mineralization (cf. Sangster, 1972, p. 27).

In those cases ((2b) above) where the pyrrhotitic mineralization forms apparently concordant, commonly footwall, portions of composite pyritic-pyrrhotitic ore bodies, the primary order of deposition is uncertain. It seems reasonable to assume that the footwall portions of volcanogenic (exhalative-sedimentary) orebodies were deposited before their hanging wall portions. In this case one is probably seeing the effects of deposition by "ore forming solutions" in which the sulphur to iron ratio increased with time. Indeed it is possible, as at the Bleikvassli Mine, northern Norway (Vokes, 1963), to document increase of sulphur towards the hanging wall even within the pyritic ore zones. At Grube Bayerland, Bavaria (Maucher, 1939; Spross, 1954) the apparent stratigraphic succession is: magnetite-bearing ore, pyrrhotite ore, pyrite ore, in a deposit of the type here under consideration, again indicating an increase in S/Fe ratio from the footwall to hanging wall. Orebodies consisting of solely pyrrhotitic type ore are also explained on this primary depositional hypothesis as being due to solutions deficient in sulphur with respect to iron, compared with pyritic type orebodies in the same district. In these instances the sequence of deposition of the two ore types is mostly impossible to determine.

B) Metamorphic processes. The prevalence of fabrics showing unequivocal evidence of deformation (Durchbewegung) in ores of the pyrrhotitic type makes it necessary to consider the role of deformation in particular, and metamorphism in general, in the formation of this type of sulphide ore. The need to consider the effects of metamorphism is emphasized by the fact that pyrrhotite is increasingly more abundant in all ores of the general type under consideration here, as the grade of metamorphism of their enclosing rocks increases. In other words pyrrhotite is commonly itself a metamorphic mineral, being formed by the breakdown of pyrite, mainly by the increasing metamorphic temperatures. At temperatures below 743°C (Kullerud and Yoder, 1959) the transformation of pyrite to pyrrhotite involves loss of sulphur, either by escape from the system or by reaction (sulphurization) with iron-bearing silicate and other rock-forming minerals to form new iron monosulphide.

The appearance of pyrrhotite in regionally metamorphosed rocks at about or just below the biotite isograd has been documented, *inter alia*, by Antun (1954) and Carpenter (1974). Most of the massive sulphide ores showing wholly pyrrhotitic paragenesis certainly appear to be in rocks of upper greenschist and higher, metamorphic grades. According to Skinner (1974) the metamorphism of the Tetagouche rocks in the Bathurst-Newcastle area has been in the greenschist facies; only in restricted areas have minor amounts of biotite made their appearance. Later work by Helmstaedt (1973, Fig. 1) has enabled the biotite isograd to be delineated in the central Bathurst-Newcastle area. Helmsteadt finds that "on the regional scale pyrrhotite-bearing assemblages are restricted to sulphide deposits in biotite-bearing country rocks" (p. 40). This statement agrees well with the present writer's own observations in the area this summer. The Brunswick Mining, Heath Steele and Chester Mines deposits all lie within the area of biotite-bearing rocks delineated by Helmstaedt. The Key Anacon area lies outside the biotite isograd but within the contact aureole of the Bathurst granite of Devonian age (Helmstaedt, 1973, p. 40).

It is thus not possible to ignore the influence of metamorphism on the occurrence of the pyrrhotitic ore types even in such relatively low metamorphic terrane as that of the Newcastle-Bathurst area.

The evidence available to date would suggest that the pyrrhotite-rich ore types may be the result of the breakdown of pyrite as a result of elevated metamorphic temperatures, accompanied by deformation and mobilization, during the Taconic orogeny. The deformation and mobilization have, in this view, been responsible for the epigenetic relationships shown in many places by the pyrrhotitic ores towards the pyritic ores.

One is, however, seemingly able to trace the influence of possible primary depositional differences through the "metamorphic veil". For example the commonly copper-rich nature of the pyrrhotitic ores as compared with the pyritic ores in the same orebody or same area is difficult to explain on purely metamorphic grounds. The metal differences in this case could be due to orebody zoning – for example, that one may be dealing with a metamorphosed feeder or stringer zone, or the copper-rich footwall portion of a composite massive ore.

References

Antun, P.
1954: Sur les paramorphoses de cubes de pyrite et le mode de formation metamorphique de quelques sulfures de l'Ardenne; Ann. Soc. Geol. Belg. Bull., v. 77, p. 343-358.

Carpenter, R.H.
1974: Pyrrhotite isograd in Southeastern Tennessee and Southwestern North Carolina; Geol. Soc. Am. Bull., v. 85, p. 451-456.

Gates, W.
1972: Heath Steele Mine; *In* Mineral deposits of southern Quebec and New Brunswick; (McAllister, A.L. and Lamarche, R.Y., eds.) XXIV Int. Geol. Congr. Montreal, 1972. Guidebook Excursion A58-C58, p. 68-72.

Helmstaedt, H.
1973: Structural geology of the Bathurst-Newcastle district; *In* Geology of New Brunswick, N.E.I.G.C., 1973 Field Guide to excursions, p. 34-46.

Kullerud, G. and Yoder, H.S., Jr.
1959: Pyrite stability relations in the Fe-S system; Econ. Geol., v. 54, p. 533-572.

Luff, W.M.
1975: Structural Geology of the Brunswick No. 12 open pit; Can. Inst. Min. Met. Bull., v. 68, no. 758, p. 64-74.

Maucher, A.
1939: Über die kieslagerstätte der Grube "Bayerland" bei Waldsassen in der Oberpfalz; Zeitsch. f. angew Min., v. 2, p. 219-275.

Rutledge, D.W.
1972: Brunswick Mining and Smelting Corporation Nos. 6 and 12 mines; *In* Mineral Deposits of southern Quebec and New Brunswick (McAllister, A.L. and Lamarche, R.Y., eds.) XXIV Int. Geol. Congr. Montreal, 1972. Guidebook Excursion A58-C58, p. 58-67.

Sangster, D.F.
1972: Precambrian Volcanogenic massive sulphide deposits in Canada: a review; Geol. Surv. Can., Paper 72-22, 44 p.

Skinner, R.
1956: Geology of the Tetagouche Group, Bathurst, New Brunswick; McGill Univ. Ph.D. Thesis (unpubl.).

1974: Geology of the Tetagouche Lakes, Bathurst and Nepisiguit Falls map-areas, New Brunswick, with emphasis on the Tetagouche Group; Geol. Surv. Can., Mem. 371, 133 p.

Spross, W.
1954: Gefügeanalyse and tektonische Geschichte der Grube "Bayerland" im Oberpfalzer Wald; Neues Jb. Mineral. Abh., v. 86, no. 3, p. 393-436.

Stanton, R.L.
1959: General features of the conformable "pyritic" ore bodies. Part II – Mineralogy; Can. Inst. Min. Met. Bull., v. 52, no. 570, p. 631-643.

Vokes, F.M.
1962: Mineral parageneses of the massive pyritic orebodies of the Caledonides of Norway; Econ. Geol., v. 57, p. 890-903.

1963: Geological studies on the Caledonian pyritic zinc-lead orebody at Bleikvassli, Nordland, Norway; Norges geol. unders., No. 222, 126 p.

1968: Regional metamorphism of the Palaeozoic geosynclinal sulphide ore deposits of Norway; Inst. Min. Met., Trans. Sect. B, v. 77, p. 853-859.

1969: A review of the metamorphism of sulphide deposits; Earth Sci. Rev., v. 5, p. 99-143.

1973: "Ball texture" in sulphide ores; Geol. Fören. i Stockholms Förh., v. 95, p. 403-406.

SCHEELITE MINERALIZATION OF SOUTHEASTERN NOVA SCOTIA

E.M.R. Research Agreement 1135-D13-4-181/75

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Two types of tungsten deposits are known to occur in southeastern Nova Scotia:

(a) quartz veins with wolframite and cassiterite with some fluorite, scheelite, molybdenite, and Cu-Fe sulphides associated with aplites and pegmatites in the more felsic phases of the Devonian batholiths; and
(b) folded quartz-carbonate veins with scheelite, arsenopyrite, and rutile, subparallel to the bedding of the metapelites of the Ordovician Meguma Group.

Four deposits of the latter type have seen limited production of tungsten ore. The general view has been to ascribe the genesis of both types to solutions generated by the granitic intrusions (Malcolm, 1912, 1929), and their spatial distribution has been described as zoned with respect to the plutons (Newhouse, 1936). There is little reason to doubt the genetic relationship between the wolframite mineralization and the felsic igneous bodies (Marc Charest, pers. comm., 1975), but our work demonstrates that this relationship is not valid for the latter group of deposits.

Field relationships, structural, textural, and detailed mineralogical studies of tungsten mineralization, especially at the Moose River scheelite orebody (Miller, 1974), indicate that the scheelite in these deposits crystallized early in the paragenesis of the veins, which, in turn, crystallized early in the regional deformation of the Meguma Group. This regional deformation, the Acadian orogeny (Schenk, 1971) produced northeasterly trending folds and axial-plane slaty cleavage (Fyson, 1966). The crystallized veins were buckled or boudinaged depending upon their changing orientation with respect to these regional stress conditions and resulting strain. The granitic intrusions cross-cut these structures. They are thus younger than the scheelite-bearing veins, and cannot have a direct genetic relationship with them.

Newhouse (1936) considered the presence of tourmaline in the scheelite-bearing veins at Moose River and elsewhere to be indicative of the contact effects of a nearby unexposed intrusion and thus evidence for a magmatic provenance of the mineralizing solutions and further confirmation of a zonal distribution of mineralization. Electron microprobe analyses and X-ray diffraction studies of the host rock and vein minerals at Moose River (Miller, 1974) demonstrate that the "tourmaline" as described by Newhouse (1936) is in fact rutile which is a common metamorphic mineral in slates.

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Compositions of mineral phases of the Fe-As-S system, a useful geothermometer (Kretschmar, 1973), and fluid inclusion data (Graves, 1975) suggest that the veins crystallized under conditions of temperature and pressure compatible with greenschist regional metamorphic conditions (Taylor and Schiller, 1966) of the surrounding country rocks.

Scheelite was one of the first minerals to crystallize in the veins. Our model (Graves, 1975; Graves and Zentilli, in prep.) visualizes major vein minerals originating from the metamorphic pore fluids. In that the most likely source of metals in this model is in the sediments themselves, our conclusion does not differ significantly from those of Faribault (1899) and Boyle (1966) in explaining the source of heavy metals in gold-bearing quartz veins that occupy a similar setting in the Meguma Group.

The proximity of the intrusions had no effect on the size or grade of this type of scheelite mineralization in Nova Scotia.

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References

Boyle, R.W.
1966: Origin of gold and silver in the gold deposits of the Meguma Series, Nova Scotia; (abstract), Can. Mineral., v. 8, p. 662.

Faribault, E.R.
1899: On the gold measures of Nova Scotia and deep mining; Min. Soc. Nova Scotia, Halifax, Nova Scotia, 11 p.

Fyson, W.K.
1966: Structures in the lower Paleozoic Meguma Group, Nova Scotia; Geol. Soc. Am. Bull., v. 77, p. 931-943.

Graves, M.C.
1975: The formation of gold-bearing quartz veins in Nova Scotia: hydraulic fracturing under conditions of greenschist regional metamorphism during early stages of deformation; unpubl. M.Sc. thesis, Dalhousie Univ., Halifax, Nova Scotia.

Kretschmar, U.
1973: Phase relations involving arsenopyrite in the system Fe-As-S and their application; unpubl. Ph.D. thesis, Univ. Toronto, 146 p.

Malcolm, W.
1912: Gold Fields of Nova Scotia; Geol. Surv. Can., Mem. 20, 331 p.

1929: Gold Fields of Nova Scotia; Geol. Surv. Can., Mem. 156, 253 p.

Miller, C.K.
1974: Scheelite mineralization in the region of the Moose River gold district, Halifax County, Nova Scotia; unpubl. B.Sc. Honours thesis, Dalhousie Univ., Halifax, Nova Scotia.

Newhouse, W.H.
1936: A zonal gold mineralization in Nova Scotia; Econ. Geol., v. 31, p. 805-831.

Schenk, P.E.
1971: Southeastern Atlantic Canada, northwestern Africa, and continental drift; Can. J. Earth Sci., v. 8, p. 1218-1251.

Taylor, F.C. and Schiller, E.A.
1966: Metamorphism of the Meguma Group of Nova Scotia; Can. J. Earth Sci., v. 3, p. 959-974.

69. PETROLOGY, GEOCHEMISTRY AND ECONOMIC GEOLOGY OF THE NEW ROSS AREA, LUNENBURG COUNTY, NOVA SCOTIA

E. M. R. Research Agreement 1135-D13-4-181/75

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The New Ross-Vaughan adamellite complex is intrusive into the granodiorites of the South Mountain Batholith (10 000 km^2). A small area (≃ 100 km^2) centred on New Ross, Lunenburg County, contains most of the known occurrences of mineralization in the entire batholith. The complex consists mainly of 2-mica porphyritic adamellite in which late-stage minor bodies of alaskite, aplite and pegmatite are intruded (McKenzie and Clarke, 1975). All rocks in the area show the effects of hydrothermal alteration. The mineralization (Sn-W-Cu-Mo) is associated with late-stage aplites-pegmatites, vein systems and greisen.

The objects of this investigation are:

1. To understand the relationship between the rock types and textures and between the mineralized areas and host lithologies by field mapping and petrography.

2. To seek geochemical parameters associated with the deposits which might indicate other favourable areas to look for mineralization elsewhere in the batholith.

3. To catalogue and analyze the economic minerals of each of the deposits.

Progress on the first of these includes recognition of the occurrence of several deposits (Reeve's Tin, Keddy's Molybdenum, Walker's Molybdenum, Wallaback Prospect) in or near alaskitic bodies. Vein systems roughly parallel one major joint direction (north-northwest — south-southeast). The geochemical study has included work on major and trace elements. Of the latter the ones showing the most significant variation are Li (granodiorite 124 ppm; adamellite 260 ppm; minor intrusives over 400 ppm) and Sn (granodiorite 5 ppm; adamellite 17 ppm; minor intrusives 52 ppm). Cu (average 10 ppm), Mo (average 1-2 ppm) and W (average < 1.6 ppm) show no systematic variation. Work on the third objective, namely cataloguing each deposit, is still in progress: to date, the following minerals have been identified: cassiterite, wolframite, chalcopyrite, digenite, covellite, bornite and molybdenite. In addition a Nb-Ta mineral (tapiolite) has been identified.

Reference

McKenzie, C. B. and Clarke, D. B.
1975: Petrology of the South Mountain Batholith, Nova Scotia; Can. J. Earth Sci., v. 12, p. 1209-1218.

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70. BARITE, FLUORITE, AND CELESTITE DEPOSITS AND OCCURRENCES IN CANADA

Project 720095

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During the field season the author visited the bedded barite deposits of Nevada and Washington States and took a brief look at the protores of the residual barite deposits of Missouri. A second trip was made to northern Ontario to examine the past producing properties and some potential producers. The Ontario occurrences are classed as hydrothermal fissure fillings and, with the exception of the Penhorwood vein and certain of the Thunder Bay silver veins, are not spatially related to nearby intrusive rocks.

In the Potosi, MO., barite camp, the barite is mined from the weathered residuum of Cambrian Eminence and Potosi Formation dolomites. The Eminence Formation, a massive- to poorly-bedded, medium- to coarse-grained dolomitic limestone is characterized by barite in irregular veinlets, vug fillings, and small areas of breccia associated with calcite, dolomite, and traces of galena, sphalerite, and pyrite-marcasite. It is well exposed by mining in the Settle Mine where the bedrock surface shows few pinnacles and an abrupt change from the residuum to bedrock. The Potosi Formation on the other hand contains a higher proportion of algal reef limestone and dolomite with a high degree of porosity and more abundant coarse voids coated by chalcedony and drusy quartz and filled by coarse bladed barite. The druses exhibit forms that suggest derivation from a variety of poorly preserved reef organisms. The interface between the Potosi dolomite and the overlying residuum is marked by numerous pinnacles that project several feet upwards. In neither formation does the barite occur in lengthy fissure veins but rather in voids in the reef dolomite and discontinuous fractures and breccia zones in the more massive dolomites. The field occurrence in the Cambrian Eminence Formation resembles the vein and cavity fillings seen in the Middle Devonian Stone and Dunedin Formations near Summit Lake (Taylor and Stott, 1973) and along 110 Creek in northeastern British Columbia (Dawson, 1975, p. 257).

The Rossi and Greystone barite mines near Battle Mountain, Nevada, were also examined. Barite mined at the Rossi Mine is partly direct shipping ore and partly a lower-grade type, both of which are mined from a stratigraphic section consisting of interbedded barite, chert, and fossiliferous siliceous argillite of the Devonian Slaven Formation. Barite at the Greystone Mine is also contained in chert of the Slaven Formation. Ore from both mines is dark grey, aphanitic, and thinly laminated.

The Slaven chert of Middle to Upper Devonian age, is a thin bedded, dark radiolarian variety with a limited number of argillaceous chert and limestone lenses. The chert layers consist mainly of quartz and chalcedony with minor dissemianted altered pyrite, clay, carbonaceous material, and white mica. The barite deposits are mainly interbedded with chert and argillaceous chert, rarely with limestone. The bedding contacts are sharp; there is no recognizable wall-rock alteration; the gangue mineral suite is small and sparsely distributed; the ore minerals are limited to barite; and, the orebodies have a few small barite, quartz, and calcite veinlets cross-cutting them. There is no spatial association with major faults or intrusive rocks, although post-ore faulting does complicate the mining of the barite. There are apparently structural and lithological similarities between these orebodies and those recently discovered in the MacMillan Pass area of the Yukon Territory.

The debate regarding the origin of the Battle Mountain area barite deposits continues and the main veiws expressed are diagenetic replacement of marine sediments by barium from the overlying sea-water or the epigenetic replacement of limestones and/or cherts by circulating mineral-bearing solutions.

Visits were made to five bedded barite deposits in Stevens County, Washington State, courtesy of Dr. J. W. Mills, Washington State University. These deposits are characterized by many features that support the sedimentary rather than epigenetic vein fillings or replacement models of origin. In general, the deposits lie conformably and co-extensively within Ordovician to Carboniferous interbedded siliceous argillite, siltstone and rarely limestone, in a low-grade metamorphic terrane that has been subjected to two periods of folding and intrusion by granitic rocks.

The barite occurs in thin laminae (less than 1.0 to 10.0 mm) conformable with the argillites. Two periods of folding have locally developed barite boudins and thickened beds at the hinges of small folds. There is no evidence of wall-rock alteration or of spatial relationship to fissures or faults or to the granite batholiths in the area. The deposits, which are poorly exposed, reach a maximum length of 1600 feet and aggregate thicknesses of a few tens of feet.

The mineralogy is uncomplicated and the barite beds range from medium- to coarse-grained pure barite to very fine grained beds that may contain up to 20 per cent impurities including quartz, calcite, muscovite, clay, and disseminated pyrite. The grain size of the clastic material, the moderate to high content of carbonaceous material, disseminated pyrite, and foetid odour suggest deposition in an anaerobic environment in deep water beyond a limestone shelf environment. The contacts between the barite and limestone are sharp whereas those with the argillite are sometimes gradational. It is probable that these beds were deposited by normal sedimentary processes on the seafloor. The enriched barium content of the bottom sediments accumulated either directly from seawater in the high carbon ooze or by the sedimentation of skeletons of organisms biochemically enriched in barium.

Barite veins and occurrences were visited at sites across northern Ontario as far west as Thunder Bay. These included barite veins in Yarrow, Langmuir, and Penhorwood townships. Silver and lead-zinc veins having barite and/or fluorite gangue were examined in Thunder Bay and Dorion townships. All veins are classed as hydrothermal fissure fillings unrelated to nearby igneous intrusive rocks with the exception of the Penhorwood veins, which occur in granite host rocks, and certain of the Thunder Bay veins that penetrate diabase sills and dykes.

The Yarrow Township barite veins, occurring in Proterozoic arkose and conglomerate southwest of Matachewan in a regional fissure system, are currently being opencut by Extender Minerals Ltd. The Premier-Langmuir Mine south of Timmins was visited and character samples were collected. The wall rocks of the vein consist of banded Archean epidote-amphibolite. Barite was produced from the vein many years ago and might support further production on a small scale. The Cryderman-Ravena vein in Penhorwood Township occupies a fissure system in an Archean granite lens enclosed in Archean mafic to intermediate metavolcanic rocks. Barite- and fluorite-bearing veins were visited at the Creswell shaft, Silver Mountain, Shuniah, Enterprise, and Dorion mines. Barite occurs near the top of vertical veins in the Rove shale beneath diabase sills and was mined from the west end of the Silver Mountain vein. There is a possibility of small barite production from one of these. The structure on which the Dorion and Bishop zinc deposits occur, although not deep, might also support limited production of barite. Fluorite is widespread as a gangue mineral in the silver and lead-zinc fissure veins in the Thunder Bay district but it is not abundant enough for economic production.

References

Dawson, K.R.
1975: Barite, fluorite, and celestite deposits and occurrences in Canada; Geol. Surv. Can., Paper 75-1, pt. A, p. 257-259.

Taylor, G.C. and Stott, D.F.
1973: Tuchodi Lakes map-area, British Columbia; Geol. Surv. Can., Mem. 373.

Project 750069

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Major uranium-bearing sandstones and lignite deposits were visited in the United States of America in order to compare with Canadian occurrences. Expert guidance and hospitality of officials of the Energy Research Development Administration were greatly appreciated. The firsthand field investigations of these important deposits have already proven, and should continue to prove useful in the study of Canadian occurrences.

Work was started on a systematic investigation in Canada of deposits of uranium in sandstone and related deposits, with field examinations initially in the Maritime provinces and in the southern parts of the three westernmost provinces. In the process of investigation, a reconnaissance gamma-ray spectrometer survey was done by automobile between Cranbrook and Vancouver, British Columbia mainly along Highway 3, and between Hope, British Columbia and Maple Creek, Saskatchewan along the Trans-Canada Highway. Additional work was done by car near Beaverdell, British Columbia, in the Rocky Mountain Foothills northwest of Calgary, Alberta, and in the Cypress Hills near Eastend, Saskatchewan. A model 423 Spectrometer, manufactured by Inex Instruments Ltd., was used for the automobile surveys.

Most of the occurrences in Carboniferous rocks in the Maritimes described by Brummer (1958), Gross (1957), Prest *et al.* (1968), Ruzicka (1971), and Lang *et al.* (1962) were visited, and for reference, the reader is referred to these publications. A brief description of new occurrences follows, including those from western Canada. A McPhar Model TV-1 three-threshold scintillometer was used in the field to estimate U and Th abundance (in ppm).

Mill Settlement, New Brunswick

Two new occurrences were investigated between Mill and West Mill settlements — the first at Pete Brook (45°34'24"N, 66°34'55"W, about 2680 m bearing 243° from Mill Settlement), and the second on Oromocto River (45°35'10"N, 66°34'30"W, about 1540 m bearing 256° from Mill Settlement). Both occurrences are apparently in the same stratigraphic unit.

At Pete Brook, friable, cross-stratified, grey-green, coarse- to medium-grained, pebbly, tuffaceous sandstone, siltstone, and conglomerate show anomalous radioactivity. *In situ* measurements indicate 35 to 81 ppm U and 16 to 30 ppm Th through perhaps 15 m of strata, dipping 15° north-northwest, apparently overlying basaltic lavas that are exposed to the south.

At Oromocto River, somewhat friable, very coarse- to medium-grained, pale brown tuff, sandstone, conglomerate, and perhaps ash flows, show anomalous radioactivity. *In situ* measurements vary considerably laterally, but are generally similar in magnitude to those at Pete Brook. However, two small pods measuring about 30 to 40 cm across, indicated approximately 500 ppm U. In these, the high radioactivity is related to black, sooty material associated with deep purple fluorite. It superficially resembles the mineralization at Harvey and York Mills (Gross, 1957) about 40 km west-northwest, and is also in Lower Mississippian rocks.

Beaverdell, British Columbia - Fuki-Donen Prospect

Sands, silts, and gravels immediately underlying plateau basalts at Dear Creek (49°32'20"N, 118°52'55"W) show very high radioactivity, with highest values immediately beneath the basalts: *in situ* measurements are 181 ppm U in the lower, heavily-fractured part of basalts, 427 ppm U in fine grained sands at top 5 cm of the sediments, dropping rapidly off to the order of 100 ppm U in the gravels 60 cm from the top contact. Ruzicka (1971) reports 300 ppm U. At this locality, the sediments overlie the Marron Volcanics; elsewhere, they may directly overlie the basement complex. Of note, surveys show higher than normal values done by automobile between this locality and Rock Creek, British Columbia, part of which is attributed to the aggregates used in constructing the roads.

Lundbreck, Alberta

About 2300 m bearing 195° from the town of Lundbreck, anomalous radioactivity occurs in dark brownish-grey, sparsely white-speckled, fine grained, flaggy sandstone in the Belly River Formation. This occurrence, on the third bench east from the main ridge top, is apparently concordant, about 2 m thick, and can be traced on strike for approximately 200 m. The uppermost beds are most highly radioactive with the upper 30 cm indicating as much as 5 times normal background. However, *in situ* determinations suggest that almost all of this is due to thorium.

Burnt Timber Creek Area, Alberta

During a survey by automobile in the Burnt Timber Creek map-area (Ollerenshaw, 1965), it became apparent that the Blackstone Formation gave gamma spectrometer readings that were slightly higher than background. At 51°30'30"N, 115°17'45", about 500 m west of the bridge over the Red Deer River, the basal transition part of the Cardium Formation showed a distinctive anomaly. Dark silty beds exposed about 100 m north of the road, immediately below fine grained, buff weathering sandstones showed 2 to 3 times background and *in situ* measurements suggested 20 to 30 ppm U and

20 ppm Th. Black silty shales of the Blackstone Formation at the river just south of the road showed twice background readings, with *in situ* determinations of about 20 ppm U and less than 10 ppm Th.

Eastend, Saskatchewan

Ravenscrag Formation

A 30-cm lignite seam on the south side of Anxiety Butte in the Ravenscrag Formation about 15 m below the top showed 3 to 4 times background with *in situ* measurements indicating about 27 ppm U in the uppermost 5 cm. The seam is within fine grained, silty sandstone and muddy siltstone. Ruzicka (1971) reports 500 ppm U, and Cameron and Birmingham (1970) report 825 ppm U_3O_8 from this area.

Cypress Hills Formation

Grit, and gravel scree in the Cypress Hills Formation (Furnival, 1950), about 5 to 15 m above the basal contact about 2 km north of the above Ravenscrag occurrence, contain silicified wood and fragments of Oligocene mammal bones. These bones (cf. Bell *et al.*, 1976) show exceptionally high radioactivity and confirm the fact that uranium-bearing ground waters moved through the Cypress Hills Formation.

References

Bell, R.T., Steacy, H.R., and Zimmerman, J.B.
1976: Uranium-bearing bone occurrences; *in* Report of Activities, Part A, Geol. Surv. Can., Paper 76-1A, rept. 72.

Brummer, J.J.
1958: Supergene copper-uranium deposits in northern Nova Scotia; Econ. Geol., v. 53, no. 3, p. 309-324.

Cameron, A.R. and Birmingham, T.F.
1970: Radioactivity in western Canadian coals; Geol. Surv. Can., Paper 70-52, 35 p.

Furnival, G.M.
1946: Cypress Lake map-area, Saskatchewan; Geol. Surv. Can., Mem. 242, 161 p.

Gross, G.A.
1957: Uranium deposits in Gaspé, New Brunswick and Nova Scotia; Geol. Surv. Can., Paper 57-2, 27 p.

Lang, A.H., Griffith, J.W., and Steacy, H.R.
1962: Canadian deposits of uranium and thorium; Geol. Surv. Can., Econ. Geol. Ser. 16, 2nd ed., 324 p.

Ollerenshaw, N.C.
1965: Burnt Timber Creek; Geol. Surv. Can., Map 11-1965.

Prest, V.K., Steacy, H.R., and Bottrill, T.J.
1968: Occurrences of uranium and vanadium in Prince Edward Island; Geol. Surv. Can., Paper 68-74.

Ruzicka, V.
1971: Geological comparison between east European and Canadian uranium deposits; Geol. Surv. Can., Paper 70-48, 196 p.

72. URANIUM-BEARING BONE OCCURRENCES

Project 750069

R.T. Bell[1], H.R. Steacy[2] and J.B. Zimmerman[3]

Introduction

Fossil bones (Fig. 72.1) including recognizable titanothere fragments, containing significant amounts of uranium were found in gravel screes on the east side of Anxiety Butte about 10 km north-northeast of Eastend, Saskatchewan. These fossils were found generally 5 to 15 m above the base of the Cypress Hills Formation, presumably derived from this formation. *In situ* determination with a McPhar TV-1 scintillometer indicated that the exceptional radioactivity was due to uranium perhaps in excess of 0.08 per cent.

The collected specimens, up to 40 cm long, are composed essentially of carbonate fluorapatite (5.5 per cent CO_2). Autoradiographs made of several polished surfaces (Fig. 72.2) reveal the radioactivity to be uniformly distributed throughout, with no apparent discrete radioactive minerals. A qualitative X-ray fluorescent scan detected strontium, yttrium and uranium, but no thorium. Chemical analyses of three separate bone samples showed 0.13, 0.15 and 0.18 per cent uranium. Values on bone and teeth material from other areas reported by Altschuler *et al.* (1958) and Davidson and Atkin (1953) range between 0.003 and 0.83 per cent uranium. In general the reported values are less than 0.10 per cent. The Cypress Hills material may therefore be considered exceptional.

In view of the foregoing, a brief survey of fossil bone and teeth was made of collections at the Paleontology Division of the National Museum of Natural Sciences, Ottawa, with the following observations:

1) all fossil material from the Cypress Hills Formation was found to be the most radioactive of all material tested;

2) no significant radioactivity was detected in any Canadian Paleozoic material;

3) Cretaceous material from near Eastend, Saskatchewan contained significant radioactivity in only two cases, a ceratopsian femur collected from the Frenchman Formation at Old Man On His Back Plateau, and weakly radioactive ceratopsian fragments from the Frenchman Formation near Morgan Creek (Rocky Creek);

4) Cretaceous material from Alberta was generally not significantly radioactive except for material from three main localities, namely: (i) ceratopsian fragments from the Edmonton Formation about 18 km southeast of Elnora on Big Valley Creek, (ii) ceratopsian fragments at Scabby Butte about 27 km north-northwest of Lethbridge in the Mary River Formation, and (iii) a hadrosaur metatarsal in section 11-14-25W near Little Bow River;

5) plesiosaur fragments (Campanian Stage?) from marine strata at Horton River, N.W.T. showed significant

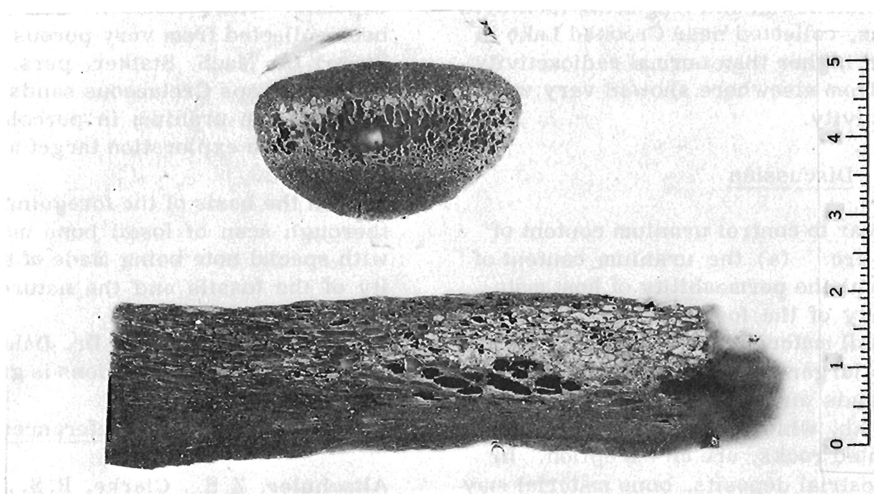

Figure 72.1. Polished longitudinal and transverse sections of uraniferous fossil bones from the Cypress Hills Formation, Saskatchewan. (Scale in cm)

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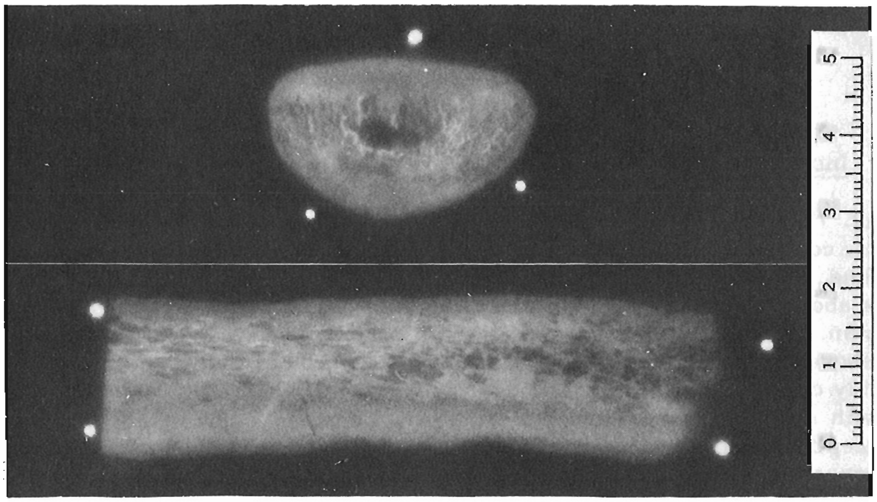


Figure 72. 2. Autoradiographs of the sections illustrated in Figure 72. 1, resulting from a six day exposure on X-ray film. The lighter portions indicate a slight concentration of uranium in the central, porous areas of both specimens, and to a lesser degree along the lower edge, or periphery, of the longitudinal section. This suggests a later enrichment of uranium in those areas providing contact surfaces for circulating ground waters. Such enrichment would not appear to be recent, however, as radium analyses indicate equilibrium conditions. The bright dots reflect grains of pitchblende mounted for orientation purposes. (Scale in cm)

radioactivity, but mosasaur fragments (Maestrichtian Stage?) collected nearby did not; and

6) Pleistocene mammoth fragments collected near Fort Qu'Appelle, Saskatchewan and fragments identified mainly as horse bones, collected near Crooked Lake in Saskatchewan showed higher than normal radioactivity. Pleistocene material from elsewhere showed very weak or negligible radioactivity.

Discussion

Four factors appear to control uranium content of fossil bone. These are: (a) the uranium content of percolating waters, (b) the permeability of host materials, (c) the porosity of the fossil material, and (d) the age of the fossil material. Davidson and Atken (1953) suggest that a larger percentage of radioactive bones is found in sands and gravels than in other sediments. Fossil fish, which are preserved in fine grained, well-indurated rocks, are an exception. In porous, clastic, terrestrial deposits, bone material may be the most favourable, and in some cases the only significant, host for uranium. Percolating ground water containing uranium may therefore produce an exceptional enrichment, as in the described Cypress Hills specimens.

Because the radioactivity of fossil specimens bears witness to the uranium content of percolating groundwater, the fossils may be used as a guide to formations and areas that are favourable to the formation of uranium deposits. In this respect the Cypress Hills Formation, albeit somewhat thin, may deserve more attention than heretofore as a potential for uranium deposits. The Pleistocene material is intriguing having been collected from very porous gravels (Sangamonian Stage) (A. MacS. Stalker, pers. comm.). The source beds (perhaps Cretaceous sands or small Tertiary outliers) of the uranium in percolating groundwaters could be the exploration target and hence could perhaps be isolated.

On the basis of the foregoing we recommend that a thorough scan of fossil bone material be carried out, with special note being made of the relative radioactivity of the fossils and the nature and permeability of enclosing materials.

The co-operation of Dr. Dale Russell in allowing access to Museum collections is gratefully acknowledged.

References

Altschuler, Z.S., Clarke, R.S., Jr., and Young, E.J.
1958: Geochemistry of uranium in apatite and phosphorite; U.S. Geol. Surv., Prof. Paper 314-D, p. 45-90.

Davidson, C.F. and Atkin, D.
1953: On the occurrence of uranium in phosphate rock; 19th Int. Geol. Cong., Algiers, Comptes rendus, sec. 11, fasc. 11, p. 13-31.

ASSESSMENT OF SOME CANADIAN URANIUM OCCURRENCES

Project 750010

V. Ruzicka
Regional and Economic Geology Division

During the 1975 field season selected Canadian uranium occurrences were examined to assess their uranium potential.

In Ontario the main attention was paid to areas with Aphebian sediments which host the conglomeratic uranium deposits in Elliot Lake, Blind River, and Agnew Lake camps. The assessment was conducted in three directions: (a) Collection of data from past and present exploration and mining operations in Sault Ste. Marie and Sudbury mining districts; (b) Field examination of Huronian sediments in Sault Ste. Marie and Agnew Lake areas, and of the uranium occurrences in the Cobalt Plate, namely in Roberts, Grigg, Parkin, Hutton, Creelman, Fraleck, Pardo and Vogt townships; (c) Mineragraphic studies relating to the genesis of the Elliot Lake deposits. These studies have been conducted in close cooperation with the Ontario Division of Mines of the Ontario Ministry of Natural Resources, in particular with J.A. Robertson, head of the Mineral Deposits Section.

For the assessment of the Elliot Lake – Blind River area 36 townships were included in the resource study and selected data from diamond drilling records were entered on computer cards and further processed. Maps, 1 inch to ¼ mile or nearest available scale, were compiled showing contours of the Archean paleosurface, isopachs of the Matinenda Formation or its equivalent and maps and sections showing distribution of factors controlling uranium mineralization. The maps and computerized information on grades, thicknesses and some other geological factors have been subjected to geostatistical analysis. The same method was applied in processing data for assessment of the uranium resources in the Agnew Lake area.

Field examination of the uranium occurrences within the Cobalt Plate was primarily oriented toward areas with recently revived exploration activity and areas covered by mineral claims. The uranium occurrences in the Cobalt Plate can be classified into three lithologic types: (1) The irregularly and weakly mineralized oligomictic conglomerates; (2) The polymictic conglomerates with pebbles of quartz, quartzite and basement rocks; this type contains low grade uranium mineralization, and is locally auriferous; (3) Uranium-bearing argillites and siltstones with relatively extensive, but low grade uranium mineralization. The last type appears to be the most promising for eventual exploitation of low grade ores in this area.

Uranium mineralization in the Aphebian sediments in the Sault Ste. Marie area occurs in oligomictic conglomerates intercalated with Duncan Volcanics and in fine conglomerates of the Aweres Formation. However, the conglomerates in the Duncan Volcanics are thin and of variable thickness and mineralization is irregularly distributed, and those in the Aweres Formation contain only low grade local mineralization.

Some other areas in Ontario were reconnoitered geologically and geophysically by M.N. Turay and R.J. Anderson, and the writer using a gamma-ray spectrometer, model INAX 234 carried in an automobile; immediate follow-up of anomalies was made using portable gamma-ray spectrometer, model McPhar TV-1. The reconnaissance was conducted along the Precambrian/Phanerozoic unconformity above the Frontenac Arch, in the Penokean Fold Belt, in the Port Arthur Homocline, and in the Nipigon Plate.

Anomalous radioactivity was detected in calc-silicate rocks immediately beneath the unconformity approximately 1 mile southeast of Phillipsville, Bastard Township, and in post Grenville pre-Nepean conglomerates and arkoses near Salem, Bedford Township. However, the radioactivity is not caused by uranium mineralization.

Radiometric anomalies related to uranium mineralization were detected in the Sibley Formation southwest of Nipigon. Two selected samples of carbonate rocks with megascopically visible galena mineralization, contained 360 and 540 ppm uranium. This type of mineralization differs from that reported by Tanton (1948) from Sibley Formation in a different locality on the Sibley Peninsula.

Mineragraphic studies on the Elliot Lake ores conducted by H.R. Steacy of the Geological Survey and by S. Kaiman of CANMET allowed some new interpretations of the genesis of mineralization in the Huronian conglomerates. The studies confirmed an apparent analogy with some genetic features observed recently on the Witwatersrand ores by Hallbauer (1975), Pretorius (1975) and by others. It was biochemically proven that biological agents, such as algae and fungi participated in the formation of some Witwatersrand ores. In the Elliot Lake mines the writer observed narrow high grade uranium-bearing black stringers occurring parallel to the bedding. This phenomenon has been known to local geologists for a long time, but no explanation of its origin has been given. If one accepts the analogy with similar phenomena in the South African deposits, then these stringers can be interpreted as first algal structures; they occur at the top of a sedimentation cycle and were deposited during quiet times between depositional cycles (Ruzicka and Steacy, 1976).

In Saskatchewan the resource assessment was oriented toward deposits in the Carswell structure and in the Beaverlodge area.

A separate report on some metallogenic features of the "D" uranium deposit at Cluff Lake, Saskatchewan, has been published (Ruzicka, 1975). The uranium mineralization in the deposits within the Carswell structure is spatially related to the Athabasca/Tazin unconformity, structurally controlled by regional fault systems and occurs in various litho-stratigraphic environments: (a) in the basement rocks of the Tazin Group, (b) in pelites and arenites of the Athabasca Formation, and (c) in breccias.

The Eldorado's Beaverlodge mining operations have reached a depth of approximately 5000 feet below the surface. A drift is being driven on the thirty-second level of the Fay Mine in order to explore and develop the 01 and 09 orebodies. No substantial changes in lithology have been observed at this depth in comparison with the upper portion of the mine.

The remaining uranium exploration targets in Saskatchewan are or will be subjects of other reports.

In Quebec, A. Boyer and R.J. Anderson of the Geological Survey conducted a radiometric survey by automobile in the western part of the eastern extension of the Mount Laurier uranium exploration area. Several radiometric anomalies of a magnitude lower than those in the target area were detected in pegmatitic granitic rocks that at the time of the survey were not covered by mineral claims.

In the Province of Newfoundland and Labrador attention was paid to uranium occurrences in the Makkovik – Kaipokok Bay area which is being explored by Brinex in a joint venture with Urangesellschaft Canada Limited. The present exploration activity is mainly concentrated in that portion of the Aillik Group where the Michelin deposit occurs, i.e. approximately 25 to 30 miles southwest of Cape Makkovik, Labrador. The uranium mineralization in the Michelin deposit is confined to an east-northeast trending zone of quartzofeldspathic rocks derived from rhyolitic tuffs and flows with hematitic alterations in the mineralized areas. The Michelin deposit is presently being explored underground by an inclined adit. Two ore zones have been identified in the mine workings. Further exploration by trenching and drilling is being done to test lateral extensions of the mineralized zones. The mineralization in the "Michelin type" environment is more regularly distributed, but of lower grade than in the Kitts deposit, where the host rock is mainly argillite. Uranium occurrences in the Makkovik-Kaipokok Bay area have been found within a zone more than 100 miles long and locally more than 20 miles wide.

In the Northwest Territories Cominco Limited conducted exploration on the land optioned from Pan Ocean Corporation and affiliated companies in the Baker Lake – Kazan Falls area. Three types of geologic environments are considered as favourable: (a) the basement rocks, which can be tentatively correlated with the rocks of the Tazin Group north of Lake Athabasca, (b) the Kazan Formation, which can be, along with the South Channel Conglomerate, tentatively correlated with the Martin Formation in Beaverlodge, and (c) the Christopher Island Volcanics, which generally exhibit anomalous uranium content. Several other companies, such as Shell Oil Canada Limited, Urangesellschaft Canada Limited and Uranerz Canada Limited, explored for uranium in the Baker Lake area. The assessment of the uranium potential of the whole area is in progress.

In order to investigate possibilities of application of various geochemical and geophysical techniques for the assessment studies two methods were examined recently: (a) the magneto-telluric method using a portable unit developed by Ecole Polytechnique, Montreal, Quebec, and (b) emanometric method using instruments produced by Scintrex (ETR-1) and by E.D.A. Electronics Limited (Radon detector RD-200).

The magneto-telluric method was successfully applied in the Carswell structure by Amok Limited for lithological and structural studies. Its further application is presently being studied by the Geological Survey.

The emanometric method was tested in shallow residual soils west of Ottawa. The soil gas, pumped out of slim holes, was analyzed using the two above mentioned emanometers simultaneously. The instruments were operated by R.J. Anderson and C.D. Laing. The tests showed a good correlation and high resolution of both instruments. It appears that in practical application for assessment purposes the ETR-1 is suitable for regional reconnaissance surveys, whereas the RD-200 can be applied for detail assessment surveys.

References

Hallbauer, D.K.
1975: The plant origin of the Witwatersrand 'Carbon'; Minerals Sci. Eng., v. 7, no. 2, April, 1975, p. 111-131.

Pretorius, D.A.
1975: The depositional environment of the Witwatersrand goldfields: A chronological review of speculations and observations; Mineral Sci. Eng., v. 7, no. 1, 1975, p. 18-47.

Ruzicka, V.
1975: Some metallogenic features of the "D" uranium deposit at Cluff Lake, Saskatchewan; *in* Report of Activities, Part C, Geol. Surv. Can., Paper 75-1C, p. 279-283.

Ruzicka, V. and Steacy, H.R.
1976: Some sedimentary features of conglomeratic uranium ore from Elliot Lake, Ontario; *in* Report of Activities, Part A, Geol. Surv. Can., Paper 76-1A, rept. 74.

Tanton, T.L.
1948: Radioactive nodules in sediments of the Sibley Series, Nipigon, Ontario; Roy. Soc. Can. Trans., v. 42, Ser. III, Sec. 4, p. 69-75.

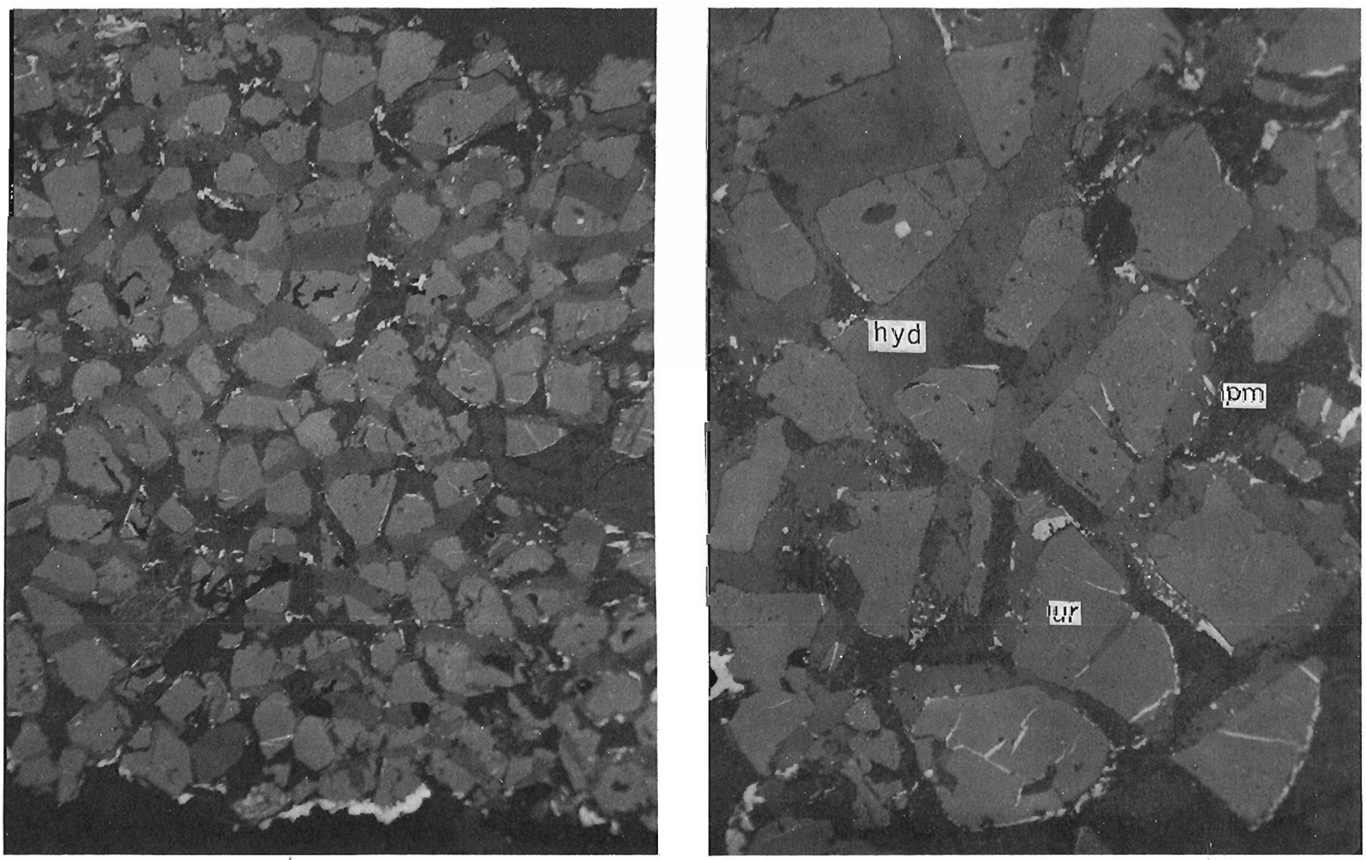
74. SOME SEDIMENTARY FEATURES OF CONGLOMERATIC URANIUM ORE FROM ELLIOT LAKE, ONTARIO

Project 750010

V. Ruzicka[1] and H.R. Steacy[2]

During an underground examination of the conglomeratic uranium ores at Elliot Lake, Ontario, the senior author observed some narrow, highly radioactive beds that possessed unique sedimentary characteristics. Accordingly, specimens were collected for closer examination in the hope that they could contribute to an understanding of the origin of the uranium mineralization. The specimens are from a stope in the lower reef at the Denison Mine.

One ore specimen, fairly typical of those collected, is illustrated in Figure 74.2 and its autoradiograph in Figure 74.3. The specimen displays several narrow beds with sharply defined contacts. The sedimentary nature is readily apparent, particularly in the lower bed which grades from a quartz-pebble-rich base upward through a pyrite-brannerite-monazite-rich section to a black highly radioactive band at the top. In polished section (Fig. 74.1a, 74.1b) this black band is composed largely of fragmented uraninite crystals cemented by a sulphur-rich hydrocarbon. Under crossed nicols (Fig. 74.4) the hydrocarbon exhibits, for the most part, a uniform extinction and a columnar structure, whose fibres are oriented normal to the bedding plane. The uraninite contains several per cent thorium.



A B

Figure 74.1. Photomicrographs of a polished section of a black, highly radioactive band showing fragments of uraninite (ur) cemented by sulphur-rich hydrocarbon (hyd). Sulphur was the only element detected in any significant amount in the hydrocarbon by energy dispersive analysis with the electron probe. Fine grained polymineralic patches (pm) occur within the band and likely represent finely-comminuted minerals and clay fraction. The white mineral is mainly galena (a-X 60; b-X 163).

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[1]Regional and Economic Geology Division.

[2]Central Laboratories and Administrative Services Division.

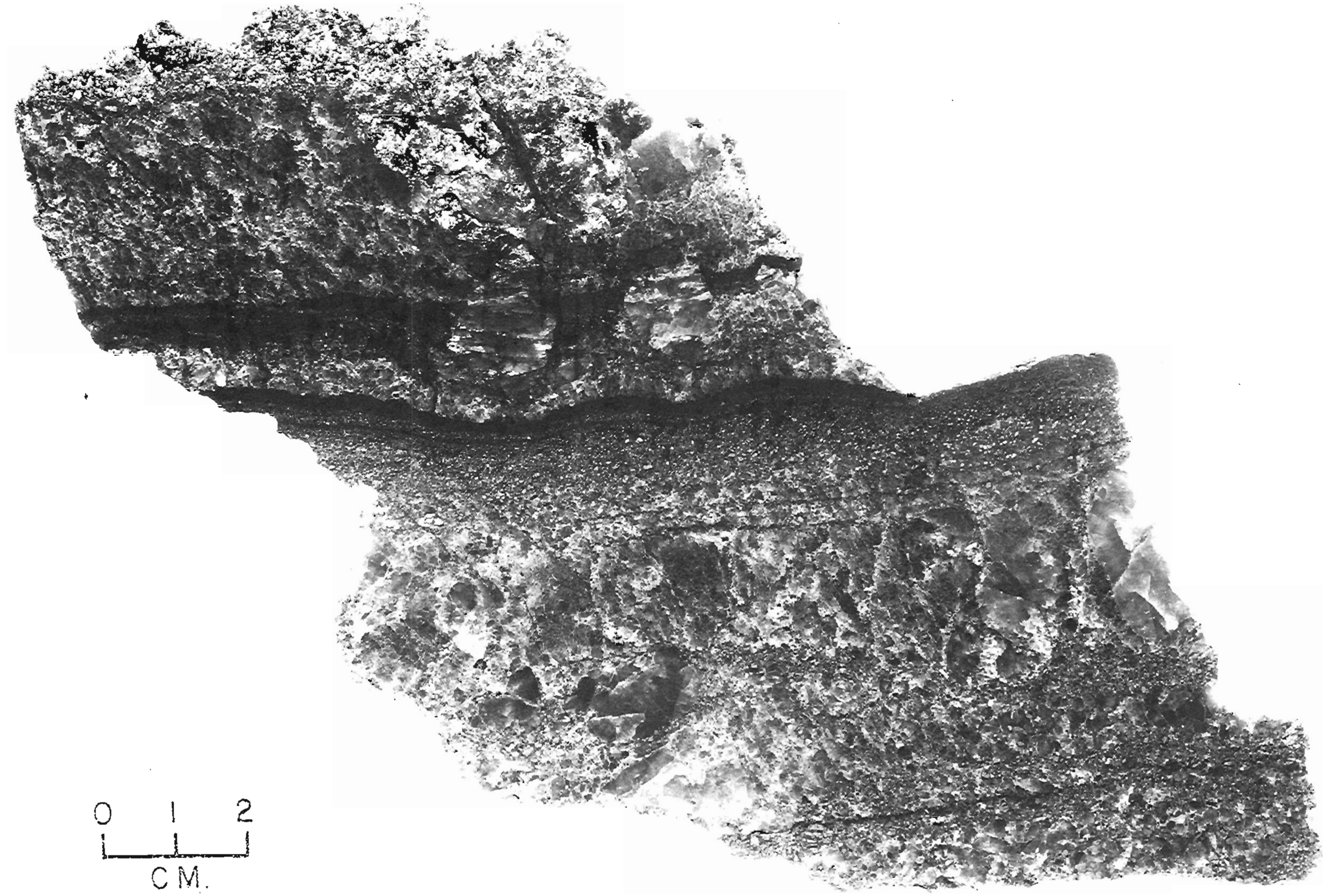

Figure 74. 2. Photograph of a polished specimen of conglomeratic uranium ore from the Denison Mine, Elliot Lake, Ontario, oriented as collected. (GSC-202872 B)

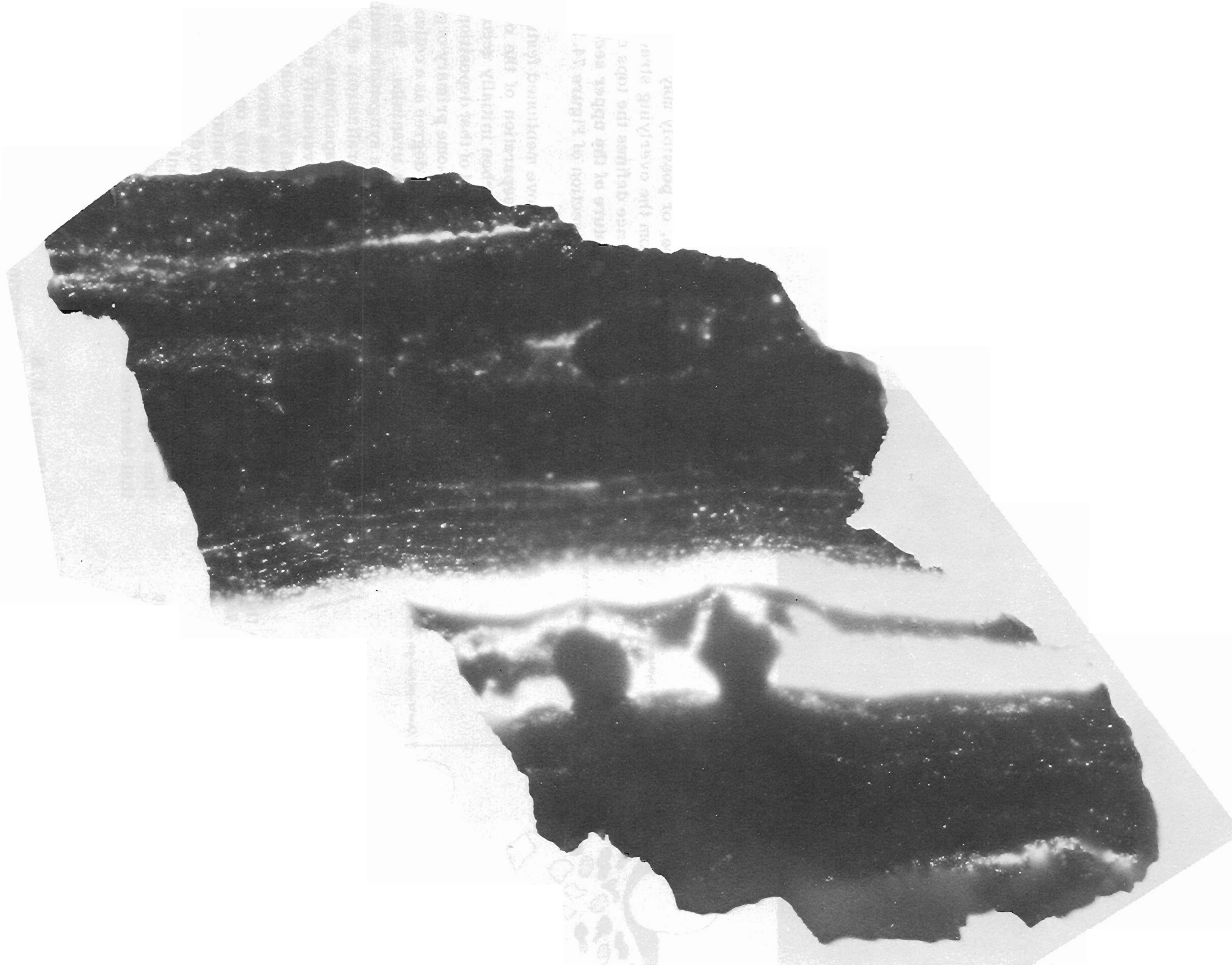

Figure 74.3. Autoradiograph of the specimen shown in Figure 74.2. Exposure 24 hours.

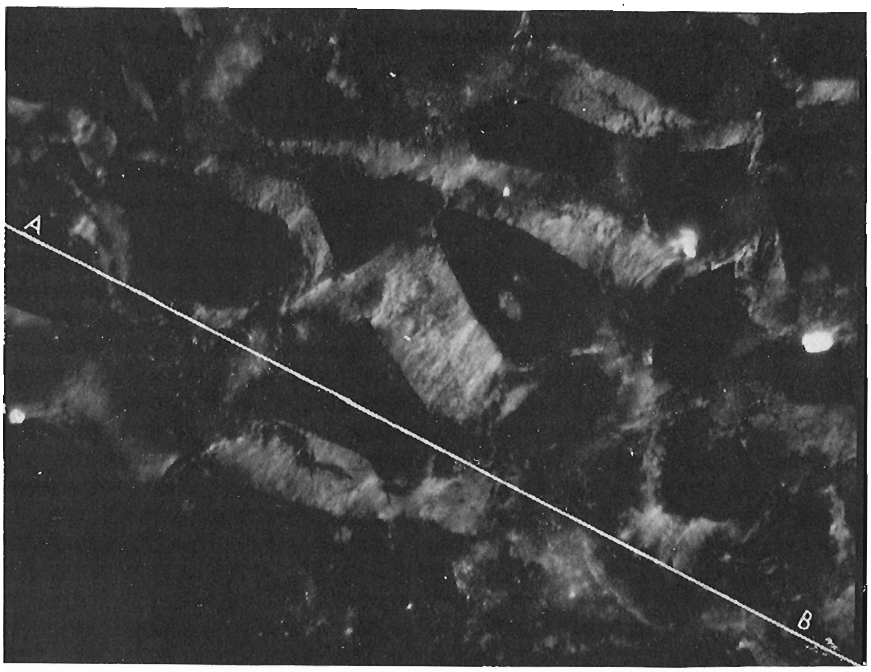

Figure 74. 4

Photomicrograph showing the columnar structure of the hydrocarbon as revealed under crossed nicols. Line A-B denotes the plane of bedding. Note that the columnar structure is roughly normal to the bedding plane, and that the long axes of the uraninite grains are roughly parallel to it (X 163).

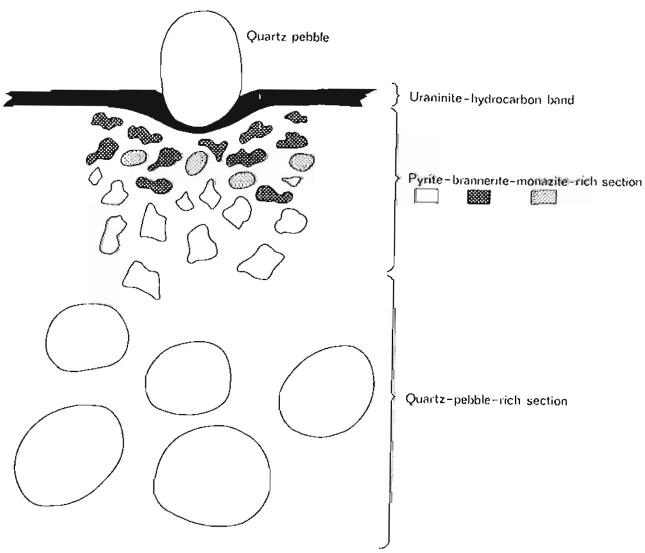

Figure 74. 5. Generalized interpretative section through one of the narrow beds illustrated in Figure 74. 2.

Brannerite is present as ragged, irregular, to, occasionally, subrounded grains, and monazite as subrounded grains. Pyrite occurs as angular to subrounded grains many of which are fractured and broken; the grains show a decreasing order in size toward the top of the bed.

Attention is drawn particularly to the small quartz pebbles which depress the top of the central bed. These pebbles, and the manner in which they and others have impacted or burrowed into the beds, initially attracted the collector's attention. They apparently appeared late in the depositional sequence, or possibly may be post-depositional and derived from the overlying stratum, but in any event their presence defines the tops of the beds and the sedimentary nature of the upper section. A generalized interpretative section of Figure 74. 2 is shown by Figure 74. 5.

Our interpretation of the above mentioned features is that they reflect a hydraulic separation of the ore constituents and that the hydrocarbon initially accumulated as an oily substance at the end of that depositional cycle, possibly being derived from some primary organic matter, and possibly acting to some degree as a collector, or natural flotation agent, for the uraninite. The features are strongly suggestive of a syngenetic sedimentary origin for the uranium mineralization, at least in the area of the collection of the specimens. The hydrocarbon in the Elliot Lake ores variously described as radioactive hydrocarbon, uranium hydrocarbon or, more generally 'thucholite' — deserves more exacting study such as that conducted recently on the Witwatersrand (Hallbauer, 1975) uranium ores. For example, the possibility that the hydrocarbon may represent some primitive form of plant life in Precambrian time as postulated by several scientists is a hypothesis that deserves consideration.

Reference

Hallbauer, D.K.
1975: The plant origin of the Witwatersrand 'Carbon'; Minerals Sci. Eng., v. 7, no. 2, p. 111-131.

75. THE NATKUSIAK BASALTS, VICTORIA ISLAND, DISTRICT OF FRANKLIN

Project 750006

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Introduction

The Natkusiak basalts are the only known surface expression of the widespread Franklin magmatic province of late Hadrynian times (Fahrig *et al.*, 1971). Accordingly, they offer the prospect of a stratigraphic framework to which features of the intrusives, for example the magnetic reversals, might be referred. In addition, the Natkusiak basalts contain numerous copper showings and, for this reason, are of considerable economic interest.

Five weeks of the past summer were spent in field studies of the Natkusiak Formation and associated intrusives in the northeastern part of the Minto Arch. This was done in conjunction with stratigraphic studies of the Shaler Group by J.D. Aitken, and paleomagnetic studies of both igneous and sedimentary rocks of the Minto Arch by J. Foster. Very able assistance was given in the field by Mimi Fortier, William N. Houston, Mary T. LaHam and Dan Gato.

General Geology

The geology of Victoria Island as it is presently known can be attributed mainly to the work of Thorsteinsson and Tozer (1962) and Christie (1964). In the northeasterly part of the Minto Arch where the present work was done, the Shaler Group comprises five formations of principally carbonates and sandstones, with lesser shale and gypsum. These are profusely intruded by dolerite sills and overlain by the Natkusiak Formation (Fig. 75.1). According to Thorsteinsson and Tozer, the contact is a disconformity. In detail it appears to be conformable, but on a general scale, evidence for an unconformity can be seen. In Figure 75.1 a lobe of the Natkusiak Formation overlaps the trace of some of the sills to the south. Assuming that the sills reflect the bedding of the sediments, a disconformity between the Shaler Group and Natkusiak basalts is indicated. This does not mean that a hiatus exists between intrusion of the sills and extrusion of the flows; the sills being simply guided by the structure of the sediments could have been introduced at any later time.

The Shaler Group is assumed to be Hadrynian, on the basis of a K/Ar age for the sills of about 640 m.y. (Christie, 1964) but an Helikian age, suggested by Aitken *et al.* (1973), is a distinct possibility in view of the disconformity which may separate the sedimentary and igneous events. Paleomagnetic studies should indicate whether or not the sedimentary and igneous rocks are of approximately the same age and, hopefully, the age of the igneous rocks can be corroborated by a Rb-Sr isochron.

Dolerite Sills and Dykes

Dolerite sills and dykes in the study region in the northeastern part of the Minto Arch are shown in Figure 75.1. The sills were numbered downward in stratigraphic order from 1 to 13. More sills are present below 13, but were not reached in this study. Most of the sills were sampled for geochemical and for paleomagnetic purposes, as shown in Figure 75.2. The numbered sills are the major ones but additional small sills not shown in Figure 75.1 do occur in places between the larger sills, and these were designated in the field as sill X±1, X±2, ... with X representing the nearest major sill. Sills in the same stratigraphic interval (i.e. between alternate numbered sills) but separated laterally, are distinguished by letters, for example 2 and 2a.

Most of the sills dip at between 5 and 10 degrees and in consequence, thicknesses are not easily measured. They form cuestas with scarps generally ranging from 50 to 150 feet high. Where it has been possible to determine the thickness of the sills with accuracy, they range from 75 to 300 feet. Most are probably about 150-200 feet thick. A few of the sills can be traced laterally for many miles; sills 3, 10, 11 and 13 are all about 30 miles long.

The sills are composed of olivine-bearing dolerite, with olivine seemingly a minor constituent. Some of the thicker sills, such as 6b, contain patches of pegmatite and patches and sharp-walled veins of pinkish granophyre.

Dolerite dykes of a northwesterly-striking swarm are, in part, feeders of the sills. This is demonstrated in Figure 75.3, which is a sketch drawn from airphotos of a set of sills just east of Killian Lake. The northwesterly-trending dyke in the centre of the figure can be seen in the field to feed westward without break into sills 7a and 6a, and eastward into 6b. All three sills are at different stratigraphic levels. Southeastward the dyke cuts through sediments that overlie sill 6a and forms the western boundary of sill 6d. Presumably it also feeds this sill. The eastern boundary (which was not examined in the field) sharply transects the trend of underlying sediments and may similarly be a dyke. It appears from the airphotos, to pass upward through intervening sediments to sill 4, where it also forms the eastern boundary of that sill. A very similar transition between dykes and sills is evident on the southwest side of Killian Lake (Fig. 75.1, sill 4a).

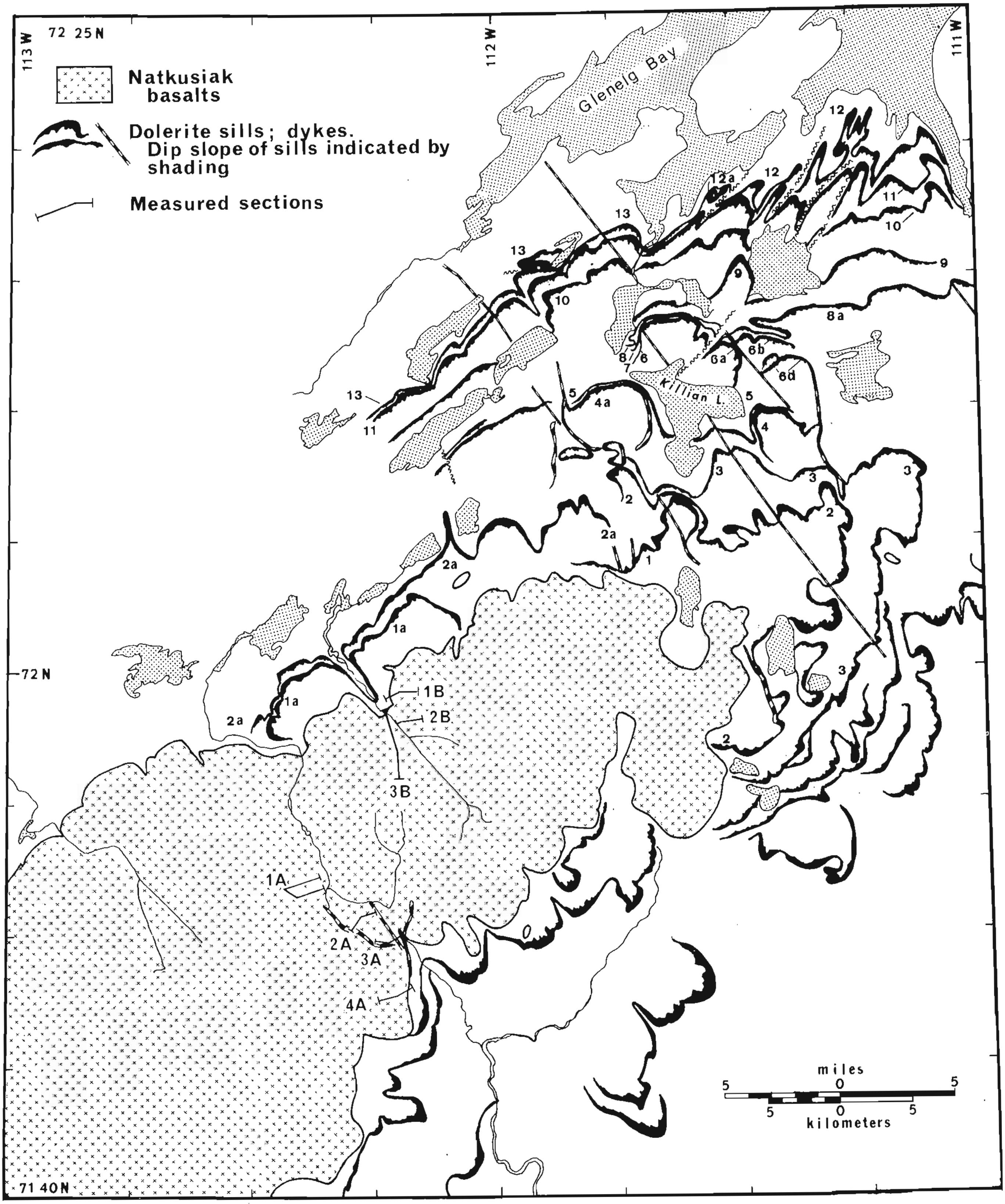

Figure 75. 1. Igneous rocks of the northeastern part of the Minto Arch, Victoria Island. The sills are numbered from 1 to 13 in descending stratigraphic order from the base of the Natkusiak Formation.

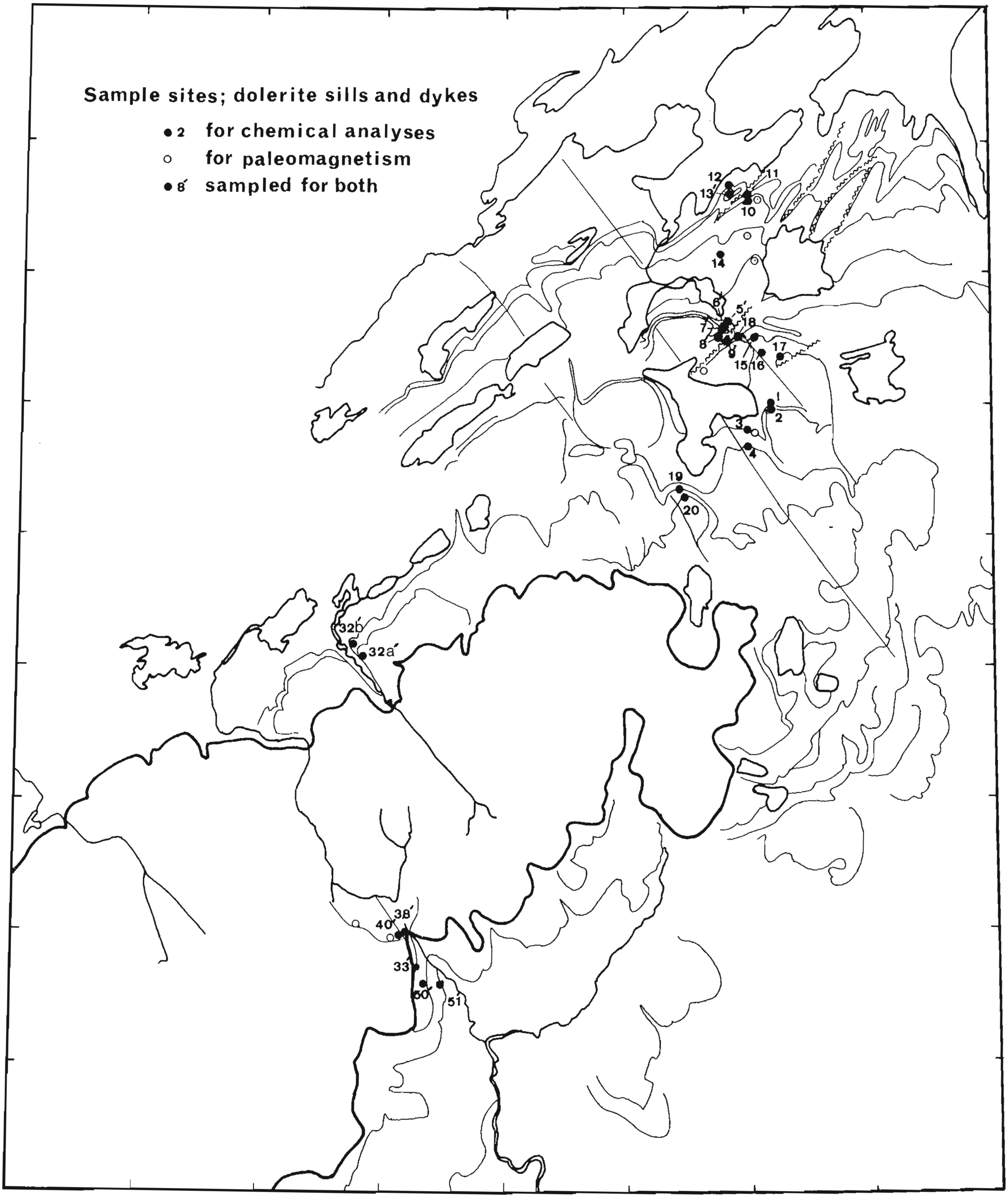

Figure 75. 2. Sample sites for dolerite sills and dykes in the same region as shown in Figure 75. 1.

The impression gained from these examples is that individual dykes may feed a number of sills on the same or alternate sides of the dyke as they pass up through the sequence. That the dykes and sills overlap in time there can be no doubt. The additional observation that northwest-trending dykes also cut the Natkusiak Formation is strong evidence that all three are facies of the same continuing magmatic episode.

Natkusiak Basalts

Seven stratigraphic sections measured in different parts of the Natkusiak Formation, together provide a fairly complete north-south cross-section of the formation at about 112°15'W (Figs. 75.1, 75.4). Both the northerly (1B-3B) and southerly (1A-4A) sets of sections are intracorrelated by flows that are continuous between adjacent sections. These are shown in Figure 75.4. The sets of sections are equated by their common base at the lower contact of the formation. The uppermost level of section 1A is the highest level in the Shaler Mountains; hence the total preserved sequence of Natkusiak basalts is probably represented by these sections. Its thickness is about 2400 feet.

Stratigraphic sections were measured from profiles mapped by chain, compass, and barometer surveys. Composite samples of clean, fresh chips for chemical analyses were taken of every flow that was suitable for sampling. The locations of geochemical and paleomagnetic samples taken through the sequence are shown in Figure 75.4.

Special mention should be made of the difficulties of attempting to do detailed stratigraphic work and sampling in this region. Frost shattering is extreme,

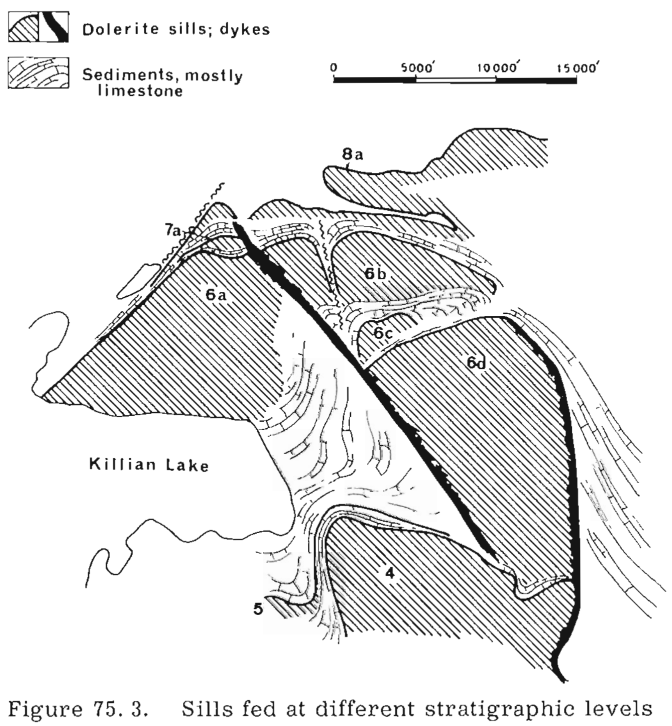

Figure 75.3. Sills fed at different stratigraphic levels and on alternate sides by a northwest-striking dolerite dyke. Sketch from airphotos.

and most slopes are composed of rubble. Even the slopes identified as escarpments on airphotos commonly prove to be steep gradients of loose material. Fortunately, in most cases, the loose blocks have not moved far from their source, and tops and bottoms of flows can be identified in the rubble. The best exposed sections are near the base of the formation, where the flows are thick and outcrop in spectacular escarpments.

The Natkusiak Formation is at the core of the gently folded Holman Island syncline (Thorsteinsson and Tozer, 1962). The synclinal axis is not shown in Figure 75.1, but lies along the centre of the belt of Natkusiak basalts. The dips on both limbs are similar and range between 2 and 8 degrees. Some faulting can be recognized in airphotos, but the major lineaments, northwest-trending valleys, are probably derived from the erosion of dykes.

The Natkusiak basalts are typical plateau basalts which have erupted, for the most part, in a subaerial environment. Each flow comprises a massive base, with the exception of a thin amygdaloidal zone at the bottom, and a highly amygdaloidal top. The top ranges from $\frac{1}{4}$ to 1/10 the thickness of the flow, although some thin flows are almost entirely amygdaloidal. Most flows have a pronounced 'sheeting' structure which is parallel or subparallel to their contacts, and is most closely-spaced at their tops. Weathering of the sheeted basalts is commonly severe. The thinner flows are sheeted throughout and the attendant weathering makes a fresh sample difficult to obtain. In the thicker flows, the lower portion is generally unsheeted and fresh. Almost certainly the sheeting structure is inherited from a primary feature such as flow layering, but it has been much modified by weathering and produces a very friable rock.

The basalts are generally dark grey to green, aphanitic or finely granular rocks with few distinguishing features. A few flows are inconspicuously feldspar phyric. Even the flows named in Figure 75.4 that were used to correlate between sections are generally indistinctive, apart from their persistence and topographic expression. The Rosetta flow does have a distinctive grey amygdaloidal top that marks it out from neighbouring flows in the three sections in which it outcrops, but this may not be a feature of great persistence.

Many of the flows weather to an orange-brown colour, suggesting that these basalts are fairly rich in iron. Amygdaloidal fillings are mainly calcite, chlorite, and quartz, but prehnite and native copper were also observed in places. Particles of native copper are minor constituents of some of the flows, but their distribution does not seem to be related to stratigraphic level as it does in the Coppermine River flows (Baragar, 1969).

In a general way, the thickness of the individual flows diminishes upward in the stratigraphic sequence (Fig. 75.4). The lowermost flows range mainly from 150 to 200 feet thick, whereas those in the upper part of the sequence are more commonly 50 to 75 feet thick. This probably records a declining supply of magma as the eruptive period progressed.

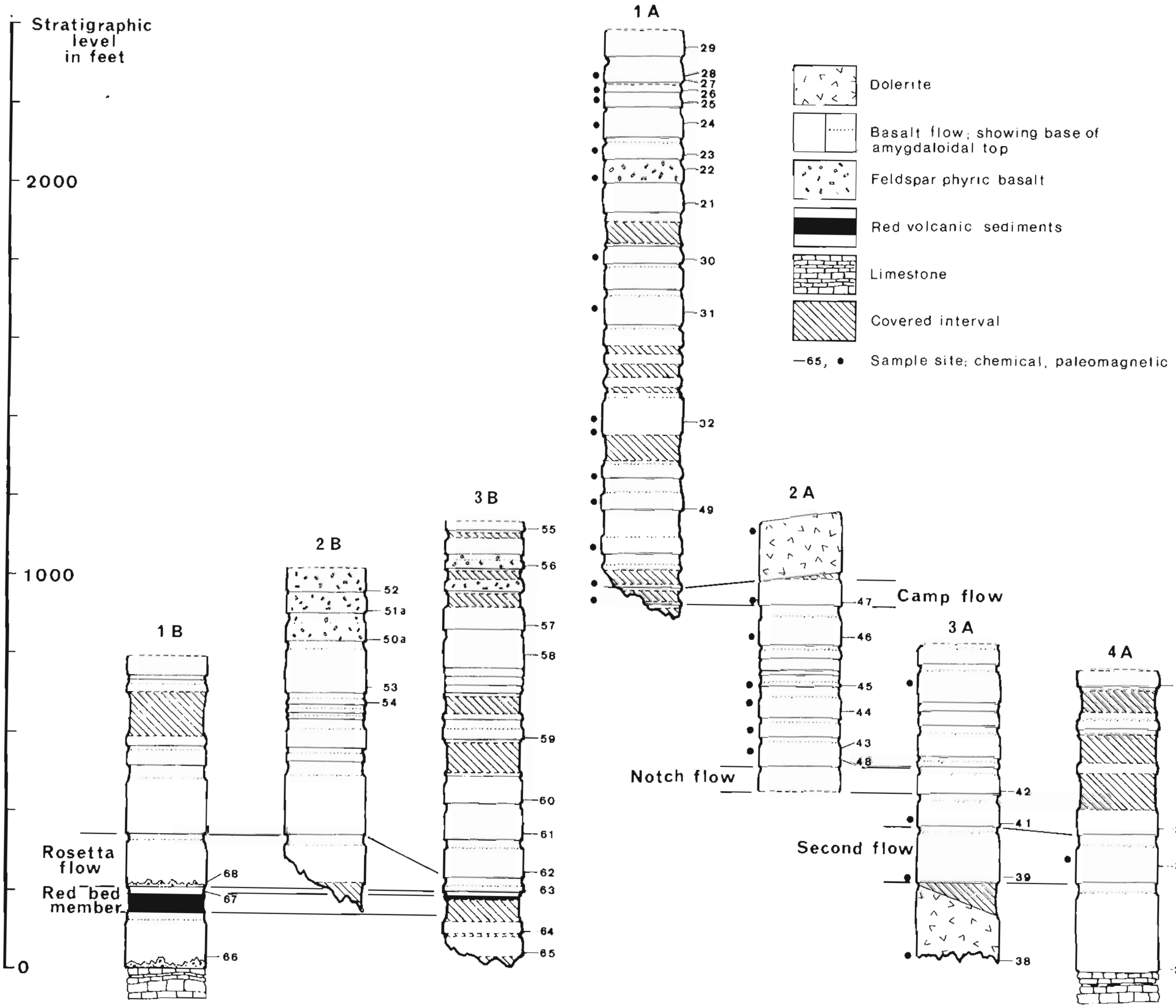

Figure 75. 4. Stratigraphic sections of the Natkusiak basalts measured along the profiles shown in Figure 75. 1. The stratigraphic level is carried between sections by flows (named in the diagram) that can be traced from one section to the next.

Individual eruptions appear to have followed one another in rapid succession throughout the sequence, for very little interflow sediment is present. The red bed member overlying the lowermost flow, in sections 1B-3B, is the only sedimentary unit of note within the formation. It is a coarse clastic composed of angular to rounded, red and green clasts, ranging from about 1 to 5 cm in diameter in a friable, reddish, fine grained matrix. No doubt it is mainly of volcanic derivation. In section 1A, much higher in the sequence, a string of shaly plates in the felsenmeer at one level and reddish soil at another, indicate the presence of some interflow material.

Although the major part of the Natkusiak basalts appears to have been erupted in a subaerial environment, the lower two flows in section 1B show evidence of extrusion in water (Fig. 75. 4). The lower 10 to 40 feet of both flows is severely shattered by polygonal fracturing and the interstices filled with hyaloclastic-like material and calcite. Aphanitic basalt re-intrudes the shattered zone in irregular dykes and masses. Upward, the zone passes into massive basalt with a gradual diminution of the polygonal fracturing. These are probably the effects produced by the extrusion of thick basaltic flows into shallow water.

Economic Aspects

Although the presence of native copper on Victoria Island was reported as early as 1913 by Stefansson (Thorsteinsson and Tozer, 1962) no serious exploration took place until recent years. In 1968 and 1969, extensive exploration of the Natkusiak Formation by several mining companies resulted in the discovery of numerous small copper prospects, but nothing of economic significance. The present field work can add little to what is already known. Rare native copper occurrences in flow tops, copper particles within the lavas, and two or three copper nuggets of a few ounces weight, were all that were seen this summer. Chemical analyses of the flows should provide some information on the distribution of copper within the stratigraphic sequence, and may help to limit the search to that part of the section which would seem to be most favourable.

References

Aitken, J.D., MacQueen, R.W., and Usher, J.L.
1973: Reconnaissance studies of Proterozoic and Cambrian stratigraphy, Lower Mackenzie River area (Operation Norman), District of Mackenzie; Geol. Surv. Can., Paper 73-9.

Baragar, W.R.A.
1969: The geochemistry of Coppermine River basalts; Geol. Surv. Can., Paper 69-44.

Christie, R.L.
1964: Diabase-gabbro sills and related rocks of Banks and Victoria Islands, Arctic Archipelago; Geol. Surv. Can., Bull. 105.

Fahrig, W.F., Irving, E., and Jackson, G.D.
1971: Paleomagnetism of the Franklin Diabases; Can. J. Earth Sci., v. 8, p. 455-467.

Thorsteinsson, R. and Tozer, E.T.
1962: Banks, Victoria and Stefansson Islands, Arctic Archipelago; Geol. Surv. Can., Mem. 330.

Project 730039

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This is a study of late Aphebian volcanism and plutonism at the north end of the Great Bear Batholith (Fraser *et al.*, 1972), the late orogenic magmatic belt of the Coronation Geosyncline (Hoffman, 1973). The volcanic rocks include enormous thicknesses of continental intermediate to acidic ignimbrites and lavas intruded by comagmatic epizonal plutons.

In 1975, the final year of the project, 1: 250 000 scale mapping of the Sloan River sheet (Fig. 76. 1) was completed. In addition, 1: 50 000 scale mapping of the volcanic rocks along the east shore of Great Bear Lake was extended from Hornby Bay to Conjuror Bay. Detailed mapping of the differentiated monzonite-syenodiorite-syenite complex and adjacent rocks south of Rainy Lake on the Camsell River (Fig. 76. 2) will be the basis of a B. Sc. thesis by Rein Tirrul. This and related igneous complexes may control the silver-uranium-arsenide vein mineralization of the region.

Structural Relations

Figure 76. 1 shows the numerous faults in the area. All the major northeast faults originate with dextral strike-slip but many have subordinate, and probably subsequent, dip-slip as well. The major faults postdate all igneous rocks except diabase, but detailed mapping reveals relatively minor synvolcanic faults of generally similar trend. The two largest faults in terms of displacement extend up the Fault River valley from Hornby Bay and up the Tilchuse River valley from Conjuror Bay.

The volcanics are broadly folded with gently curved hinges and limbs dipping generally less than 45 degrees. The folds trend northwest, normal to the trend of faults and oblique to the belt as a whole. Northeast dipping fold limbs are the longer and thus successively younger volcanic units are exposed to the northeast. Along the eastern edge of the belt, units near the top of the volcanic pile rest unconformably on synorogenic metamorphic and granitic rocks of the Coronation Geosyncline. This indicates that the older volcanic units do not project indefinitely to the northeast beneath younger units and implies a northeastward shift in volcanism with time. Basement is not exposed within the belt in the Sloan River area but McGlynn (1976) reports that rocks perhaps correlative with the oldest volcanic unit rest unconformably on granitic rocks at Hottah Lake, 70 km south of Conjuror Bay. These granitic rocks may be part of the prevolcanic basement or an unroofed synvolcanic pluton.

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Volcanic Stratigraphy

The volcanic pile is here subdivided into 10 stratigraphic units corresponding to map Units 1 to 10 in Figure 76. 1. A crude alternation exists such that the odd numbered units are dominated by basic to intermediate volcanics and the even numbered units by acidic volcanics and sediments. In the only previous formal stratigraphic subdivision (Robinson, 1933; Kidd, 1933), the Echo Bay Group corresponds closely to Unit 3 and the Cameron Bay Group to Unit 4.

Unit 1 (4200 + m)

The oldest stratigraphic unit is known only in the Conjuror Bay area but may be correlative with rocks described by McGlynn (1976) at Hottah Lake. In Conjuror Bay, the lower part of the unit (Fig. 76. 2, Unit 1) consists of crossbedded white quartzite interbedded with dark siltstone and cut by many basalt dykes. The upper part (Fig. 76. 2, Unit 2) comprises basalt flows with thin interbeds of laminated air-fall tuff and mudstone (Fig. 76. 2, Unit 2a), quartz-pebble conglomerate (Fig. 76. 2, Unit 2b), and argillaceous dolomite which is exposed on islands just north of the boundary of Figure 76. 2). A few of the basalt flows are pillowed and some, perhaps sills, are very coarsely plagioclase-porphyritic.

Units 2 (4500 m)

The bulk of this unit consists of rhyolite ignimbrite (Fig. 76. 2, Unit 3) characterized by subrounded clasts of rhyolite lava, broken phenocrysts of quartz and alkali feldspar, and flattened pumice fragments dispersed in a flesh-coloured matrix of welded ashstone. Discontinuous horizons of basalt flows and air-fall tuff (Fig. 76. 2, Unit 3s) with lenses of lithic sandstone and conglomerate are in the ignimbrite pile. Conglomerates near the base of the unit contain cobbles of quartzite, rhyolite lava, porphyritic basalt or andesite, and medium grained quartz monzonite. The latter show how early in the eruptive history subvolcanic plutons were emplaced and unroofed. Funnel-shaped plug-domes of massive rhyolite (Fig. 2, Unit 4), are interspersed in the ignimbrite pile with wedges of rhyolite-pebble conglomerate and ignimbrite with unusually coarse rhyolite clasts around them. South of Rainy Lake, the ignimbrite pile is overlain by cross-bedded white orthoquartzite (Fig. 76. 2, Unit 5) but, elsewhere, conglomeratic sandstones high in the ignimbrite pile are derived from the ignimbrite and are much less mature. The top of the unit consists of laminated tuffaceous mudstone with calcareous lenticles (Fig. 76. 2, Unit 6), in which are many thin rhyolite lava flows with aprons of felsite-lithic sandstone. Some representatives of the monzonite-syenodiorite-syenite

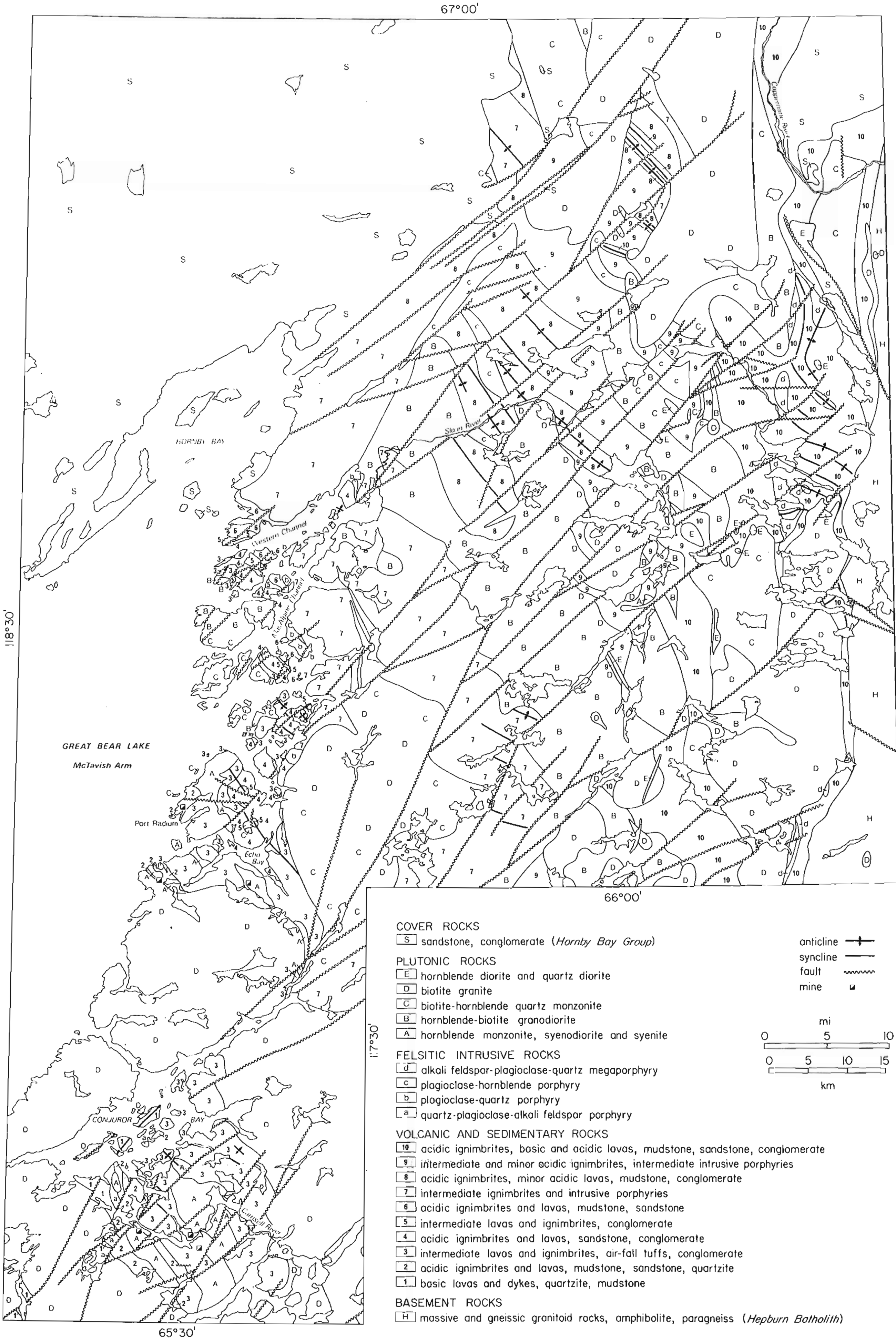

Figure 76.1. Geology of the Sloan River map-area.

Figure 76. 2. Geology of the Rainy Lake area showing two intrusions of the monzonite-syenodiorite-syenite suite.

suite intrude the upper part of Unit 2 and the mudstones are silicified, brecciated and intruded by magnetite-apatite pipes.

Unit 3 (4200 m)

The lower part (Fig. 76. 2, Unit 7) of this unit consists of coarsely plagioclase-porphyritic andesite lava flows intercalated with bedded crystal-lithic air-fall tuffs of similar composition. The tuffs predominate at Rainy Lake but at Echo Bay, where this part of the unit attains its maximum thickness of 2600 m, flows prevail except near the base. Many of the flows have oxidized and brecciated tops, and some have flow-foot breccias. Some of the tuffs are rippled and cross-bedded, and are more commonly related genetically to the overlying than the underlying flow. At the east end of Rainy Lake and the Southwest Arm of Echo Bay, many of the lower flows are hornblende-porphyritic basalts or basaltic andesites. The flows and tuffs are overlain by an andesite-pebble conglomerate (Fig. 76. 2, Unit 8) of variable thickness. Locally, this conglomerate rests unconformably on, and contains boulders of, plutonic rocks of the monzonite-syenodiorite-syenite suite that intrudes the underlying andesite pile, thus proving that this plutonic suite is synvolcanic. The unconformity is well exposed at Uranium Point on Balachey Lake.

The upper part (Fig. 2, Unit 9), of the unit consists of dacite ignimbrite composed of broken phenocrysts of plagioclase and quartz in a dark green or red-brown matrix of welded ashstone. Much of the ignimbrite pile is massive but horizons containing flattened, highly recrystallized, pumice fragments occur near the base of the pile. North of Balachey Lake, air-fall tuffs and lava flows or hornblende-porphyritic basalt or andesite (Fig. 76. 2, Unit 9a) occur locally within the ignimbrite pile. On the west side of Achook Island, south of Western Channel, a conglomerate lens within the ignimbrite pile contains boulders of granodiorite derived from the large pluton exposed on Hogarth Island to the south. This pluton intrudes the andesite lavas and tuffs below the ignimbrite pile and, therefore is also demonstrably synvolcanic. As shown in Figure 76. 2, the ignimbrite pile truncates the older rocks of Unit 3, in places resting directly on rhyolite ignimbrites of Unit 2. This discordance probably results from warping and block-faulting of the older rocks during emplacement of the synvolcanic plutons.

Unit 4 (1160 m)

The bulk of this unit consists of coarse friable lithic sandstone and conglomerate, and rhyolite ignimbrite rich in clasts of rhyolite lava. Their mutual relations are complex both in section and plan, and it is evident that the individual ignimbrite flows were dissected soon after deposition, producing a highly irregular topography where remnant pinnacles of ignimbrite were surrounded by channels filled in part by alluvial sediments and in part by succeeding ignimbrites. Thus, although the unit consists mainly of ignimbrite on Mackenzie Island in Lindsley Bay, the ignimbrite is intimately mixed with sediments on both the north and south sides of the bay; ignimbrites are all but excluded at Echo Bay, 5 km to the south. In addition, a massive rhyolite plug-dome, at least 6 km in diameter, is exposed on Rocher Rouge and Achook islands near the mouth of Western Channel, and a strongly flow-banded rhyolite lava flow occurs on islands near the mouth of Lindsley Bay. The sediments of Unit 4 are derived mainly from the various rhyolites and sediment transport was to the north and northeast.

Unit 5 (810 m)

This unit is similar to but thinner than Unit 3. Its lower part comprises an andesite stratovolcano, centred over Achook and Cornwall islands, of plagioclase-porphyritic hornblende andesite flows and breccias. It attains a maximum thickness of 810 m but thins to less than 110 m within 5 km of the centre. The lower flanks of the volcano are overlain by conglomerate containing boulders mainly of andesite, but also of granodiorite and aplite from the Hogarth Island pluton. Above the conglomerate is a thick dacite ignimbrite, massive except near the top, that almost buries the underlying volcano.

Unit 6 (1200 m)

This unit closely resembles Unit 4. Rhyolite ignimbrites, most strongly eutaxitic and many rich in clasts of rhyolite lava, prevail in the lower part of the unit. They contain horizons of conglomerate, lithic sandstone and laminated desiccated mudstone. On Doghead Peninsula, north of Western Channel, the uppermost ignimbrite is capped by 10 m of stromatolitic dolomite. Above this marker, ignimbrites south of Western Channel grade northward into a sequence of thin-bedded silicified mudstone, resembling ribbon chert, with thick graded beds of lithic sandstone. At the top, these sediments contain stubby flows (?) of massive rhyolite lava.

Unit 7 (10 000 + m)

This is the first of three units composed almost entirely of ignimbrite and is best exposed across the 30-km-wide northeast-dipping homocline along the east side of MacAlpine Channel. In contrast to the rhyolite ignimbrites below, this unit is remarkably free of sediments. The ignimbrites are dacites and rhyodacites, composed of broken phenocrysts of plagioclase and hornblende with variable but subsidiary proportions of alkali feldspar and quartz dispersed in a red-brown matrix of densely welded ashstone. Much of the ignimbrite is massive but weak eutaxitic foliation defined by strongly flattened and recrystallized porphyritic pumice fragments is not uncommon. Sills and irregular intrusions of plagioclase-hornblende porphyry compositionally identical with the ignimbrites but with unbroken phenocrysts occur in the upper part of the unit.

Unit 8 (4500 + m)

The ignimbrites of this unit range in composition from rhyolite to rhyodacite and most have subequal proportions of phenocrystic plagioclase, alkali feldspar and quartz. They are highly variable in colour, have generally better developed eutaxitic structure and more abundant lithic clasts than the more basic ignimbrites above and below. There is a large dome, 4 km in diameter, of flow-banded rhyolite lava exposed 15 km east of the mouth of Sloan River. The dome was extruded in a shallow depression in the ignimbrite terrain marked by laminated mudstone. Around the dome are aprons of rhyolite-pebble conglomerate.

Unit 9 (11 000 m)

This unit, best exposed between Junius and Jaciar lakes, consists of dark green, massive to moderately eutaxitic, dacite ignimbrites. The dacites are separated by discontinuous and much thinner horizons of brick-red, strongly eutaxitic, phenocryst-poor, rhyolite ignimbrite. All of the ignimbrites are densely welded and extensively recrystallized. They are intruded by irregular masses of fine plagioclase-biotite porphyry, intruded in turn by coarse plagioclase-hornblende porphyry with variable but subordinate phenocrysts of alkali feldspar and quartz. This unit, like Unit 7, is remarkably free of sediment.

Unit 10 (12 000 + m)

This unit resembles those exposed on Great Bear Lake in having mafic lavas and abundant sediments in addition to ignimbrites. The ignimbrites range from rhyolite to rhyodacite and the lower ones are more strongly eutaxitic. They are interrupted by lenses of lithic sandstone and conglomerate, and by four mappable sub-units (*see* Hoffman and Bell, 1975, Fig. 5) of dark laminated silty mudstone. Capping the mudstones are basalt flows, some coarsely plagioclase-porphyritic, and local rhyolite domes. The lavas pinch out westward and are fed by basalt and rhyolite dykes that cut the basement rocks beyond the erosional limit of the volcanics. The mudstones are intruded by distinctive porphyry laccoliths that contain rounded alkali feldspar megacrysts up to 5 cm in diameter, in addition to smaller euhedra of plagioclase, quartz and biotite. Broken feldspar megacrysts of identical type occur in some of the ignimbrites proving that the porphyries are their intrusive equivalents.

The basalts are thickest along the eastern edge of the volcanics, where they are intercalated with conglomerates containing clasts of basement rocks. This line is interpreted as a buttress unconformity, controlled by west-side-down movement during volcanism along a major synvolcanic fault system that truncates the older volcanic rocks at depth and marks the eastern limit of all but the smallest plutons of the Great Bear Batholith.

Plutonic Rocks

The plutons of the Great Bear Batholith are epizonal. Most are massive, and all have sharp contacts and very narrow contact metamorphic aureoles. There is, however, an anomolous area of variable but locally intense tectonic foliation that involves both plutonic and volcanic rocks between Handley-Page and Gagne lakes. The pluton between St. Germain Lake and Coppermine River is locally foliated with numerous inclusions because the present erosion surface coincides with its near horizontal roof.

The plutonic rocks are subdivided into five classes, the age relations of which are indicated in the legend of Figure 76.1. This is not meant to imply that, for example, all biotite granites are coeval, but only that they tend to have consistent cross-cutting relations with the other plutons with which they are in contact. Thus, granites intrude granodiorites and quartz monzonites but are cut by diorite and quartz diorite stocks. Granites on Great Bear Lake, however, may be older than granites, or even granodiorites, along the eastern margin of the batholith.

Reference has already been made to the demonstrably synvolcanic plutons but those of the monzonite-syenodiorite-syenite suite deserve further comment. The most interesting of these occurs south of Rainy Lake (Fig. 76.2) and is considered to be responsible for the vein mineralization at Terra Mine (Shegelski and Scott, 1975). This is a gravity differentiated sill-like body that was intruded into the upper part of Unit 2 before or during the deposition of Unit 3. Its lower part is a syenodiorite comprised of densely-packed cumulate plagioclase euhedra and clotted hornblende prisms with interstitial alkali feldspar. The upper part is a syenite, cut by aplite dykes, containing sparse plagioclase in a matrix, becoming progressively finer grained upward, of alkali feldspar with minor interstitial quartz. There is generally a narrow chilled border phase of plagioclase-porphyritic monzonite, above which are satellitic intrusions of fine grained hornblende syenodiorite. Other intrusions of this suite, including those at Balachey Lake (Fig. 76.2) and Echo Bay (Fig. 76.1, Unit A), are much less well differentiated and are best described as low-quartz hornblende monzonites.

It was noted (Hoffman and Bell, 1975) that the vein mineralization at both Echo Bay and Terra Mines occurs near the top of Unit 2, suggesting a stratigraphic control. The Norex and Federated Mines at the east end of Rainy Lake, however, occur in Unit 3, and the silver veins of the old Contact Lake Mines south of Echo Bay have a plutonic host, now known to belong to the monzonite-syenodiorite-syenite suite. The fact that all six mines in the region occur either within these intrusions or within the zones of intense alteration adjacent to them supports Shegelski and Scott's conclusion and provides an obvious target for exploration.

References

Fraser, J.A., Hoffman, P.F., Irvine T.N., and Mursky, G.
1972: The Bear Province; *in* Variations in Tectonic Styles in Canada, edited by R.A. Price and R.J.W. Douglas; Geol. Assoc. Can., Spec. Paper 11, p. 454-503.

Hoffman, P.F.
1973: Evolution of an early Proterozoic continental margin: the Coronation geosyncline and associated aulacogens of the northwestern Canadian Shield; *in* A Discussion on the Evolution of the Precambrian Crust, edited by J. Sutton and B.F. Windley; Phil. Trans. Roy. Soc. London, v. 273, p. 547-581.

Hoffman, P.F. and Bell, I.
1975: Volcanism and plutonism, Sloan River map-area (86K), Great Bear Lake, District of Mackenzie; *in* Report of Activities, Part A, Geol. Surv. Can., Paper 75-1A, p. 331-337.

Kidd, D.F.
1933: Great Bear Lake area, Northwest Territories; Geol. Surv. Can., Summ. Rep., 1932, pt. C, p. 1-36.

McGlynn, J.C.
1976: Geology of the Calder River (86 F) and Leith Peninsula (86 E) map-areas, District of Mackenzie; *in* Report of Activites, Part A, Geol. Surv. Can., Paper 76-1A, Rep. 77.

Robinson, H.S.
1933: Notes on the Echo Bay District, Great Bear Lake, Northwest Territories; Can. Mining Met. Bull., v. 26, p. 609-628.

Shegelski, R.J. and Scott, S.D.
1975: Geology and mineralogy of the silver-uranium-arsenide veins of the Camsell River District, Great Bear Lake, N.W.T.; *in* Abstracts with Programs, Geol. Soc. Am., v. 7, no. 6, p. 857-858.

77. GEOLOGY OF THE CALDER RIVER (86 F) AND LEITH PENINSULA (86 E) MAP-AREAS, DISTRICT OF MACKENZIE

Project 730041

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Mapping of the Calder River map-area was completed during the past summer, along with those portions of Leith Peninsula map-area to the west, that are underlain by Precambrian rocks. In the Leith Peninsula area the supracrustal rocks in the northern part of the map-area were mapped by Hoffman (Hoffman, 1976). The map-area spans two major tectonic divisions of the Wopmay Belt (Stockwell, 1970) which has been called the Wopmay Orogen by Fraser *et al.* (1972). These tectonic divisions are the Great Bear Batholith and the Hepburn Metamorphic-Plutonic Belt, and their mutual boundary is defined by the Wopmay Fault (Fraser *et al.*, 1972).

In the Calder River area the essential elements of the geology have been described in previous reports (McGlynn, 1974, 1975). During the past summer unmapped areas of supracrustal rocks and the granitic rocks in the north half of the sheet were mapped. The oldest rocks occur just east of the Wopmay Fault and comprise a conformable sequence of siltstone and shales overlain by basic volcanic rocks. This sequence thins toward the north and disappears into the Hepburn gneisses just south of the north boundary of the map-area. Thickness of the siltstone-shale unit is unknown as it is cut by granitic rocks. The top of the volcanic unit is not exposed, but only a few hundred feet are present in the sequence, whereas in the south and central parts of the area some thousands of feet are exposed. A thin (up to 10 feet) but persistent dolomite unit occurs between the volcanic and siltstone-shale unit at the north end of the belt. This dolomite is not present in the sequence in the central part of the belt, but is found in the south part of the map-area. These rocks, to the east, become involved in migmatite phases of the Hepburn gneisses or are cut by massive granodiorites and quartz monzonites that also intrude the Hepburn gneisses.

Work in the Hepburn gneisses east of the Wopmay Fault more or less confirms earlier work (McGlynn, 1974). The rocks are complexly folded. Early folds whose axial planes appear to trend northeasterly to east are refolded about northerly or west of north-trending axes. Mylonite zones, so common in these gneisses, occur along faults that breach second folds or along limbs of second folds. The gneisses consist of metasedimentary rocks, migmatites, veined gneisses, and granitic gneisses. They are cut by several plutons of massive granodiorite to quartz monzonite that contain porphyritic phases.

West of the Wopmay Fault, in the Great Bear Batholith, most of the supracrustal rocks had previously been mapped, and the past summer's work confirmed previously reported results. A number of large masses of intrusive porphyry were outlined in the central part of the north half of the Calder River map-area. These rocks, which may be quartz latite or rhyolite porphyries, cut the supracrustal rocks and granitic rocks.

Of special interest is the geology at Hottah Lake in the Leith Peninsula map-area (Fig. 77.1) where there is a sequence of basic volcanic and sedimentary rocks, whose thickness is about 8000 feet, and which rest unconformably on a basement underlain by granitic rocks. These rocks are unlike any of the supracrustal rocks in the Calder River map-area.

The oldest unit, a lithic sandstone and conglomeratic sandstone (unit 2a, Fig. 77.1) rests unconformably on basement (unit 1) composed of medium grained, massive, buff-weathering grey to buff granodiorite containing plagioclase, quartz, potash feldspar and chloritic mafic minerals that probably included biotite and possibly, hornblende. Sediments at the base of the sequence are poorly sorted and contain blocks and clasts of granodiorite and lenses of conglomerate with granodiorite clasts and quartz pebbles. Upwards from the base, the rocks are less feldspathic and lithic, and finer grained. Higher in the sequence are beds up to 4 feet thick of fining-upward conglomerates that grade to sandstone and locally, shale. This unit is about 700-800 feet thick. The sandstones are overlain conformably by a sequence of purple to grey massive mudstones (2b) that contain thin quartz-pebble conglomerate horizons, iron-formation beds, and a thin possibly discontinuous unit of massive basic lava. The unit is as much as 1000 feet thick. It is overlain by a sequence of massive basalts (3a) that are about 2000 feet thick, and that may include some intrusive phases. A thin unit of pillowed basalt occurs within the upper part of this unit. The massive lavas are overlain by a thick sequence of pillowed basalts (3b) with local thin chert beds. The pillowed sequence attains a thickness of about 2000 feet and comprises flows with pillows that vary in maximum dimensions from 6 inches to 8 feet. Smaller pillows tend to be round, and larger ones, round to slightly elongate. Pillows are amygdaloidal near their margins and at the centre of most are round gas cavities or vugs that are lined with quartz and epidote. Their dimensions vary with the size of pillow, being about one foot in the largest pillows. Quartz masses are common in pillow interstices. Phases of both massive and pillow basalts are glomeroporphyritic, with clusters of light greyish green equant plagioclase grains measuring about one half inch in diameter.

The pillowed basalts are overlain by a relatively thin unit of massive basaltic flows (unit 4) which are overlain by gritty siliceous sandstones (unit 5) fining-upward to grey siliceous mudstones and chert. Only about 200 feet of these sediments are found on shoreline exposures on islands.

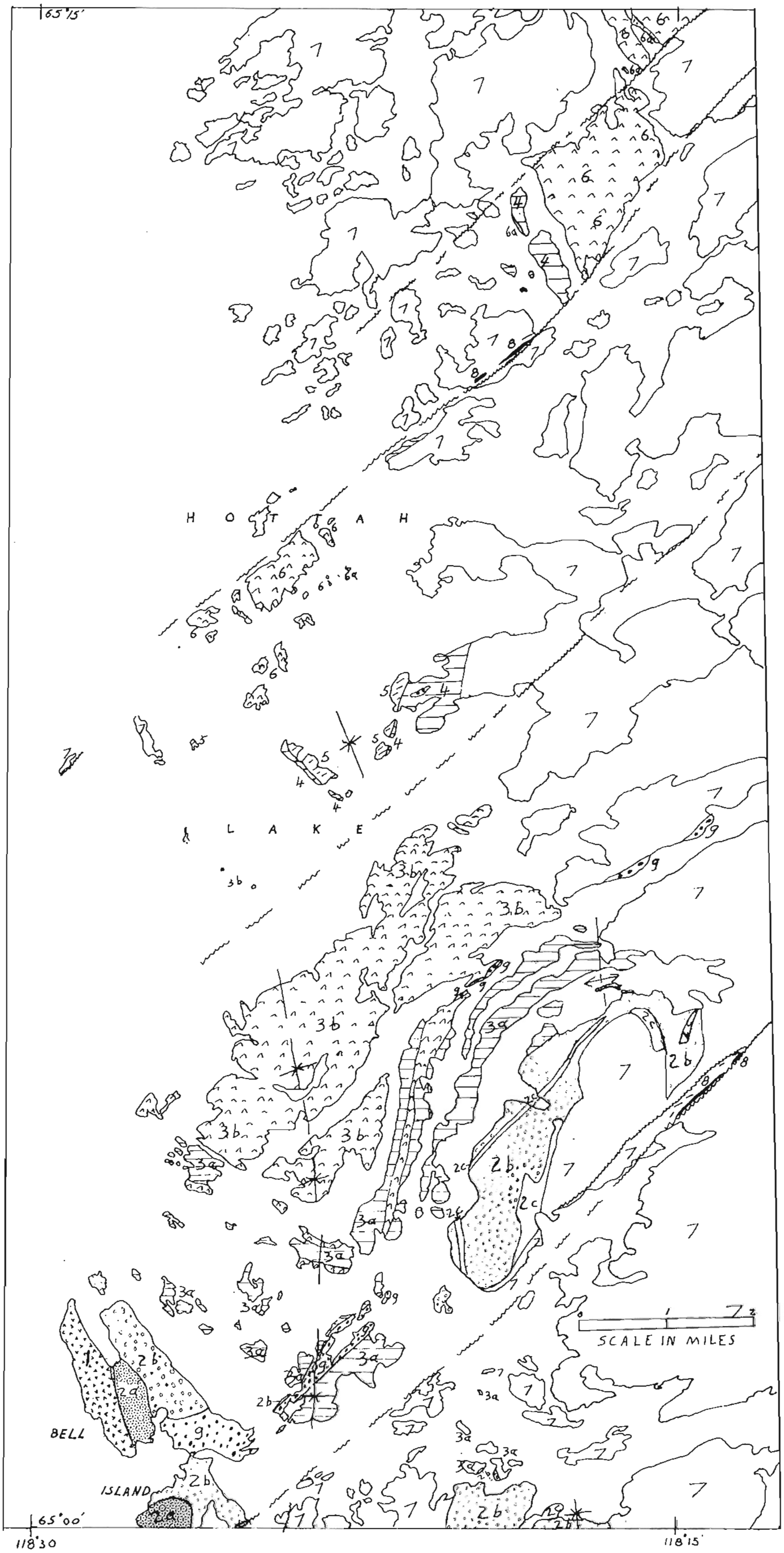

Figure 77. 1.

Geological sketch map of part of Hottah Lake area, Leith Peninsula map-area.

1. Granodiorite.
2a. Lithic sandstone, conglomeratic sandstone.
2b. Mudstone, sandstone, quartz pebble conglomerate, in part intensely silicified.
2c. Basalt flows.
3a. Massive basalt flows, possibly with some interrelated gabbro sills.
3b. Pillowed basalt.
4. Massive basalt flows.
5. Quartzite, chert, mudstone.
6. Pillowed basalt with some interbedded massive flows.
7. Granodiorite, quartz monzonite, granite.
8. Quartz stockwork.
9. Diabase.

This unit is succeeded by another thick sequence of pillowed basalts (unit 6) with an exposed thickness of about 1500 feet. These pillowed lavas are similar in all respects to those described above. The top of the sequence is cut by granitic rocks. Several thin units of interbedded fine grained quartzites and grey mudstones (6a) occur in the upper part of the pillow lavas.

The sequence is folded about north-trending axes. In the central part of the area a syncline and anticline occur in which the basalts are simply folded. The sedimentary rocks in the core of the anticline are more intensely deformed, with minor folds occurring on the limb of the major fold. These sedimentary rocks are also intensely silicified, with mudstones converted to chert-like rocks.

The supracrustal rocks are cut by granodiorites and quartz monzonites typical of the Bear Batholith. The volcanic rocks are metamorphosed to amphibolites in narrow zones about the younger granitic rocks. Away from the younger intrusives the degree of metamorphism is as low as greenschist facies.

All of these rocks are cut by four major northwest-striking right lateral faults with displacements of two to three miles. Quartz stockworks (unit 8) occur along the faults.

Volcanic and sedimentary rocks south of the most southerly of these faults are grouped with the succession north of the fault because of similar lithology and stratigraphy, but their precise stratigraphic relationship to the sequence north of the fault is unknown. It is possible that further work south of the map-area may show that they are an unrelated succession.

The supracrustal sequence at Hottah is unlike anything found in the Calder River map-area and its stratigraphic relationship to rocks in other parts of the Bear Batholith is uncertain. They may relate to the lowest part of the succession in the northern part of the Leith Peninsula map-area around Conjuror Bay (described by Hoffmann, 1976 as unit 1), in which quartzites and mudstones are overlain by basalts with interbeds of sedimentary rocks.

The age of the basement of the sequence at Hottah Lake is unknown at present, but may be resolved by further work that will include age determinations from samples collected for this purpose.

References

Fraser, J.A., Hoffman, P.F., Irvine, T.N., and Mursky, G.
1972: The Bear Province; *in* Variations in tectonic styles in Canada, edited by R.A. Price and R.J.W. Douglas, Geol. Assoc. Can., Spec. Paper 11.

Hoffman, P.H., Bell, I.R., and Tirral, R.
1976: Sloan River map-area, Great Bear Lake, District of Mackenzie; *in* Report of Activities, Part A, Geol. Surv. Can., Paper 76-1A, rept. 76.

McGlynn, J.C.
1974: Geology of the Calder River map area (86F), District of Mackenzie; *in* Report of Activities, Part A, Geol. Surv. Can., Paper 74-1A, p. 383-385.

1975: Geology of the Calder River map-area (86F), District of Mackenzie; *in* Report of Activities, Part A, Geol. Surv. Can., Paper 75-1A, p. 339-341.

Stockwell, C.H.
1970: Geology of the Canadian Shield; *in* Geology and economic minerals of Canada, edited by R.J.W. Douglas, Geol. Surv. Can., Econ. Geol. Rept. no. 1, p. 44-54.

THE BACK RIVER VOLCANIC COMPLEX, DISTRICT OF MACKENZIE

Project 740019

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Regional and Economic Geology Division

Introduction

The 1975 field season was the first in a two-year project to map, at a scale of 1: 25 000, an intermediate to felsic volcanic complex in the Slave Province. The complex, centred on 65 degrees north latitude and 108 degrees east longitude, lies about 300 miles (485 km) northeast of Yellowknife, N.W.T. It occupies a crudely triangular area between the Back River on the east and the Contwoyto River on the west (Fig. 78.1).

The purpose of this project is to map in detail a large Archean, felsic volcanic complex and determine: (1) the stratigraphy and internal structure, and their relationship to the surrounding granitic and sedimentary rocks, and to mineral occurrences of possible economic interest; (2) criteria for identification and interpretation of metamorphosed felsic volcanics in the Slave Province; and (3) the mode and environment of volcanic eruption. This area was chosen because of its relatively low grade of metamorphism (greenschist to lower amphibolite facies) and apparent low degree of deformation as suggested by the investigations of Baragar (1975) and Henderson (1975).

Barnes and Lord (1954), Fraser (1964), Wright (1967) and Tremblay (1971) reconnoitered parts of the area during regional mapping projects. Frith and Hill (1975) mapped the northern margin of the complex as well as other volcanic belts to the north and west. Baragar and Henderson made detailed investigations in selected areas of the complex.

A. Miller, J. Ostler, R.M. Easton and J. Lafleur provided superb assistance during both geological mapping and operation of field camps. Ostler mapped part of the area along the Back River as part of an M.Sc. thesis at Carleton University.

Stratigraphy

The volcanic rocks are mainly andesitic to rhyolitic breccias, lavas and tuffs {nomenclature of volcanic rocks is based on field identification and specific gravity determinations (Lambert, in press)}. They are divided into five formations (Units 2 to 6 in Figs. 78.1 and 78.2) with considerable overlap in composition. The stratigraphic succession varies in different parts of the complex. At the present stage of mapping, a stratigraphic succession representative of the complex has not been established. A schematic representation of facies relations in the northern parts of the complex, however, is shown in Figure 78.2.

Previous workers have assigned the volcanic and sedimentary rocks of this complex to the Yellowknife Supergroup of Archean age. Zircons from greywacke along the eastern margin of the complex yield a preliminary $^{207}Pb/^{206}Pb$ age of 2670 m.y. (R.K. Wanless, pers. comm.).

Unit 1. Sediments of the Yellowknife Supergroup surround the complex and form a prominent embayment in the volcanics southwest of Regan Lake. In this embayment, the rocks are dominantly dark grey shale, slate and argillite, with minor greywacke. These rocks become lighter in colour and coarser grained toward the contact between sediments and volcanics. Coarse- to fine-grained greywackes are the major constituent toward the eastern side of the complex and along the Back River.

Iron-formation outcrops along the shoreline and north of Wolverine Lake, for at least 20 km along the Back River, and intermittently along the southwestern margin of the area mapped. The variations from oxide to sulphide facies have been outlined by Baragar (1975) and Henderson (1975). The iron-formation is associated always with black carbonaceous slates and siltstones and commonly with water laid tuffs, tuff breccia, and well-bedded volcanic clastic sandstones. Ooliths have been reported from some of these rocks near the Back River. All rocks contain pyrite and locally, pyrrhotite. Near Wolverine Lake, brown weathering, carbonate cemented andesite tuff or tuff-breccia lies between the iron-formation and overlying units of coarse volcanic breccia.

The contact between the sediments and volcanics is conformable. In the northern part of the volcanic complex, sediments generally lie beneath the volcanics or are interfingered with them, whereas locally, along the Back River the sediments overlie volcanic rocks. In view of the general lack of marker horizons and abrupt facies changes within the volcanic succession, the iron-formation may be a very important time stratigraphic horizon.

Unit 2. Dacite tuffs, breccias and massive units are most abundant in the central parts of the complex. They represent volcanic domes and associated lavas and ash-flows. These rocks, which weather grey, white, pink and green, contain sparse feldspar and quartz phenocrysts in an aphanitic matrix. Some pink and white varieties may be rhyolite. Ash-flow tuff units north of Wolverine Lake contain dark green fiamme of pumice in a medium green or grey matrix. These rocks have conchoidal fracture and preserve delicate textures such as cuspate shards and pumice, and perlitic cracks.

Breccias consist of dacite blocks (up to 25 cm) in a matrix of dacitic grit that weathers grey to pale brown. Units are generally unsorted and unlayered except for local conglomeratic horizons.

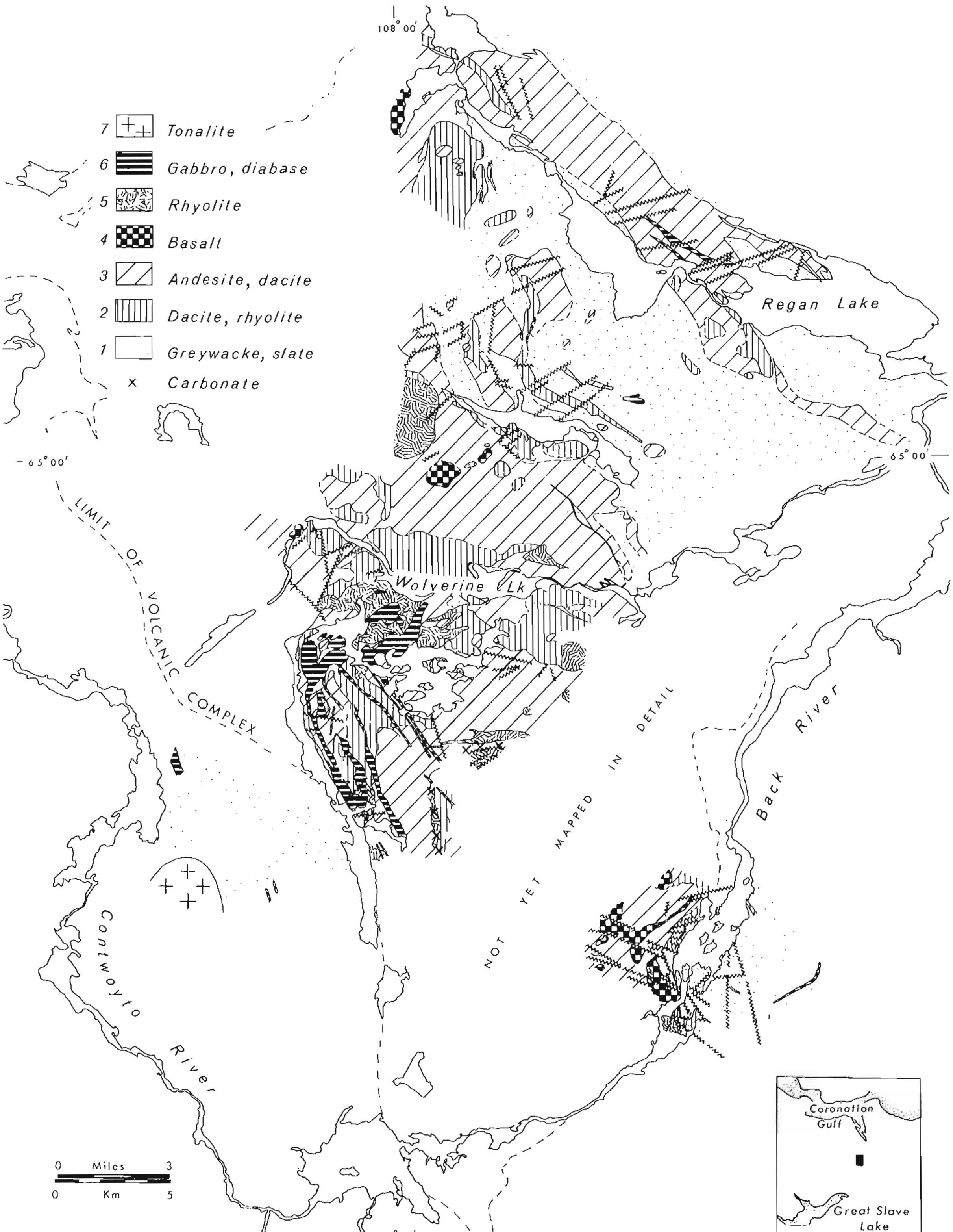

Figure 78. 1. Geology of parts of the Back River volcanic complex, Slave Province, N. W. T.

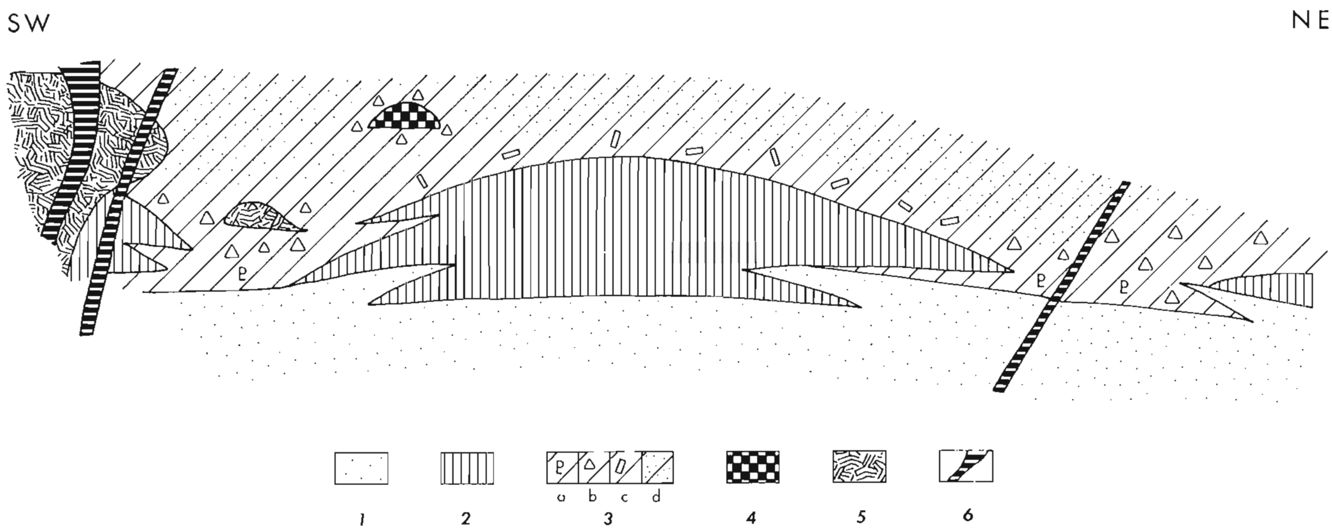

Figure 78.2. Facies relations in the northern parts of the Back River volcanic complex: 1) greywacke, mudstone; 2) dacite, rhyolite; 3) andesite, dacite -- (a) pillow lava, (b) breccia, (c) feldspar porphyry, (d) pyroclastic rocks; 4) basalt; 5) rhyolite; 6) gabbro.

Unit 3. Andesite forms a diverse unit comprising massive porphyries, pillow lavas, pillow breccias, tuffs, breccias and conglomerates. Andesite breccias, pillow lavas and massive units form a northwest-southeasterly trending belt about 2.5 km wide and 20 km long northwest of Regan Lake. Pillow lavas are most common in the west-central and northern parts of this belt. Breccias are composed of angular to subrounded porphyritic andesitic blocks, in a green andesitic matrix of fine breccia and grit. Constituents are unsorted and the rocks generally are not layered.

Coarse breccias along the southwestern side of the map-area are interpreted as landslide rubble. They are polymictic breccias and conglomerates containing blocks, up to 30 cm across, of andesite, dacite and, in places, sedimentary rocks. Units are massive, unsorted and unlayered.

In the northwestern corner of the area mapped, and in some areas south of Wolverine Lake, andesites are massive porphyries and columnar jointed lavas. These rocks contain up to 25 per cent plagioclase phenocrysts, with smaller amounts of amphibole phenocrysts, in an aphanitic, green to grey matrix. Some of the rocks, with specific gravities in the range 2.7 to 2.76, probably are between andesite and dacite in composition.

Ash flow tuffs compose most andesitic areas north and south of Wolverine Lake. They are thick, massive units with no layering (except for local concentrations of coarse blocks) and poor sorting. They are fine grained, grey-green and unwelded to partly welded. Densely welded tuffs, and air fall tuffs, are rare. Quartz and plagioclase crystal tuffs and lapilli tuffs contain clasts of porphyritic andesite, vesicular pumice, and massive to eutaxitic tuff. In many areas delicate pumice and shard structures are visible in hand specimen.

Unit 4. Basaltic rocks occur in isolated areas in the northwestern part of the map-area, north of Wolverine Lake and near the Back River. They are dark green pillow lavas and breccias.

Unit 5. Rhyolite domes, flows, and dykes occupy a large circular area northwest of Wolverine Lake and form several crude, circular to irregular areas, south of Wolverine Lake. These bodies are flow brecciated, flow layered and massive. Some are composite bodies having cores of massive quartz phyric rhyolite, and margins of quartz and feldspar phyric dacite. Some rhyolitic units that have a sucrose texture, may be hypabyssal intrusive equivalents of the brecciated flows and domes.

South of Wolverine Lake, white to brown weathering carbonate occurs along arcuate zones, up to 50 m wide, of intense shearing and faulting in rhyolite. Generally the carbonate has filled fractures. In at least one locality, however, it appears to be a bedded unit.

Unit 6. A swarm of metagabbro dykes intrudes all formations along the western side of the complex. These rocks weather dark grey to dark green and form massive outcrops with blocky fracture. Locally they have columnar jointing perpendicular to dyke contacts. Fine grained diabase at the contact of the dykes grades into medium- and coarse-grained gabbro with distinct ophitic texture, near the centre. Locally, gabbros are quartz-bearing and contain mauve-grey quartz eyes to 2 mm.

Structure

The volcanic rocks exhibit deformation patterns that differ from those of the adjacent sedimentary rocks. Thus it cannot be assumed that the structures in the volcanics are similar to those in the sediments. Folding in the volcanic rocks is very difficult to establish because of the general lack of marker horizons, and

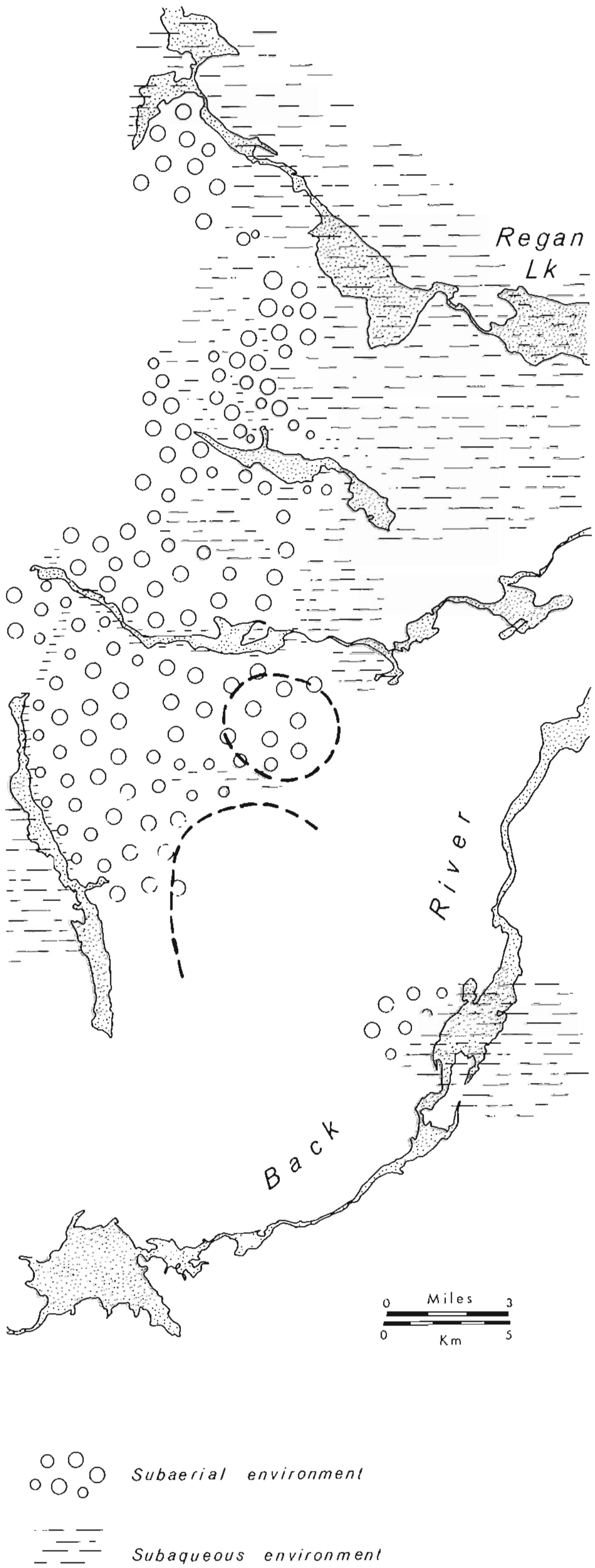

Figure 78.3. Depositional environments in parts of the Back River complex. Heavy dashed line denotes position of circular structures.

the massive, non-layered character of the units. The outcrop pattern (defined by the contact between volcanics and sediments in the northern parts of the area) and the general lack of steep dips in the volcanic rocks, suggests that they are deformed into gently plunging, open folds. In contrast, sediments adjacent to or intercalated with the volcanic rocks are tightly folded. Penetrative structures in sediments are not evident in adjacent volcanics.

A first phase of deformation formed upright concentric folds with horizontal axes trending east-west in the iron-formation and related sediments exposed in a gorge at the east end of Wolverine Lake, whereas a second phase formed broad open warps with axes plunging northerly. Near the Back River, axial traces of a first phase of folding have variable trends that swing abruptly to conform to the contact between the sediments and the volcanics. These folds generally become tighter toward the contact. A second phase of folding has north to northwesterly-trending axes.

Faults define prominent linear and arcuate topographic features that, in the northern part of the area, trend northeasterly or north to northwesterly. The intensity of brecciation and shearing in massive rocks generally increases towards these lineaments.

A zone of intense shearing is subparallel to the northern contact of the belt northwest of Regan Lake. This zone contains abundant tectonic breccias. Pillows and clasts are almost obliterated or have an apparent flattening and steeply-plunging elongation within planes of shearing.

South of Wolverine Lake, a prominent circular feature marked by a series of arcuate lakes and stream valleys is bounded on the northern and western sides by steeply dipping fault zones. Large rhyolite dykes and zones of carbonate follow this fault system. Arcuate features such as this are present in other parts of the area and may be ring fractures related to cauldron subsidence features. Further work is required to substantiate this hypothesis.

Interpretation

Subaqueous environments are indicated by pillow lavas, pillow breccias, coarse breccias and intercalated sediments, rocks interpreted as subaqueous andesite breccias (northwest of Regan Lake), and greywackes and mudstones of the Yellowknife sediments.

The iron-formation with its associated oolitic horizons, and its close association with subaerial tuffs and breccias is indicative of shallow marine deposition that probably marks the transgression from subaqueous to subaerial environments. Welded tuffs that are thick massive units, andesite flows having columnar joints, and rhyolitic to dacitic bodies that are flow layered, flow brecciated and massive, are interpreted as subaerial tuffs, flows and exogenous domes. Some coarse polymictic breccias are interpreted as landslide rubble of subaerial origin. A schematic distribution of subaerial and subaqueous environments in the parts of the volcanic complex mapped in detail are shown in Figure 78.3. Prominent circular features defined by fault zones, rhyolitic intrusions and arcuate drainage patterns, occur in subaerial environments, and may be related to caldera structures.

References

Barnes, F.Q. and Lord, C.S.
1954: Aylmer Lake, District of Mackenzie, Northwest Territories; Geol. Surv. Can., Map 1031A.

Baragar, W.R.A.
1975: Miscellaneous data from volcanic belts at Yellowknife, Wolverine Lake and James River, N.W.T.; *in* Report of Activities, Part A, Geol. Surv. Can., Paper 75-1A, p. 281-286.

Fraser, J.A.
1964: Geological notes on northern District of Mackenzie, Northwest Territories; Geol. Surv. Can., Paper 63-40.

Frith, R.A. and Hill, J.D.
1975: The geology of the Hackett-Back Rivers Greenstone Belt – Preliminary report; *in* Report of Activities, Part C, Geol. Surv. Can., Paper 75-1C, p. 367-370.

Henderson, John B.
1975: Sedimentological studies of the Yellowknife Supergroup in the Slave Structural Province; *in* Report of Activities, Part A, Geol. Surv. Can., Paper 75-1A, p. 325-330.

Lambert, M.B.
1975: Anatomy of a greenstone belt – Slave Province, N.W.T.; *in* Volcanic Regimes in Canada, Geol. Assoc. Can., Special Paper (in press).

Tremblay, L.P.
1971: Geology of Beechey Lake map-area, District of Mackenzie; Geol. Surv. Can., Mem. 365.

Wright, G.M.
1967: Geology of the southeastern barren grounds, parts of the Districts of Mackenzie and Keewatin; Geol. Surv. Can., Mem. 350.

79. GEOLOGY OF THE KILOHIGOK BASIN, GOULBURN GROUP, BATHURST INLET, DISTRICT OF MACKENZIE

Project 740092

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Introduction

Mapping and stratigraphic studies of Goulburn sediments within map-areas 76L, 76K, 76J, 76W, 76O were completed this year (*see* Fig. 79. 1 and Table 79. 1). Stratigraphic and sedimentological data were collected from the Buchans Bay area (77A), Beechey Lake area (76G), Contwoyto Lake area (76E) and Rockinghorse Lake area (86H).

On the basis of these data the regional characteristics of the Kilohigok Basin are defined. In general, the sedimentary history of the area is similar in all parts of the basin, illustrated by the remarkable continuity of most formations, members and even beds. However, the units show distinct east to west facies changes and some distinctive north to south facies changes, affecting mainly the sediments of the Western River and Brown Sound formations sediments. The facies changes and other data indicate that the preserved area of Goulburn sediments formed the northeastern part of a much larger basin.

A general summary of the stratigraphic and sedimentological features and a tentative genetic appraisal are given, by subdividing the stratigraphy into four major periods of sedimentation: deltaic, clastic, carbonate and red clastic (Fig. 79. 2).

Stratigraphic additions and revisions resulting from further field work are discussed. One major discovery was the recognition of cross-cutting pipes and dykes of lithic breccia similar to the 'exotic' breccias of Reinhardt (1972) in the East Arm of Great Slave Lake; these are briefly discussed.

F. H. A. Campbell and M. P. Cecile are jointly responsible for mapping and collection of data within the Kilohigok Basin. F. H. A. Campbell is principally responsible for the detailed descriptions of the Western River to Quadyuk formations, and M. P. Cecile is principally responsible for the detailed descriptions and definitions of the units above the Quadyuk Formation, up to and including the Ellice Formation; these units are the subject of his Ph. D. dissertation.

Kilohigok Basin

The Kilohigok Basin (Fig. 79. 3) was a large, extensive (> 7000 km^2) intracratonic basin situated to the east of the correlative Coronation Geosyncline. The Goulburn Group (Fig. 79. 3) was deposited within the north, northeast and northwest parts of this basin. Paleocurrents and facies changes from Goulburn clastic sediments define a major basin paleoslope to the west. Carbonate phase sediments are generally not influenced

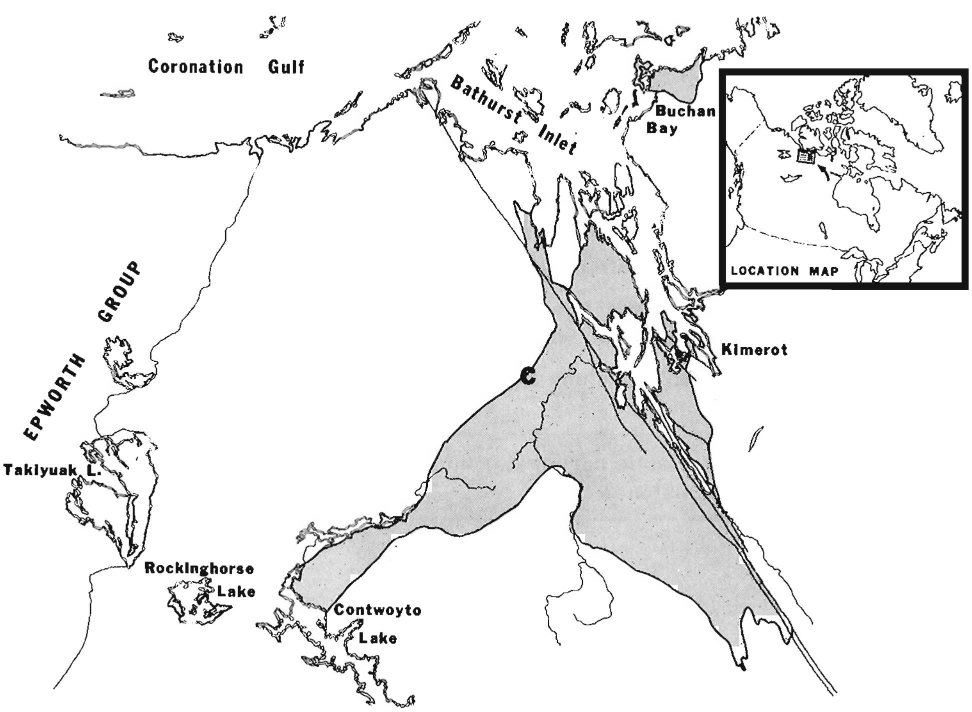

Figure 79. 1

Distribution of the rocks of the Goulburn Group, within the known limits of the Kilohigok Basin (stippled). Location names are referred to in the text. The large C refers to the location described in the economic geology portion of the report.

[1]Carleton University.

Table 79.1

Stratigraphy of the Goulburn Group

ANGULAR UNCONFORMITY		
AMAGOK FM. (N→S; >800->1000 m)	A	- white to mauve coarse grained moderately indurated arkose; minor conglomerate.
BRECCIAS		- pipe and dyke breccia complexes, younger than $\underline{B}_{3C}$; possibly younger than Amagok and overlying angular unconformity.
BROWN SOUND FM. (N→S; 2300-800 m)	$\underline{B}_{3C}$	- red, well-indurated arkose successions interstratified with white and mauve coarse grained moderately indurated arkose successions (N→S; 850-0 m).
	$\underline{B}_{3B}$	- thin vesicular basalt flows interstratified with red arkose (N→S; 50-0 m).
	$\underline{B}_{3, 3A}$	- red, medium to fine grained well indurated arkose (N→S; 400-250 m).
	$\underline{B}_{2}$	- ferruginous, calcareous muddy siltstone and very fine sandstone (N→S; 350-250 m).
Omingmaktook Member	$\underline{B}_{1O}$	- chaotic intraformational breccia and slumped exotic carbonate blocks (between 70 and 200 m N and S).
Omingmaktook Member	$\underline{B}_{1S}$	- buff-brown medium to coarse grained arkose (N→S; 60-0 m).
Omingmaktook Member	$\underline{B}_{1}$	- ferruginous, calcareous mudstone; salt casts abundant at the base (200 m).
KUUVIK FM. (300 m)	K_4	- stromatolitic carbonate, clastic carbonate, mudstone; edgewise conglomerate abundant, oncoliths.
	K_3	- stromatolitic carbonate, clastic carbonate; intraformational conglomerate abundant, minor mudstone.
	K_2	- very thick beds of interstratified carbonate and red and green mudstone (> 50% carbonate).
	K_1	- thin bedded carbonate-mudstone rhythmites (> 50% carbonate)
PEACOCK HILLS FM. (P_1→P_3; 300 m)	P_5	- red and green mudstone and siltstone with minor carbonate (Contwoyto Lake equivalent to P_1-P_3; 30 m).
	P_4	- thin bedded red and green mudstone-carbonate rhythmites with carbonate concretions (eastern facies equivalent to P_1-P_3; 70-150 m).
	P_3	- thin bedded mudstone-carbonate rhythmites; minor concretionary mudstone.
	P_2	- thin bedded green, red and red-brown mudstone rhythmites; massive, thick bedded siltstones with rare concretion or lens of carbonate
	P_1	- thin bedded red and green mudstone rhythmites; minor concretionary mudstone and carbonate beds.
QUADYUK FM. (0-50 m)	Q	- stromatolitic carbonate; clastic carbonate; minor calcareous quartzite; intraformational breccia; minor mudstone.

	B_S	- pisolitic quartzose dolomite; granular hematite ironstone; minor red or pink quartzite (30 m maximum).
BURNSIDE RIVER FM. (W→E; 200→>2500 m)	B_1	- red siltstone; fine grained subarkose; minor red mudstone and shale; minor rare stromatolitic carbonate (N→S; 100-500 m).
	B	- pink, white, red quartzite and minor subarkose; quartz-pebble conglomerate; rare shaly or muddy partings (> 500 m).
	B_M	- red mudstone, minor dolomite (0-25 m, in south only).
	B_D	- arenaceous dolomite; doloarenite; minor stromatolitic carbonate; dolomitic quartz-pebble conglomerate (W-E; 100-0 m).
WESTERN RIVER FM. (100→2200 m)		
Upper Argillite Member (W_4)		
	4B	- grey, buff, red siltstone and mudstone; minor quartzite.
	4A	- red mudstone and argillite.
Quartzite Member (W_3)		
	3D	- fine grained red and pink quartzite and rare arkose.
	3C	- red mudstone and argillite; minor turbidites (A-E beds).
	3B	- stromatolitic carbonate; doloarenite and dolosiltite; minor fine grained white or pink quartzite (BEECHEY PLATFORM).
	3A	- grey-green coarse grained turbidites (A-E beds).
Red Siltstone Member (W_2)		
	2E	- red siltstone and argillite; minor quartzite.
	2D	- quartzite; minor dolomite and mudstone.
	2C	- red, grey concretionary mudstone.
	2B	- grey concretionary mudstone; minor dolomite.
	2A	- white, pink protoquartzite; minor quartzose carbonate.
Lower Argillite Member (W_1)		
	1D	- stromatolitic carbonate; clastic carbonate; minor quartzite and calcareous quartzite (KIMEROT PLATFORM).
	1C	- thin bedded limestone, dolomite and siltstone.
	1B	interbedded siltstone, quartzite, thick bedded quartzose turbidites (A-E beds); minor subarkose.
	1A	- quartzite, argillite, quartz-pebble conglomerate; minor carbonate; regolith.
ANGULAR UNCONFORMITY		
ARCHEAN		

by this paleoslope, and during regressive phases prograde from the northern basin margin towards the south. Early in the depositional history of the basin a prominent basement 'high' north of the basin centre produced marked facies and thickness changes along a generalized north to south axis without affecting paleocurrent patterns of the coarse grained clastic sediments.

Goulburn sedimentation was dominated by terrigenous clastics. Simple stratigraph successions (very few cycles) are widely correlative. Clastic deposition was punctuated by hiatuses marked by the spread of stromatolitic carbonate successions that are locally capped with sedimentary soil profiles (pisolites, iron concretions, etc.).

The sediments of the Goulburn Group are interpreted to be the result of deposition within a large stable intracratonic sea. This basin had a simple and long-lived morphology that underwent subsidence accompanied by basin filling. The total thickness of Goulburn sediments, 7 to 8000 m, demands that some fundamental tectonic feature of crustal dimension must be involved in its origin and subsequent history.

Goulburn Sediments - A Brief Genetic Summary

Goulburn sediments comprise a succession of seven formations and numerous members (Table 79.1). These sediments are grouped into four major depositional phases of the basin's history (*see* Fig. 79.2); an early deltaic phase (1), clastic phase (2), a major carbonate phase (3), a period of red clastic sedimentation (4).

Deltaic Phase

Following initial transgression of the gneissic Archean basement, with accompanying deposition of quartzite and quartz-pebble conglomerate of the Western River Formation (W_{1A}) a major stromatolite platform - the Kimerot Platform (W_{1D}) - developed east of Bathurst Inlet (Fig. 79.1).

The Kimerot Platform stromatolites locally directly overlie a thin dolomite with molar tooth structure. The stromatolites appear to be transitional from the molar tooth structure into small mounds (Fig. 79.4). A second, overlying, stromatolite platform, the Beechey Platform (W_{3B}), was identified in the Beechey Lake area, and is correlative with an identical stromatolite succession east of Bathurst Inlet. The Beechey Platform (first tentatively identified by Tremblay, 1971) spread much farther to the west than did the initial, Kimerot Platform, and its basinal equivalents have been mapped around the entire present-day rim of the basin, from Arctic Sound in the north to Rockinghorse Lake in the west. Stromatolite facies in the Beechey Platform are correlative across the Bathurst Trench, and indicate a displacement on the fault in the "trench" of at least 40 miles. A return to deltaic deposition resulted in the burial of the Beechey Platform, and fine grained clastics spread across the basin. Continued regression resulted in the spread of coarser fluviatile sediments, the Burnside River Formation, marking the end of Western River Formation sedimentation.

The four members of the Western River Formation, first identified by Tremblay (1971), extend across the basin, indicating continued stable conditions, punctuated by brief periods of uplift.

Clastic-Fluvial Phase

Following deposition of the dominantly deltaic and stromatolitic Western River Formation, continued uplift in the source area produced the cyclic fluviatile quartzites of the Burnside River Formation which prograded across the basin, from an easterly and south-easterly source area. The majority of the Burnside is dominated by cyclic alternation of trough-crossbedded fine- to coarse-grained quartzite and quartz-pebble conglomerate as lateral accretion deposits (Duff *et al.*, 1967). The complete ideal cycle (essentially a fining-upward cycle) is rarely present, but is illustrated in Fig. 79.5, and described in Table 79.2.

A major, but brief, transgression during deposition of the formation resulted in the deposition of a terrigenous clastic dolomite member in the deeper parts of the basin (B_D member), but fluvial deposition was renewed and eventually spread across the entire basin as far west as Rockinghorse Lake (Fig. 79.1). Gradual uplift in the west-central part of the basin, together with subsidence in the east, resulted in the deposition of the red siltstone member (B_1). This member is weakly radioactive, and it will be discussed further under economic geology. Regression continued, with the eventual development of a pisolitic and/or granular hematite paleosoil across much of the basin (B_S member).

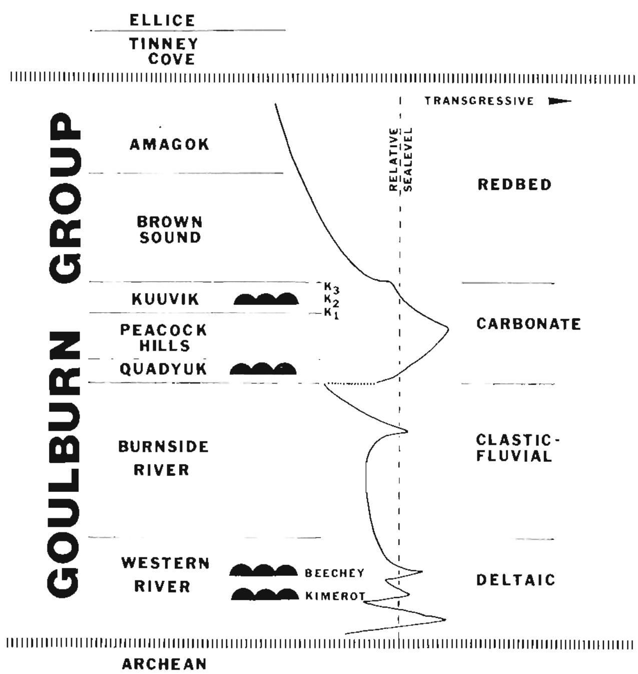

Figure 79.2. Diagrammatic summary of the genetic interpretation and stratigraphy of the Goulburn Group.

Table 79.2

Lithologic description of the sub-units in the ideal complete cycle in the Burnside River Formation, illustrated in Figure 79.5

E	-	Massive mottled red mudstone or red laminated shale; commonly occurs as a thin "film" on top of D sub-unit; occurs as abundant clasts in the overlying beds. Rarely preserved.
D	-	Parallel-laminated fine- to medium-grained reddish purple quartzite, occasionally with deep purple ferruginous quartzite interbeds; rarely with small-scale slumped trough crossbeds; unit may or may not be present.
C_2	-	Medium- to coarse grained, small-scale trough crossbedded pink to white quartzite; may have isolated pebbles up to 2.5 cm; almost always present.
C_1	-	Medium- to coarse-grained, large-scale trough-crossbedded pink to white quartzite or grit with abundant pebbles up to 15 cm, commonly at the bases of the crossbeds, but also scattered throughout the sub-unit; commonly has fragments and chips of E mudstones; almost always present.
B	-	Massive, structureless to weakly parallel-bedded, coarse grained grey to white quartzite, with rare isolated pebbles up to 15 cm; may have large-scale, trough-crossbeds developed near the upper contact; very rarely present.
A	-	Conglomerate with scoured base, in channels with maximum relief to 2.5 m; clasts of sub-unit E mudstone (clasts 5-20 cm) common in finer conglomerates, but absent in coarser varieties; majority of clasts derived from underlying succession, dominated by vein quartz, red, pink; white quartzite (C_1, C_2, D sub-units), and rare, possibly Archean basement-derived, clasts; maximum clast size observed is 2.5 m (pink quartzite); sub-unit is relatively common, but thickness is extremely variable, from 10 cm to 4 m.

Paleocurrents from over 6000 trough-crossbeds in the Burnside River Formation show almost unimodal transport direction to the west (± 30°) throughout the basin. The thickness of the formation east of Bathurst Inlet from measured sections indicates that the actual thickness is less than 2500 m. However, accurate thicknesses are not possible due to faulting and absence of outcrop.

In the northernmost part of the basin, in the Buchans Bay area, a succession of coarse grained, trough-crossbedded, pink and white quartzite rests directly on the Archean basement, and is lithostratigraphically correlative with the Burnside River Formation, as originally defined by Fraser (1969). Paleocurrents from trough-crossbeds indicate that the transport direction in this area was identical to that in the Burnside and Western River formations.

Carbonate Phase

Following a period of quiescence and regional soil formation (Burnside pisolite) the sea slowly transgressed, but without an accompanying influx of clastic sediments (Fig. 79.2).

Sediments deposited during this transgression were a thin complex of stromatolitic carbonates containing digitate deep-water mounds, with a very local platform-shoaled facies in the east, (Quadyuk Formation) followed upwards in the stratigraphic succession by their equivalent basin facies, the mudstone-carbonate turbidites of the first member of the Peacock Hills. The initial transgression was succeeded by a minor renewal of clastic sedimentation bringing in mudstone turbidites and massive, graded siltstone units of the second member of the Peacock Hills Formation. This was in turn followed by a regression and deposition of the upper Peacock Hills and Kuuvik formations, representing in stratigraphic succession, basin turbidites (P_3), proximal basin turbidites (K_1), transition from basin to carbonate bank (K_2) and a carbonate bank complex (K_3) comprised of subtidal stromatolite mounds giving way upwards to intertidal stromatolites.

Red Clastic Phase

Kuuvik sediments were succeeded by the deposition of coarsening-upwards ferruginous, calcareous clastic sediments (Brown Sound to Amagok formations). This succession, which attains thicknesses in excess of 1800 m in the south and in excess of 3000 m in the north, is composed of a lower red calcareous mudstone, discontinuous thin sandstone, and a slump breccia ($\underline{B}_1$, $\underline{B}_1$S, $\underline{B}_{10}$), a middle red muddy siltstone ($\underline{B}_2$), and an upper arkose succession with minor basalts ($\underline{B}_{3A-C}$), all overlain by coarser white-mauve arkoses of the Amagok Formation. Sediments of the red clastic phase had south to southwestward directed paleocurrents and distinct north-south facies changes. Facies changes and the stratigraphy of the red clastic sediments are illustrated in Figure 79.6. A tentative interpretation of this complex is that it formed by the progradation of a distant deltaic complex into a restricted marine basin.

Several units of interest occur with the red clastic sediments. The most prominent is an olistostrome ($\underline{B}_{10}$). This unit, somewhere between 70 and 200 m thick, is an horizon of a slump breccia derived from a once-conformable carbonate-mudstone succession that occupied the same stratigraphic position. Detailed mapping during the 1975 field season revealed that the olistostrome is cut and intruded locally by younger cross-cutting pipe and dyke breccia complexes. These breccias are discussed briefly in the next section of this report. Other anomalous units of the red clastic phase include thin vesicular basalts within the $\underline{B}_2$ member (Fig. 79.6) and a conglomerate with silicic volcanic pebbles within the Amagok Formation. This conglomerate indicates that the probable source of red clastic phase sediments was an extension of or an equivalent to the Great Bear Batholith, as originally

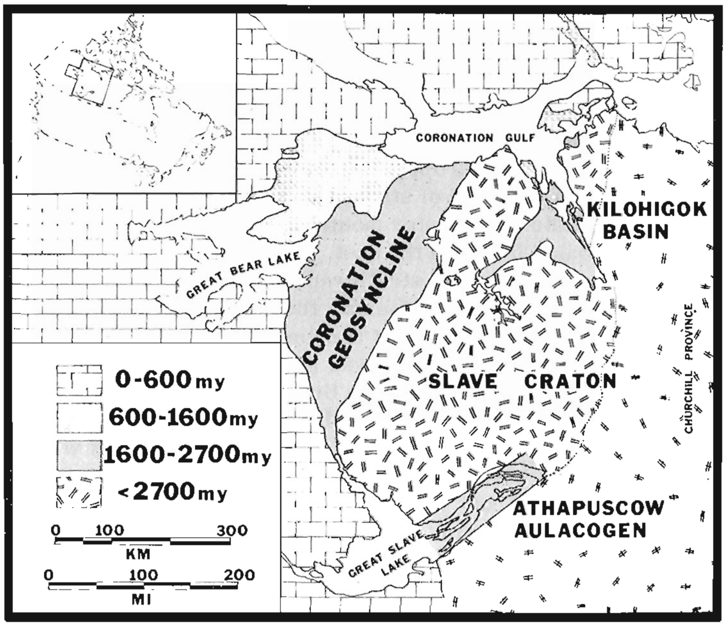

Figure 79. 3

Regional setting of the Kilohigok Basin, relative to the other elements of the Coronation Geosyncline.

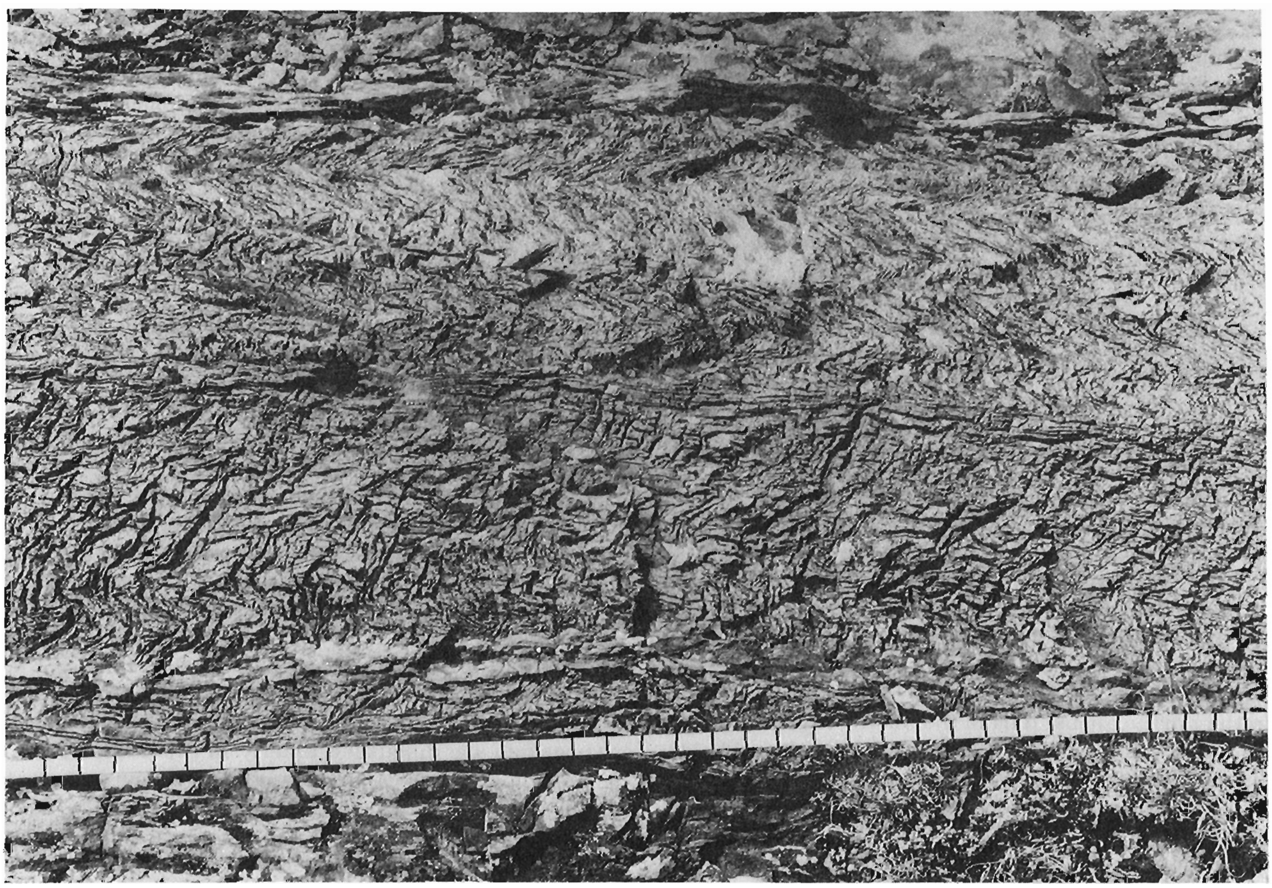

Figure 79. 4. Molar Tooth structure developed beneath a stromatolite mound in the Kimerot Platform Succession in the Western River Formation. Scale in inches (G. S. C. 202917-R).

proposed as a provenance for the correlative red molasse sediments of the Coronation Geosyncline (Hoffman, 1973).

Pipe and Dyke Breccia Complexes

Detailed examination of breccias forming large 'fissure-like' bodies below the olistostrome (previously described as 'channels', Campbell and Cecile, 1975; Table 1) has led to the recognition of the 'intrusive' nature of these bodies. Additional mapping revealed that as well as 'intruding' below the olistostrome, the breccias 'intrude' above and into the olistostrome (Table 79.1). Expansion of these breccia complexes at the stratigraphic level of the olistostrome resulted in the incorporation of a considerable volume of slump breccia material into these younger breccia complexes. The two, however, have vastly different origins.

The pipe and dyke breccia complexes form cross-cutting cylindrical and dyke-like bodies that intrude above, below, and into the olistostrome. The pipe and dyke breccias are characterized by a 'flow-banded' lithic breccia with dolomite megacrysts that everywhere rims and dips steeply along the contacts of the dykes and pipes. The dyke and pipe breccias 'intrude' and brecciate other horizons such as the Burnside River Formation (Fig. 79.7). The olistostrome, on the other hand, is a sedimentary slump breccia with 'slump blocks' and crude stratification that predates the formation of the pipe and dyke breccias by at least a few thousand metres of sediment ($\underline{B}_2$, $\underline{B}_3$).

Figure 79.7 is a detailed map of the smallest of the pipe and dyke complexes, illustrating some of the features of these breccias. The northeastern part of this intrusive consists of several circular structures which coalesce to form an 'intrusive' plug infilled with chaotic and brecciated interstratified carbonates and mudstone. This plug is fringed by a small volume of 'flow-banded' breccia that dips steeply everywhere along the contact. The western part of the complex is a megabreccia dyke of Kuuvik carbonate blocks fringing a 'block' of Burnside River Formation. The 'block' of Burnside quartzite is believed to have been originally in fault (strike-slip) contact with the Peacock Hills or Kuuvik formations prior to 'intrusion' of the breccia complex. The contact with the quartzite and the breccia complex is an irregular breccia of quartzite fragments grading into the more typical carbonate-mudstone breccias within the dyke itself.

At the present time it is thought that these pipes and dyke breccia complexes are diatremes or similar intrusive bodies; detailed examination of the 'flow-banded' breccias in thin section and other studies are necessary before positive conclusions are reached.

It is interesting to note the economic importance of these breccias. Similar breccias intruding units correlative with the Goulburn Group have been identified in the East Arm of Great Slave Lake (Reinhardt, 1972). According to Reinhardt, the uranium mineralization on one of the Simpson Islands is associated with the breccias.

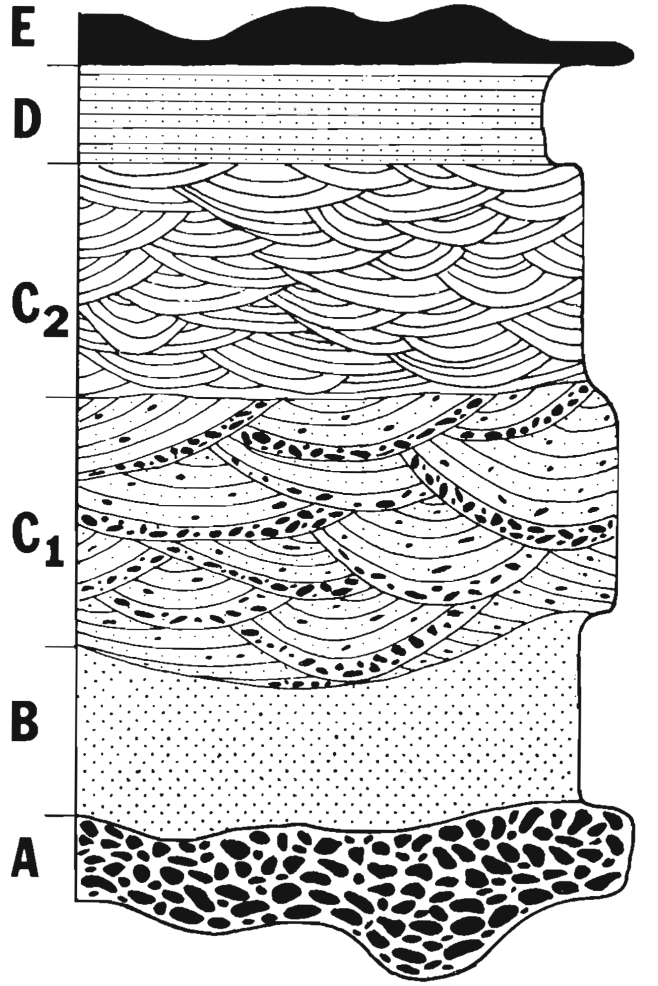

Figure 79.5. 'Ideal Complete Cycle' in the Burnside River Formation (B Member). Descriptions of the A to E sub-units are given in Table 79.2. Not to scale.

Stratigraphic Additions and Revisions

Western River Formation

The members of the formation outlined in Campbell and Cecile (1974) have been grouped into the members as initially outlined by Tremblay (1971). Locally the formations contain more sub-units than those first described, but Tremblay's original four members can be mapped everywhere in the basin.

Burnside River Formation

One major addition has been made to the member subdivisions of the formation, with the addition of the pisolitic carbonate (B_S) at the top of the unit.

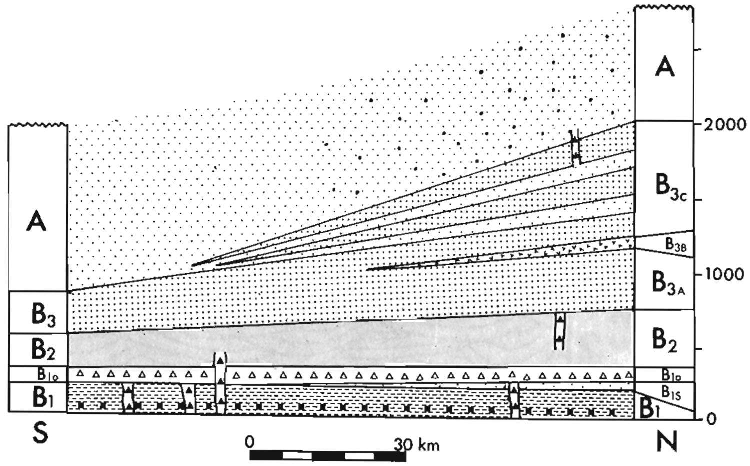

Figure 79. 6

Cross-section of the Brown Sound and Amagok Formations showing north-south facies changes across the preserved area of the Kilohigok Basin (for legend refer to Table 79. 1).

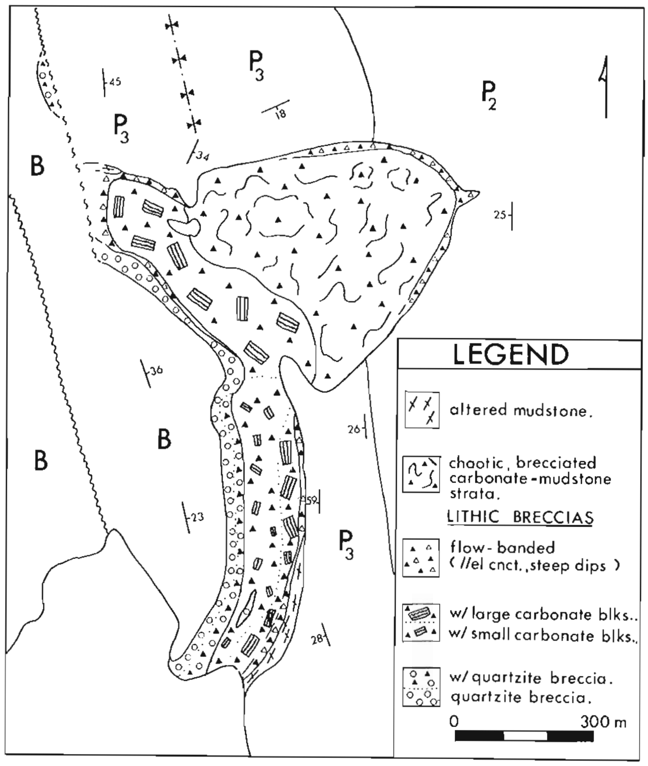

Figure 79. 7. Detailed map of a breccia complex; the northeastern part is a complex of pipes in a large cylindrical body, the western part is a dyke of megabreccia (for additional legend refer to Table 79. 1).

Peacock Hills Formation

One change has been made to the member subdivisions of the Peacock Hills Formation. Member P_2 (Campbell and Cecile, 1975), a concretionary red mudstone, has been included in the lower P_1 member of this formation after it proved unmappable, and with the recognition of similar lithologies elsewhere in the stratigraphy. Overlying members have also been relabelled to suit this change (*see* Table 79. 1).

Brown Sound Formation

The subdivisions of the Brown Sound Formation were expanded to accommodate facies changes in the northern part of the area, as illustrated in Table 79. 1 and Figure 74. 6. In addition, pipe and dyke breccia 'intrusive' complexes were recognized cutting the succession (*see* preceding section on pipe and dyke breccias). Detailed examination of the 'regolith developed upon diabase sill associated with the olistostrome' (Campbell and Cecile, 1975) revealed that the outcrops examined were part of very large blocks of altered material that are large 'clasts' within the olistostrome. This makes the diabase older than the olistostrome, but it is doubtful if the alteration of these blocks represents a paleosoil.

Economic Geology

Three areas of economic interest were noted during the course of mapping. Two of these occur in the lower part of the Goulburn Group, while the third unit is confined to the upper Goulburn Group. These are:

1. Disseminated chalcocite which occurs in fractures and small veinlets in the basinal stromatolite facies of the Beechey Platform, along the northwestern margin of the basin, in the area between the Booth River and the contact with an extensive diabase sill (Fig. 79. 1). The mineralization occurs between two north-northwest-trending diabase dykes, which are apparently younger than the diabase sill. Mineralization in the form of disseminated chalcocite is present for at least 300 feet along strike, across a 6-foot zone, concentrated in the stromatolite zone and associated sedimentary sequence.

2. The B_1 member at the top of the Burnside River Formation contains unidentified radioactive minerals disseminated in the mudstones and siltstones. These locally are up to 20 times the background defined in the other formations and members. The radioactivity within the member occurs throughout the basin. The highest scintillometer readings were obtained south of Kuuvik Lake, in the central part of the basin, where

the unit is approximately 300 metres thick. The readings are considerably lower in the east, but are still significantly above the background. The mineralization is concentrated in beds, which give noticeably higher scintillometer readings than the remainder of the formation. Ruzicka (pers. comm., 1975) determined that the unit contained approximately 30 ppm uranium, but that the majority of the radioactivity is caused by potassium.

3. Pipe and dyke breccia complexes that are mostly confined to the Brown Sound Formation are of potential economic interest since similar breccias in the East Arm of Great Slave Lake are associated with known uranium mineralization there (Reinhardt, 1972).

References

Campbell, F.H.A. and Cecile, M.P.
1975: Report on the Geology of the Kilohigok Basin, Goulburn Group, Bathurst Inlet, N.W.T.; *in* Report of Activities, Part A, Geol. Surv. Can., Paper 75-1A, p. 297-306.

Duff, P. McL.D., Hallam, A., and Walton, E.K.
1967: Cyclic sedimentation; Developments in Sedimentology 10, Elsevier Publishing Co., New York, 280 p.

Fraser, J.A.
1964: Geological notes on the northeastern District of Mackenzie; Geol. Surv. Can., Paper 63-40, 20 p.

Hoffman, P.F.
1973: Evolution of an early Proterozoic Continental margin: the Coronation Geosyncline and associated aulacogens of the northwestern Canadian Shield; Roy. Soc. London, Phil. Trans., Ser. A, v. 273, p. 547-581.

Reinhardt, E.W.
1972: Occurrences of exotic breccias in the Petitot Islands and Wilson Island map-areas, East Arm of Great Slave Lake, District of Mackenzie; Geol. Surv. Can., Paper 72-25, 43 p.

Tremblay, L.P.
1971: Geology of the Beechey Lake map-area, District of Mackenzie — A part of the western Canadian Precambrian Shield; Geol. Surv. Can., Mem. 365, 56 p.

Project 740004

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Mapping during the 1975 field season, for publication on a scale of 1: 250 000 was concentrated in the east half of the map-area. The project is scheduled for completion in 1976. The area is a part of a region previously mapped on a scale of one inch to eight miles (Wright, 1967). The reconnaissance mapping outlines the broad distribution of the Helikian Dubawnt Group sedimentary and volcanic rocks and the underlying basement igneous and metamorphic rocks.

In the southeast quarter of the map-area the rocks have a pronounced northeast trend (Fig. 80. 1), similar to that in the adjoining map-areas to the south and east. Metavolcanic rocks in the extreme southeast corner are of basic composition for the most part, but contain lenses of intermediate to acid composition. Scattered small granodiorite to quartz monzonite plutons intrude the metavolcanics. To the northwest of the metavolcanic band is a band of granodiorite gneiss with some amphibolite, and then a wide band of migmatized paragneiss. The paragneiss is commonly garnetiferous and sillimanite was recognized in a few places. White pegmatitic leucosome bands are prominent in this rock. Mineral associations indicate middle and upper amphibolite facies of metamorphism in the paragneiss.

The paragneiss is bounded on the northwest by grey to dark grey biotite – hornblende granodiorite gneiss to quartz diorite gneiss containing abundant inclusions of amphibolite. This unit is distinctive on the aeromagnetic maps as the rocks have a much higher magnetic response than do those in adjoining map-units. A few small granodiorite plutons occur within this map-unit.

To the northwest of the inclusion-rich gneiss there is another band of migmatized paragneiss and it in turn is bounded to the northwest by granodiorite gneiss.

Just north of 62°30', near the south end of Yathkyed Lake, a previously unmapped band of basic metavolcanic rocks strikes approximately east-west. This band is apparently fault-bounded for the most part. The metabasalt is chiefly massive, but pillows are present here and there. Minor ultrabasic intrusions or zones occur within the metavolcanics.

A large pluton of granodiorite to the west of Yathkyed Lake is massive and homogeneous in its central part, but toward the margins is foliated and contains abundant mafic inclusions.

All the above described basement rocks are intruded to some degree by veins, stringers or irregular bodies of pink, medium grained to pegmatitic quartz monzonite.

Porphyritic quartz monzonite, present in a small pluton east of Tulemalu Lake, is pink, medium- to coarse-grained and contains prominent feldspar phenocrysts. Mafic content, estimated to be 6 per cent or less, is predominantly hornblende with minor biotite. Unlike the previously described rocks, this quartz monzonite is almost undeformed and is considered to be late Aphebian in age. It is overlain by Dubawnt volcanic rocks.

Gabbro to metagabbro dykes are abundant in the area, the majority trending between 85 and 115 degrees. Metamorphism of these dykes varies from slight to moderate. Earlier dykes, trending north to northeast, are more metamorphosed, and less abundant. A few fresh gabbro dykes of the northwest trending Mackenzie swarm are present.

Unmetamorphosed sedimentary and volcanic rocks of the Dubawnt Group occur north and west of Yathkyed Lake (Fig. 80. 1). A lower red bed sequence consists of a conglomeratic unit, overlain by arkosic sandstones and mudstones. An upper volcanic sequence consists of mixed flows and fragmental rocks of andesite and trachyte composition. The sequences are probably equivalent respectively to the South Channel Formation, the Kazan Formation and the Christopher Island Formation of Donaldson (1965). The sedimentary rocks are typically coloured deep maroon, purplish red or red brown. The pebbles and cobbles of the conglomerate consist of granodiorite, granodiorite gneiss, vein quartz, quartz monzonite, and amphibolite or basic metavolcanics in decreasing abundance. There is a wide range in shape and roundness of the clasts, from well-rounded spherical boulders to rectangular angular blocks. The matrix of the conglomerate is massive, medium- to fine-grained, and in some places it is somewhat calcareous. Bedding in the conglomerate is marked by intercalated lenses of sandstone and pebbly sandstone.

Pink to maroon coloured, fine- to medium-grained sandstone with distinct bedding forms the lower part of the Kazan Formation in this area. Where present, the upper part of the formation consists of deep maroon to reddish brown siltstone and mudstone. In the basin west of the Kazan River the upper part of the Kazan Formation is extensively developed, and some of the mudstone layers show well preserved mudcracks and mudstone or siltstone chips occur in sandy beds. Ripple-marks are abundant in some of the fine grained rocks. Overlying the well laminated mudstones and siltstones in this basin, is a thick sequence of massive fine grained red to maroon rocks with no visible bedding. These could possibly be tuffs.

The volcanic rocks of the Christopher Island Formation typically form prominent hills. Their thickness probably exceeds 100 m, but is variable as the formation was deposited on an irregular topography. No evidence was seen in this map-area of erosion of the lower sedimentary formations preceding the volcanism, as is noted by Donaldson (1965) in the type area. The lower part of the volcanic unit consists of dark green porphyritic lavas, probably andesitic composition, with

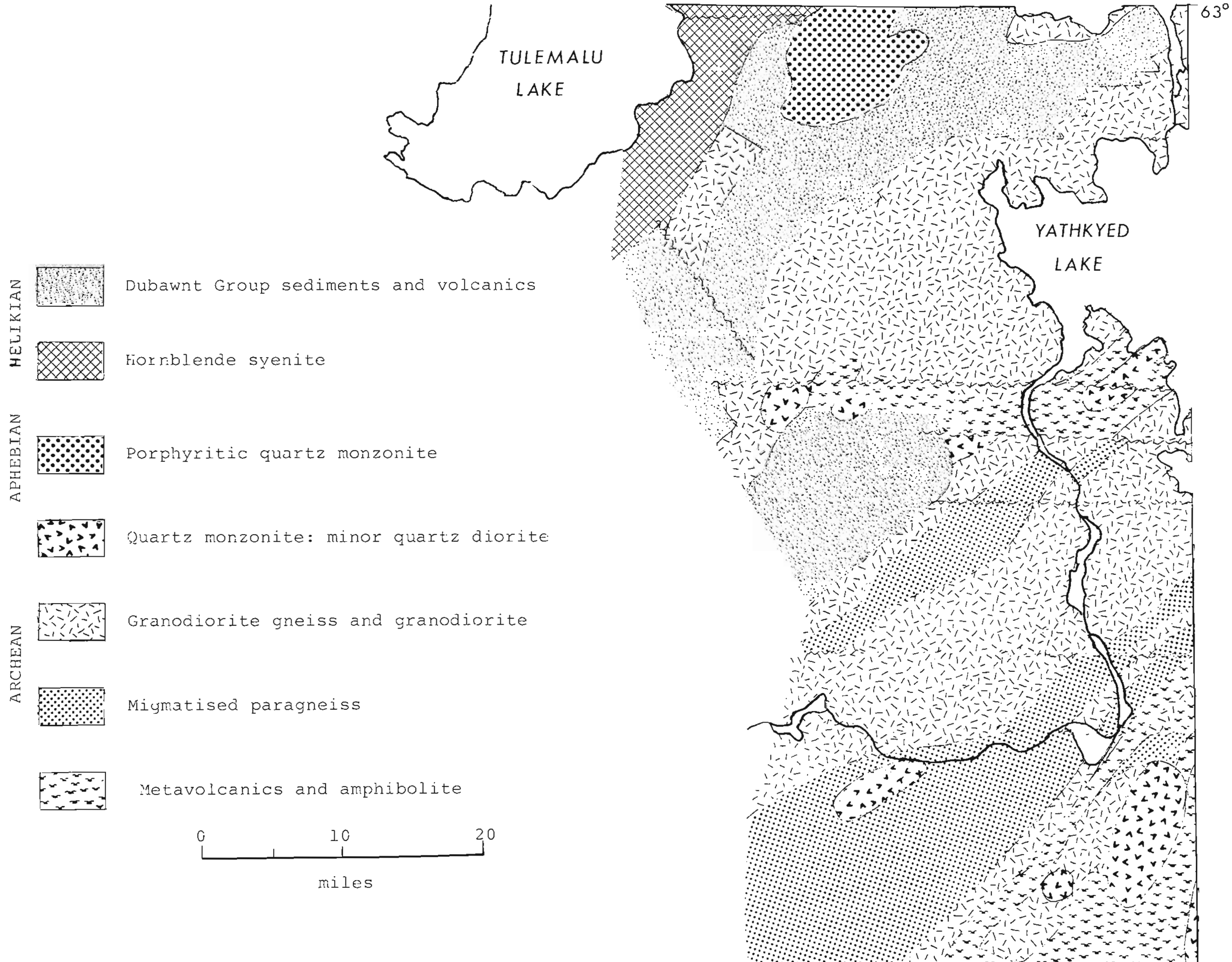

Figure 80. 1. Simplified geological sketch map of the east half of the Tulemalu Lake map-area.

biotite and feldspar phenocrysts, and less commonly, pyroxene phenocrysts. The dark green lavas are overlain by purplish red and red porphyritic lavas, agglomerates and tuffs, probably of trachytic composition. Some of the feldspar porphyry lavas in the upper part of the sequence may be equivalent to the Pitz Formation of Donaldson (1965) although quartz phenocrysts are rare or absent, unlike the type locality of the Pitz Formation.

A belt of medium grained, massive, pink hornblende syenite on the east side of Tulemalu Lake is probably correlative with the Martell syenite (Donaldson, 1965) but no age relationship with the Dubawnt Group rocks could be established.

In the basement rocks two periods of deformation are recognized. The earlier deformation results in northeast trending foliation, typically 045 to 060 degrees, except where earlier plutons result in interference of the regional pattern. The folding typically plunges gently to the northeast. During the late stages of the folding, the medium grained to pegmatitic quartz monzonite veins and stringers were emplaced, injecting generally parallel and subparallel to the northeast foliation, although locally they cross-cut the foliation. However, the veins themselves are flattened and show preferred mineral orientation, apparently due to the same deformation. Further, boudinaging of amphibolite layers during the deformation has resulted in infilling around boudins by the quartz monzonite.

The later deformation results in broad open flexures, with minor axial plane cleavage, striking approximately 150 degrees with variable but steep dips. Although both the quartz monzonite veins and stringers and the rocks they intrude are warped by these flexures, some shears associated with the flexures are filled by quartz monzonite veins, indicating that the quartz monzonite was still mobile during the second episode of folding, and that probably the second folding followed closely in time the dominant northeast folding. The abundant east-trending gabbro and metagabbro dykes postdate both deformations.

Folding in the Dubawnt Group rocks is limited to minor warps associated with faults that cut the rocks.

The map-area is characterized by prominent east-trending faults, eight having been recognized in the northern three-quarters of the area. Although their age is unknown, it is suggested that they are old and have been rejuvenated from time to time. Along with younger northwest trending faults, they cut Dubawnt Group rocks and in part control the outcrop distribution of the Dubawnt rocks. In the granodiorite west of Yathkyed Lake a major northeast trending zone of faulting and shearing results in a cataclastic zone at least five miles wide. It probably represents a major regional shear zone along which there has been periodic movement.

No occurrences of mineralization of potential economic interest were seen. Checks with scintillometers on the Dubawnt sedimentary and volcanic rocks indicated only background or slightly higher readings. Minor occurrences of quartz magnetite and chert pyrite iron-formation are associated with the basic metavolcanic rocks and in one place with the migmatized paragneiss. A few small occurrences of sulphide mineralization, pyrite with very minor chalcopyrite, are present in small shear zones in the metavolcanics.

References

Donaldson, J. A.
1965: The Dubawnt Group, Districts of Keewatin and Mackenzie; Geol. Surv. Can., Paper 64-20.

Wright, G. M.
1967: Geology of the southeastern barren grounds, parts of the District of Mackenzie and Keewatin; Geol. Surv. Can., Mem. 350.

MACQUOID LAKE (55 M WEST HALF) AND THIRTY MILE LAKE (65 P EAST HALF) MAP-AREAS, DISTRICT OF KEEWATIN

Project 750005

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Field work from June 5 to August 20, 1975 initiated a two-year project to map, at 1: 250 000 scale, the west half of MacQuoid Lake (55M) and the east half of Thirty Mile Lake (65P) map-areas. The area was previously examined during Operation Baker (Wright, 1955, 1967) and the Dubawnt Group rocks, which underlie the north part of the area, were subsequently mapped by Donaldson (1965). Present investigations are intended to upgrade knowledge of the Aphebian and Archean geology south of the Dubawnt Group, and to assess the mineral potential of pre-Dubawnt and Dubawnt Group rocks. Mapping in 1975 concentrated in the west half of 55M and immediately north and south of Thirty Mile Lake (Fig. 81. 1).

Large areas are underlain by granitoid gneisses of diverse composition, texture, and structural aspect. Locally mappable gneiss units and thin amphibolite and paragneiss bands within areas dominated by granitoid gneiss have been omitted from the geological sketch map (Fig. 81. 2) to emphasize the location of the major metasedimentary-metavolcanic belts and the distribution of plutonic bodies.

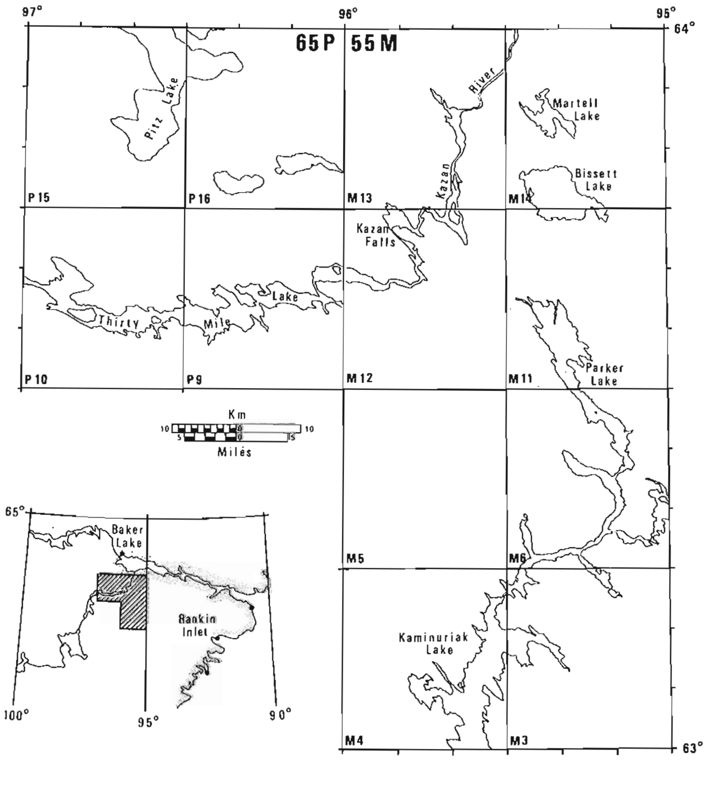

Figure 81. 1. Location of work in 1975.

Kaminak Group (?)

Two east-northeast trending belts of metavolcanic and metasedimentary rocks are tentatively correlated with the Kaminak Group (Davidson, 1970). The southern belt is continuous with the 'amphibolite-pelitic gneiss belt' mapped by Reinhardt and Chandler (1973). The belt extends across the south half of map-sheet 55M/6 and is characterized by two main groups of lithologies:

(1) fine grained, thin layered, amphibolite and hornblende-plagioclase gneiss (metamorphic equivalents of mafic to intermediate flows and tuffs),
(2) biotite-quartz-feldspar paragneiss, garnet-biotite schist, and laminated quartz-magnetite iron-formation (metamorphic equivalents of greywacke, pelitic sediments, and iron-formation.

Amphibolites are the dominant lithology in the north half of the belt. Thin bands of amphibolite are interbanded with thick metasedimentary units to the south. The steeply dipping east-northeast trending metasediments and metavolcanics envelop small elliptical and domal masses of gneissic granodiorite (Fig. 81. 2).

Garnet- and biotite-rich paragneiss and schist are interlayered with thin bands of oxide facies iron-formation (maximum thickness 7 m). Iron-formations are complexly folded and boudinaged but discontinuous bands have been traced along strike for up to two kilometres. Deformed and boudinaged bodies of white pegmatite are present throughout the belt. Garnet, muscovite, and tourmaline are common accessory minerals in the pegmatites.

Metamorphic mineral assemblages are indicative of the amphibolite facies. The common assemblage in quartzofeldspathic gneisses is quartz-microcline-biotite-muscovite-plagioclase. Associated pelitic assemblages contain garnet and rarely kyanite or cordierite. Hornblende-plagioclase and hornblende-plagioclase-garnet-biotite-quartz are typical assemblages in metavolcanic units. Iron-formation may contain two amphiboles and garnet, in addition to quartz and magnetite. Apatite and tourmaline are common accessory minerals.

A group of lower grade metavolcanic and metasedimentary rocks form an east-west trending belt across the north half of map-area 55M/11. The dominant lithology is a fine grained, dark green chloritic schist. Minor amphibole-bearing schist, amphibolite, oxide

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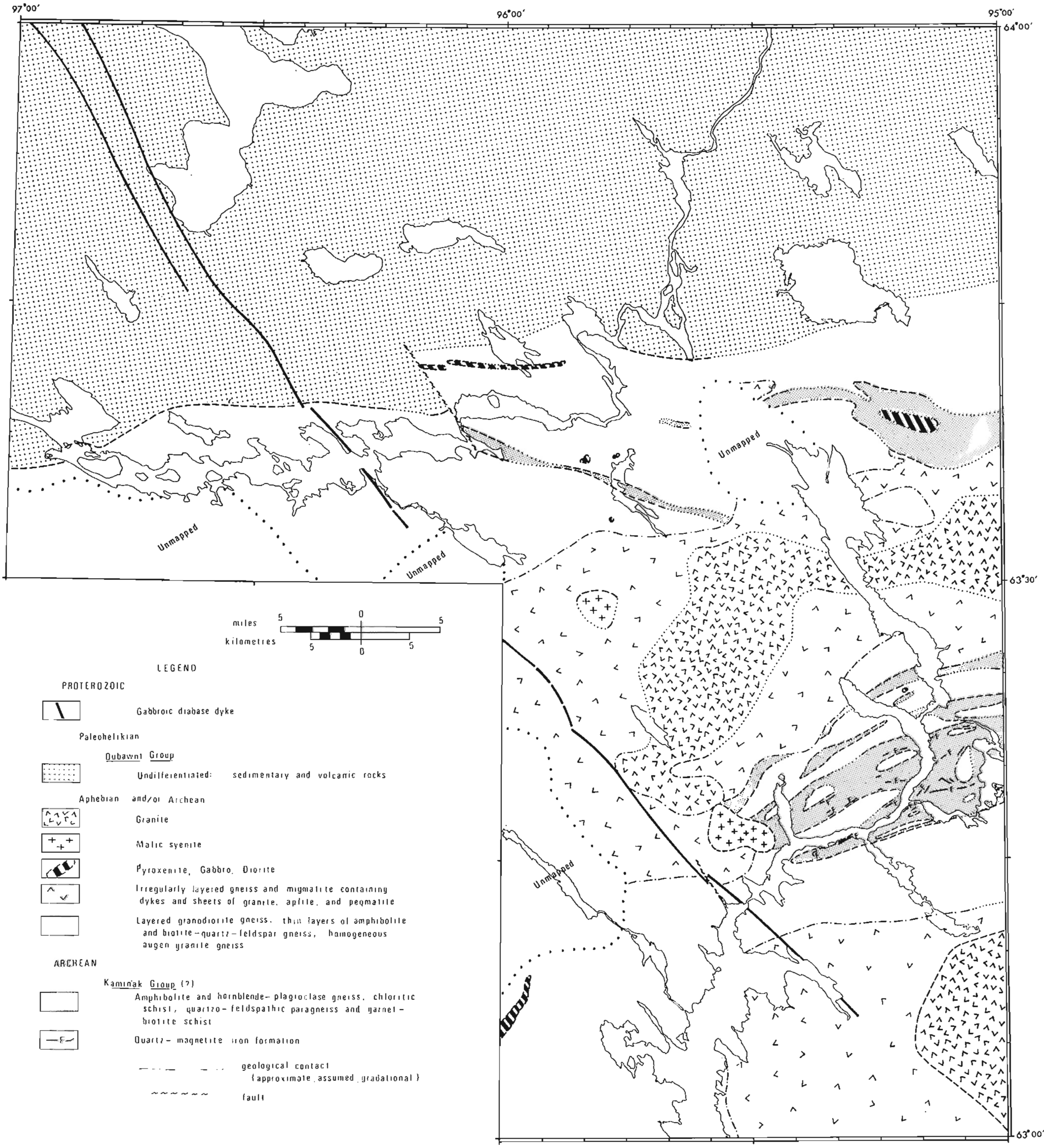

Figure 81.2. Geologic sketch map of MacQuoid Lake (55M west half) and Thirty Mile Lake (65P east half).

iron-formation, and paragneiss, are also included in the unit. These lithologies are common near the margins and within the western half of the belt.

Granitoid Gneisses

Layered grey granodiorite gneiss and gneissic granodiorite are intimately infolded with paragneisses and amphibolites of the Kaminak Group (?) in map-area 55M/6. North and south of the belt of Kaminak Group (?) rocks the layered gneisses become progressively more granitic in composition, with an accompanying decrease in the continuity of units within the gneiss complex. Large areas of map-areas 55M/3, 55M/5 and

55M/11 are underlain by migmatitic gneisses characterized by pinch and swell, irregular layering and granoblastic texture. Discontinuous thin metasedimentary bands, including garnet-sillimanite-bearing paragneiss and oxide iron-formation, are present in the southeast corner of map-area 55M/3. Numerous sheets and dykes of massive, fresh, pink aplite, granite, and pegmatite have intruded the gneisses, producing migmatitic complexes surrounding large plutons of fresh granite (Fig. 81.2). Foliations within the gneisses are broadly conformable with the plutonic contacts.

North of the granite body in map-area 55M/5 and 55M/11 the gneisses vary widely in composition and texture. Several bodies of homogeneous medium- to coarse-grained augen granite gneiss with 3- to 5-cm potash feldspar megacrysts form east-west trending bodies both north and south of the metavolcanic belt in map-area 55M/11 and along Thirty Mile Lake in the west half of map-area 65P/9. Cataclastic textures and chlorite, epidote, quartz, and carbonate veining are prominently developed in an east-west zone several kilometres wide south of the Dubawnt unconformity. Within this zone thin units of steeply dipping east-west trending amphibolite and paragneiss are interlayered with granitic gneisses. Chloritic schists and intensely altered deep pink to red granite gneiss cut by quartz stockwork breccias were mapped at several locations immediately south of the Dubawnt unconformity.

Diabase, Metadiabase and Amphibolite Dykes

East to east-northeast trending metadiabase dykes are common throughout the area but are particularly abundant south of Thirty Mile Lake and in the south half of map-area 55M/6. Dykes commonly retain fine grained chill margins but are typically marginally sheared, and boudinaged or faulted into short segments. Despite deformation and metamorphism, original porphyritic and glomeroporphyritic textures are commonly preserved, except in the north part of map-area 55M/11 where a penetrative tectonic fabric is developed.

A younger set of east-southeast-trending relatively undeformed diabase dykes cuts the east-northeast-trending set. Both sets are segmented by numerous southeast-trending faults.

Lamprophyre and Feldspar Porphyry Dykes

South-southeast- to southeast-trending porphyritic dykes, ranging in width from several centimetres to 10 m are present throughout the map-area and are unusually plentiful cutting Kaminak Group (?) rocks in map-area 55M/6. Rock types range from fine grained grey biotite lamprophyre to biotite-augite-feldspar porphyry (phenocrysts to 2 cm). The dykes are commonly emplaced along fault zones and may be fresh and undeformed, marginally sheared, or altered to chloritic schist in reactivated shears. Lamprophyre and porphyry dykes cut metadiabase dykes and are cut by granite dykes marginal to the plutons in map-areas 55M/3 and 55M/12. Biotite-augite-feldspar porphyry dykes may be related to the mafic syenite stocks discussed below.

Pyroxenite, Gabbro, Diorite

Small plugs and dykes of massive, medium- to coarse-grained pyroxenite, gabbro, and diorite intrude both granitoid gneisses and Kaminak Group (?) rocks (Fig. 81.2). Marginal deformation with replacement of clinopyroxene by amphibole and biotite is common. Breccias with abundant mafic inclusions in a gabbroic matrix are present along the margins of the dyke-like body in map-area 55M/4.

Syenite

Small stocks of porphyritic mafic-rich syenite form prominent hills in map-area 55M/5. Phenocrysts of grey potassium feldspar (to 3 cm), altered clinopyroxene and biotite are set in a dark grey matrix. The bodies intrude granitoid gneisses and Kaminak Group (?) rocks, contain xenoliths of the country rocks and are locally marginally sheared with development of amphibole and biotite.

Granite

The eastern half of map-area 55M/3 and a broad belt across map-areas 55M/11 and 55M/5 are underlain by large plutons of undeformed, fresh, pink granite. Large areas are covered by plains of granite felsenmeer with sparse outcrop. Typical granite is light pink, massive, and leucocratic (colour index 2-5), with textures ranging from aplitic to pegmatitic. Central areas of the plutons show remarkable textural and compositional homogeneity but border phases are crowded with xenoliths and migmatitic zones extend for several kilometres into the surrounding gneisses. Granite and pink pegmatite dykes and sheets intrude all units previously described. An undeformed pegmatite containing fluorite, topaz, and a rose mica cuts chlorite schists in map-area 55M/11.

Dubawnt Group

The north part of the map-area is underlain by poorly exposed sediments and volcanics of the Dubawnt Group. Donaldson (1965, 1967) has mapped a lower redbed sequence (South Channel and Kazan formations), a middle volcanic sequence (Christopher Island and Pitz formations), and an upper conglomerate-sandstone sequence (Thelon Formation). These rocks unconformably overlie cataclastically deformed granitoid gneisses (Fig. 81.2).

At the western end of Thirty Mile Lake the unconformity between South Channel conglomerate and basement gneiss is well exposed. The redbed sequence dips north at 55 to 75 degrees, exposing a 2000-m-thick basal conglomerate unit which passes transitionally upwards into sandstones of the Kazan Formation. The redbed sequence extends to the east into map-area 65P/9 where shallow north dips of 20 to 30 degrees are typical. North of the sediments purple, maroon, and brown porphyritic volcanics of the Christopher Island Formation are exposed along 20- to 40-m-high ridges. Ridge exposures of Christopher Island volcanics are typically the first outcrops of Dubawnt Group rocks exposed north of the unconformity in map-areas 65P/9 and 55M/12.

Gabbroic Diabase

Two 50- to 100-m-wide southeast-trending gabbroic diabase dykes are the youngest rocks in the area (Fig. 81.2). The massive, medium- to coarse-grained gabbro contains 5-10 per cent magnetite which accounts for the prominent magnetic anomaly that confirms the continuity of the dykes in areas of poor outcrop.

Economic Geology

Minor pyrite and pyrrhotite are present as disseminations in magnetite-quartz iron-formation. Metavolcanic rocks within the Kaminak Group (?) contain concentrations of iron sulphides plus minor chalcopyrite in thin gossan zones parallel to layering. Arsenopyrite, pyrite, and chalcopyrite occur in a 5- to 30-cm-wide gossan zone associated with amphibolite layers in granitoid gneiss just south of the gabbro dyke in map-area 65P/9.

The pyroxenite-gabbro dyke in map-area 55M/4 contains minor concentrations of iron sulphides.

Traces of molybdenite were noted in several pegmatite bodies in map-area 65P/9 and within granitoid gneisses in map-area 55M/12.

Uranium-copper mineralization in structurally controlled zones in basement gneisses south of the Dubawnt unconformity in map-area 55M/12 has been described by Laporte (1974). Copper mineralization was noted in a west-trending cataclastic zone marked by intense chlorite-epidote-calcite alteration on the north shore of Thirty Mile Lake in map-area 65P/10.

North of the Dubawnt unconformity uranium-copper mineralization occurs in sandstones and conglomerates of the lower redbed sequence (map-areas 55M/11, M/13, M/14 and 65P/10), and in northeast- and northwest-trending zones in porphyritic volcanics of the Christopher Island Formation (map-area 65P/9).

References

Davidson, A.
1970: Precambrian geology, Kaminak Lake map-area, District of Keewatin; Geol. Surv. Can., Paper 69-51.

Donaldson, J.A.
1965: The Dubawnt Group, Districts of Keewatin and Mackenzie; Geol. Surv. Can., Paper 64-20.

1967: Two Proterozoic clastic sequences: a sedimentological comparison; Geol. Assoc. Can. Proc., v. 18, p. 33-54.

Laporte, P.J.
1974: TMT project, Central District of Keewatin; p. 66-70, *in* Mineral Industry Report, 1969 and 1970, v. 2, Northwest Territories East of 104° West longitude, 191 p.

Reinhardt, E.W. and Chandler, F.W.
1973: Gibson-MacQuoid Lakes map-area, District of Keewatin; *in* Report of Activities, Part A, Geol. Surv. Can., Paper 73-1A, p. 162-165.

Wright, G.M.
1955: Geological notes on central District of Keewatin; Geol. Surv. Can., Paper 55-17.

1967: Geology of the southeastern barren grounds, parts of the Districts of Mackenzie and Keewatin; Geol. Surv. Can., Mem. 350.

STRATIGRAPHY, STRUCTURE AND METAMORPHISM OF THE WEST HALF OF THE FOXE FOLD BELT, BAFFIN ISLAND

Project 740020

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Work during the 1975 field season continued the study, initiated the previous year (Morgan *et al.*, 1975) of the Foxe Fold Belt on Baffin Island (Fig. 82.1). Approximately three months' field work was undertaken in 1975, using a Bell 47G4-A helicopter and canova boats to supplement foot traversing. To facilitate logistics a basecamp was set up in the vicinity of the DEW Line Station at Dewar Lakes (Fox 3) on 12 June, and was moved to Longstaff Bluff (Fox 2) on 1 August. Fixed-wing support, required at the beginning and end of the field season, and for the basecamp move, was by DC-3 aircraft supplied by Nordair from Frobisher Bay. Mapping, for publication on a scale of 1:250 000, was completed in Foley Island (37A), Lake Gillian (3D), and Koch Island East (37C east half) map-areas (Fig. 82.1). A limited amount of local mapping at 1:50 000 scale was also undertaken, chiefly from fly-camps, in the area (parts of 27B/5, B/11, B/12, B/13; 37A/8, A/9, A/10; 37C/9, C/15, C/16; 37D/4, D/5, D/7, D/8). No mineral occurrences of economic importance were located. Future field work that will involve mapping the Home Bay (27A), Ekalugad Fiord (27B) and McBeth Fiord (27C) map-areas for publication at 1:250 000 is scheduled for 1978 and 1979.

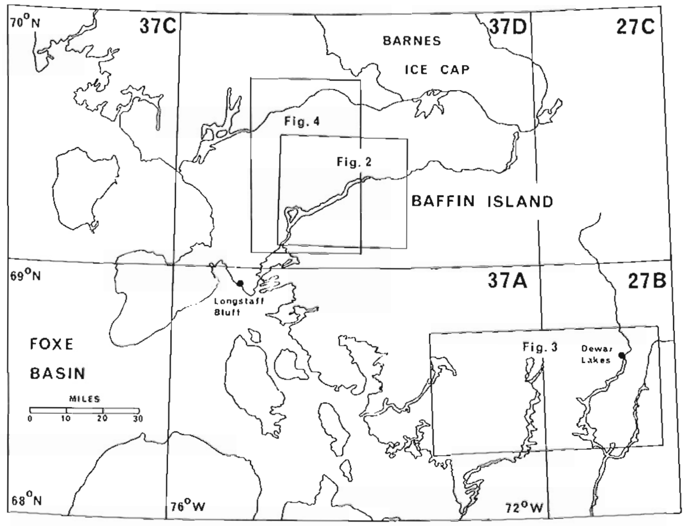

Figure 82.1. Index map of area investigated.

Stratigraphy of the Piling Group

The group consists of a miogeoclinal sequence of rocks, outcropping along the northern and southern margins of the Piling Basin, and a eugeoclinal sequence that occupies the central part of the basin.

The general stratigraphy of the miogeoclinal quartzite-marble-schist sequence that overlies basement gneisses in the north (Fig. 82.2) has been outlined by Morgan *et al.* (1975). No basal metaconglomerate was identified on the basement, which is generally overlain by a thin zone of sheared muscovite schist, followed by quartzite. The quartzite is sheared, massive or bedded, contains rare indistinct cross-beds, is of variable purity and, although commonly thin, ranges in thickness about 10 to 100 feet. A distinctive unit of carbonate rocks that includes marble, dolomite and calc-silicate gneisses, occurs above the quartzite, but locally, in highly deformed zones, it is in contact with basement gneisses. Thin horizons of quartzite, muscovite schist, rusty weathering graphitic sulphide schist, amphibolite, and silicate-oxide facies iron-formation occur in this unit. There is a considerable range in the total thickness of the carbonate rocks and although this is locally related to recumbent folding (Fig. 82.4), part of the variation is probably original. Thicknesses in excess of 600 feet have been measured in the Flint Lake region (Fig. 82.2) where folding is not intense. The upper part of the miogeoclinal sequence consists of rusty

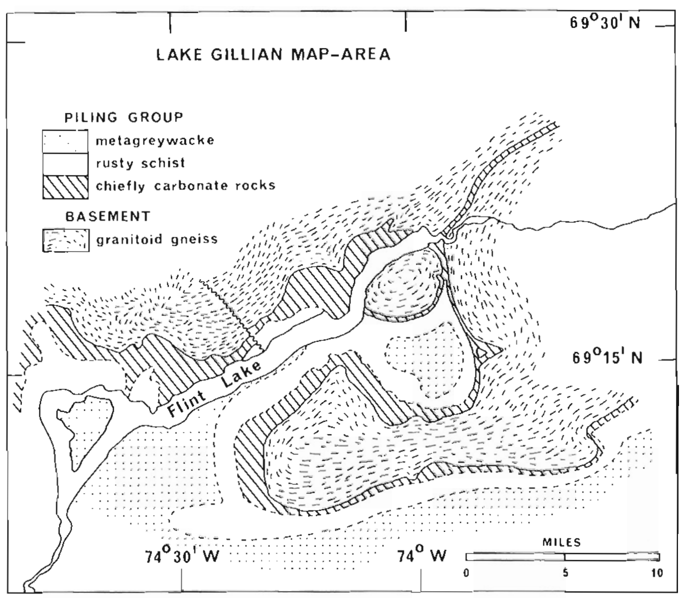

Figure 82.2. Distribution of carbonate rocks and structure at the northern margin of the Piling Basin, Flint Lake region.

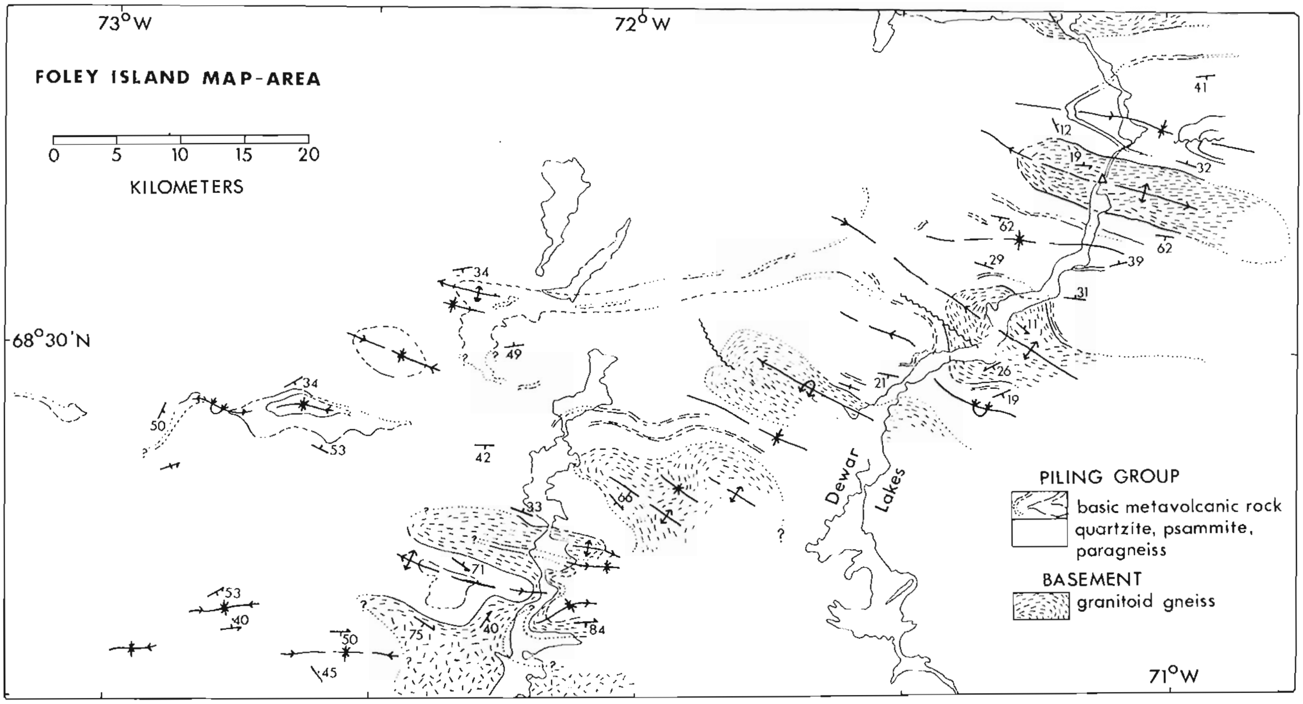

Figure 82. 3. Interference structures at the southern margin of the Piling Basin.

weathering graphitic sulphide schists with horizons of sulphide facies iron-formation.

No carbonate rocks occur in the miogeoclinal sequence that has been examined along the southern margin of the Piling Basin where the basement is commonly overlain by a zone of mica schist that is succeeded by a basal pink feldspathic quartzite. However, a few thin bodies of sheared, highly deformed metaconglomerate have been mapped locally in contact with granitoid gneiss. Most of the quartzite is white or grey, well-bedded, has excellent cross-beds, and contains horizons of mica schist. In the Dewar Lakes region the quartzite unit thins rapidly to the southwest, over a distance of 14 miles, from approximately 3000 feet to 100 feet.

Over 10 000 feet of very uniform, thin to thick bedded monotonous rocks that are chiefly metagreywacke, psammite and slate, occupy the central and greater part of the Piling Basin. The metagreywacke sequence is also well exposed near basement gneisses along the southern margin of the basin, where the underlying miogeoclinal sequence is thin, and is lithologically more varied. In addition to horizons of rusty graphitic sulphide schists, sulphide facies iron-formation, and calc silicate rocks that occur throughout the eugeoclinal sequence, the succession in the south contains amphibolite, basic metavolcanic rocks, quartzite, carbonate rocks and lean silicate-oxide facies iron-formation.

Structural Geology of the Piling Group

1. Dewar Lakes Area.

Polyphase structures developed during five episodes of deformation were mapped in the supracrustal-basement succession in the Dewar Lakes area (Fig. 82. 3). Near the DEW Line Station, basement gneissic rocks possess a relict fabric, very attenuated fold structures and evidence of flow under high temperature conditions. Development of these features was apparently accompanied or closely followed by generation and injection of pegmatitic sills. Truncation of these sills on the megascopic scale by basal quartzite and metaconglomerate of the Piling Group suggests that this intense phase of deformation preceded deposition of the group. Mesoscopic evidence for this presumed earliest phase (D_1) was not observed in the supracrustal succession.

Both basement and supracrustal rocks were deformed (D_2) about easterly trending axes forming recumbent, nearly isoclinal folds best developed in cross-bedded quartzite and pelite of the Piling Group. Subsequently, tight, upright approximately coaxial folds formed (D_3) and deformed earlier structures. Northerly trending fractures, some filled with pegmatitic dykes, observed within the basement and the metasedimentary rocks, may be related to this phase of folding. Upright open warps with north-northeasterly axes constitute the last folding phase (D_4) and produce variations in plunge of earlier folds. Extensive fracturing along an east-southeast direction (D_5) resulted from crustal movements and presumably

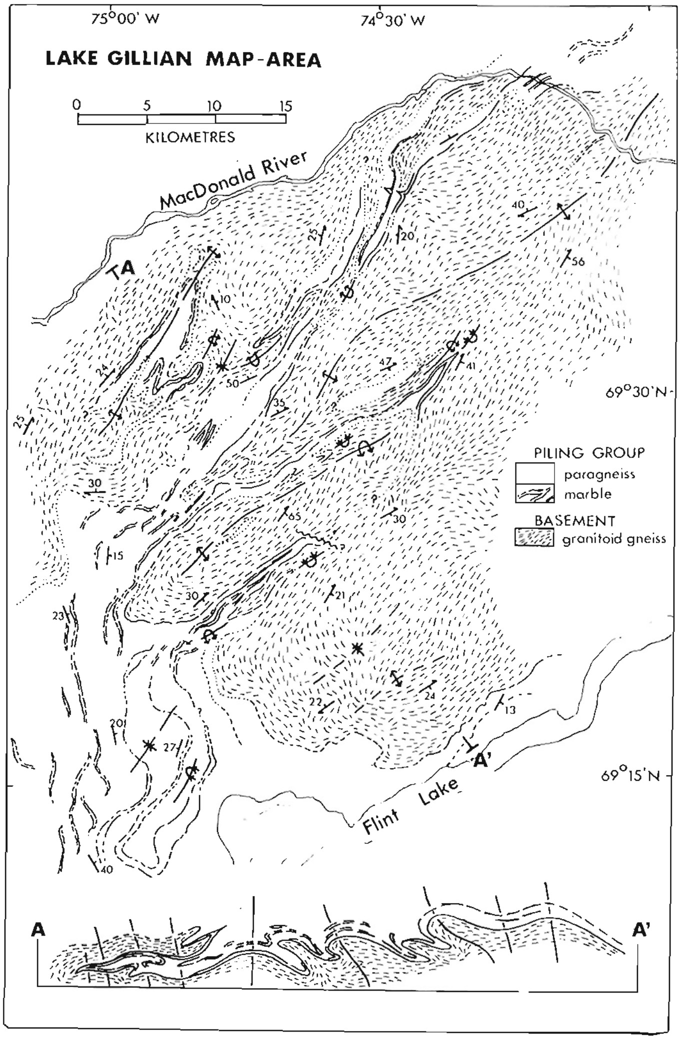

Figure 82.4. Basement-supracrustal relationships at the northern margin of the Piling Basin, between Flint Lake and MacDonald River.

tensional conditions associated with emplacement of large post-tectonic diabase dykes.

Interference between megascopic D_4 and D_3 folds results in an en echelon array of domal features within which basement is exposed (Fig. 82.3). These basement domes were mapped south and southwest into a highly mobilized zone of recumbent nappe structures cored by basement gneiss.

2. Foley Island map-area: southeastern part

Structural style in the Dewar Lakes area changes gradually to the southwest as the southern mobile borders of the Piling Basin are approached. Earliest (D_1) structures in the basement are rare or are no longer seen. In places the gneiss has been sufficiently fluid to become massive and structureless. The supracrustal succession, here predominantly paragneiss or psammite laced by pegmatite and other leucocratic intrusions, contains only pervasive foliation with rare rootless isoclinal folds as evidence of early deformation (D_2). Mappable megascopic structures indicate considerable mobility of the basement in this area. The southwest part of Figure 82.3 illustrates a recumbent nappe-like fold cored by basement gneiss with a southern root zone of massive granitoid rock and upper and lower limbs of supracrustal paragneiss, schist, quartzite and minor basic metavolcanic rocks. The domal aspect of this nappe is a result of refolding about east-west (D_3) and northerly (D_4) axes.

To the west, an en echelon array of basins and domes with unusual tightly appressed troughs and crests has formed within a succession composed predominantly of paragneiss overlain by amphibolite. Little recognizable basement gneiss was seen here. The structures are interpreted to have formed from interference of very tight upright D_3 folds and approximately orthogonal open D_4 folds.

3. Central Piling Basin

Fold structures within low grade psammite, metagreywacke and metavolcanic rock of the Piling Group are seen best on the megascopic scale by aerial reconnaissance. Few mesoscopic features developed within the thick monotonously bedded succession. Available data suggests that the intensity of deformation, concomittant with the metamorphic grade, is low compared to that observed to the east and south. Tight, upright to overturned, east-west trending folds (D_2) are folded by coaxial open warps (D_3). Slight variations in plunge of D_2 and D_3 structures may be a reflection of very gentle D_4 flexing.

4. Lake Gillian map-area: central part

Study of the region northwest of Flint Lake northward to MacDonald River suggests that here a tectonic regime differing somewhat from that described previously, controlled basement-supracrustal deformation. Pre-existing inhomogenieties in the basement may have been instrumental in determining later structures. Exposed areas of basement rocks are large and, except to the southwest, the supracrustal succession of paragneiss, carbonate rocks and minor quartzite, forms long narrow septa within the granitoid gneiss.

Evidence of D_1 structures within the basement gneiss is rare. Nearly isoclinal recumbent folds, overturned to both the northwest and southeast, with northeasterly trending axes are the earliest structures developed within the supracrustal rocks (D_2). Approximately coaxial, upright to overturned folds (D_3) deform early folds. These features are illustrated in Figure 82.4 and are accompanied by a tentative interpretation of megascopic structures. Resolution of the form of many of these has not been possible and the section shown should be considered as indicative only of the geometry of the folds. It can be seen, however, that a transition in structural style, from broad open antiformal warps near Flint Lake to tight, overturned folds to the northwest near MacDonald River,

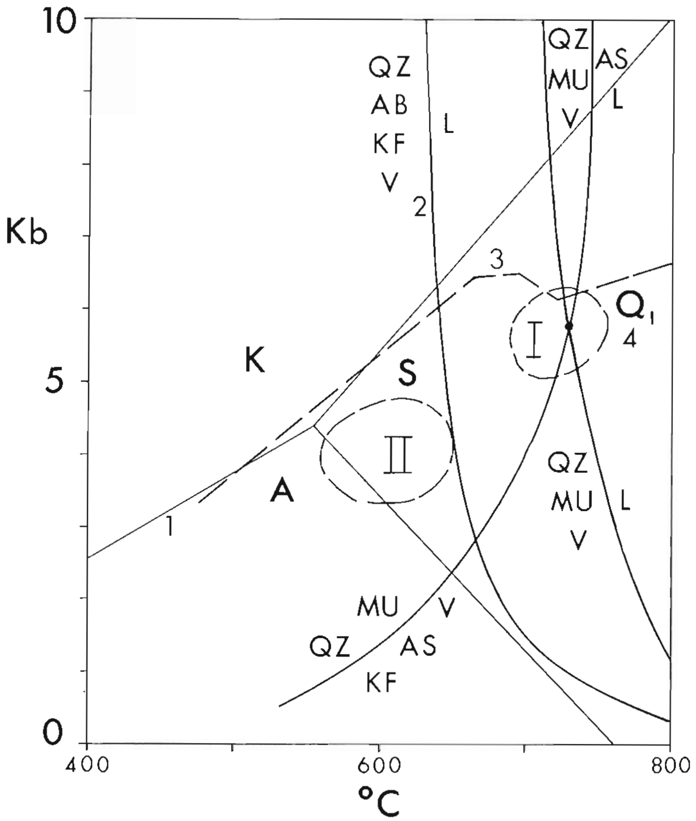

Figure 82. 5

Possible maximum pressure-temperature conditions of metamorphism of the Piling Group on the southern (I) and northern (II) margins of the Piling Basin. Experimental data include:

1. Al_2SiO_5 system (Froese and Gasparrini, 1975);
2. granite melting (Luth *et al.*, 1964);
3. upper stability limit of Mg-cordierite (Hess, 1969; Newton *et al.*, 1974);
4. invariant point Q_1 and associated equilibria after Huang and Wylie (1974).

Mineral abbreviations:

A	-	andalusite,	KF	-	K-feldspar,
K	-	kyanite,	AS	-	alumino-silicate,
S	-	sillimanite,	AB	-	albite,
QZ	-	quartz,	L	-	liquid,
MU	-	muscovite,	V	-	vapour (H_2O).

is present (Fig. 82. 2 and 82. 4). This is similar to that transition observed southwest of Dewar Lakes. In some areas, thrusting of gneiss over supracrustal rocks has taken place, possibly related to extreme closure of D_2 or D_3 folds.

Metamorphism of the Piling Group

Field observations of mineral assemblages in pelitic schist, calc-silicate gneisses, quartz segregations, psammite and metagreywacke, together with the regional distribution of pegmatite, indicate that metamorphic grade increases from upper greenschist facies (?) in the middle of the Piling Basin to upper amphibolite and granulite facies to the north, south and southwest. The most prominent metamorphic marker is the dramatic appearance of white, pegmatoid quartz-feldspar sills and dykes in the psammite-metagreywacke sequence and the associated transformation of the metasediments to biotite-quartz-feldspar paragneiss. The simple mineralogy of the pegmatoid rocks (quartz, biotite, plagioclase, K-feldspar? ± garnet, sillimanite), their contact relations with the psammite-paragneiss and the absence of plutonic rocks in the area, support the contention that the pegmatoid sills and dykes are derivatives of the psammite-metagreywacke sequence. The assemblage biotite-garnet-sillimanite is widespread in quartz muscovite pelitic schist. In these pelitic rocks, evidence for melting occurs at lower grade than the pegmatite zone boundary in psammite-metagreywacke.

Extensive melting of Piling Group metasediments has occurred without development of a prominent K-feldspar-sillimanite zone between pelitic rocks containing quartz-muscovite-biotite-sillimanite-garnet and their migmatitic equivalents. Migmatitic gneiss typically contains the assemblage biotite-garnet-sillimanite-cordierite-quartz-plagioclase-K-feldspar. Less commonly, pegmatoid bodies containing garnet-plagioclase-K-feldspar-cordierite-quartz or K-feldspar-sillimanite-quartz occur. In the southern high grade region, muscovite is rare in the pegmatitic derivatives of both the quartz-muscovite pelitic schist and the psammite-metagreywacke-paragneiss. The absence of an obvious K-feldspar-sillimanite zone and the mineral assemblages in the migmatitic gneiss suggest that these rocks were subjected to pressures of 5-6 kb (Fig. 82. 5) during metamorphism. These values would be maximum pressures because the composition of cordierite and muscovite in the rocks probably does not correspond to that of the end member compositions used in the experiments. Furthermore, the K-feldspar is probably microcline or orthoclase and not sanidine, as was used by Huang and Wylie (1974). The widespread occurrence of andalusite, cordierite and/or sillimanite along the northern margin of the Piling Basin is consistent with lower pressure, indicating that the depth of erosion there is less.

The extensive area of apparently low grade rocks in the central Piling Basin could represent a still shallower depth of erosion and a slower rate of post-metamorphic uplift relative to that in the high grade terranes to the north and south. If uplift was approximately uniform throughout, that is, the pressure difference between the northern and southern margins at the time of metamorphism was less than that indicated on Figure 82.5, the low grade area could represent a broad thermal depression between two thermal highs.

Sillimanite oriented parallel to hinge lines of folds that deform the regional foliation (S_2) and the pegmatoid sills described above, suggest that metamorphic grade increased during D_2 folding, attained a maximum during D_3 and declined during D_4.

References

Froese, E. and Gasparrini, E.
1975: Metamorphic zones in the Snow Lake area, Manitoba; Can. Mineral., v. 13, p. 162-167.

Hess, P.C.
1969: Metamorphic paragenesis of cordierite in pelitic rocks; Contrib. Mineral. Petrol., v. 24, p. 191-208.

Huang, W.L. and Wylie, P.J.
1974: Melting relations of muscovite with quartz and sanidine in the K_2O-Al_2O_3-SiO_2-H_2O system to 30 kb. and an outline of paragonite melting relations; Am. J. Sci., v. 274, p. 378-395.

Luth, W.C., Jahns, R.H. and Tuttle, O.F.
1964: The granite system at pressures of 4-10 kb.; J. Geophys. Res., v. 69, p. 759-773.

Morgan, W.C., Bourne, J., Herd, R.K., Pickett, J.W. and Tippett, C.R.
1975: Geology of the Foxe Fold Belt, Baffin Island, District of Franklin; *in* Report of Activities, Part A, Geol. Surv. Can., Paper 75-1A, p. 343-347.

Newton, R.C., Charlu, T.V. and Kleepa, O.J.
1974: A calorimetric investigation of the stability of anhydrous magnesium cordierite with application to granulite facies metamorphism; Contrib. Mineral. Petrol., v. 44, p. 295-311.

GEOLOGY OF ISLAND LAKE MAP-AREA (53 E), MANITOBA AND ONTARIO

Project 740016

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With contributions by G. Delaney and W. Fyson

Regional geological reconnaissance mapping on 1:250 000 scale was completed between June 3 and August 26, 1975. This season's helicopter-supported operations surveyed the whole area (53°N to 54°N, 94°W to 96°W), but mapping was concentrated on the mainly plutonic terranes surrounding the major Archean supracrustal belts mapped in 1974 (Ermanovics, 1975; Ermanovics *et al.*, 1975). Detailed structural studies on the Island Lake greenstone belt were initiated by Fyson (*see* below). Several sampling traverses were made for metamorphic and geochemical studies of the Island Lake belt by Delaney (*see* below). Samples of major plutonic and hypabyssal intrusive rocks were collected for age dating.

Chemical analyses, mainly of supracrustal rocks, are in progress.

General Geology

Surface Features

Mapping in the region is hampered by dense forest cover. Outcrops are plentiful in the north, especially around lake shores. Streams and lakes in the north are faithful topographic expressions of geological contacts, fault systems, and joint patterns while inland, tree growth similarly mimics fracture patterns, contacts, and the boundaries of glacial deposits. To the south and southwest there is considerable spruce-tamarack swamp and string bog, and large areas south and southeast of Island Lake are blanketed by Pleistocene glaciofluvial and lacustrine deposits.

Supracrustal Rocks: Major Greenstone Belts

Introduction. All regions of metavolcanic and metasedimentary rocks in the northern half of the map-area (Stevenson Lake – Collins Bay, Island Lake, Wapus Bay, Bigstone – Knight – Picket lakes: Fig. 83.1) probably were once continuous. Marginal intrusion of granodiorite – quartz diorite plutons, faulting along supracrustal/plutonic contacts, regional movements along major northeast-southwest, east-west and northwest-southeast faults, and erosion, have obscured the original continuity.

Each belt contains similar felsic to mafic extrusive and pyroclastic rocks, mafic sills, ultramafic rocks, porphyritic felsic intrusives, greywacke-siltstone-mudstone and arkosic sediments, and conglomerate horizons. While internal stratigraphic relations have been disrupted by folding, faulting, and shear movements, there is sufficient evidence to subdivide each belt into a lower predominantly volcanic sequence, and an upper predominantly clastic sedimentary sequence (Godard, 1963a, 1963b; Ermanovics *et al.*, 1975). In the Island Lake belt the lower sequence is termed the Hayes River Group, and the upper sequence constitutes the Island Lake "Series" (Wright, 1928; Quinn and Meinert, 1959). The exact relationship between the Island Lake "Series" and the Hayes River Group is uncertain. Godard (1963a, p. 34) summarized evidence for a parallel unconformity or disconformity.

Conglomerates. It is now clear that conglomerates at or near the base of the upper sequence in each belt provide much critical stratigraphic evidence. At three localities in the Island Lake belt (Fig. 83.1), conglomerate of the Island Lake "Series" containing quartz diorite boulders appears to unconformably overlie quartz diorite – granodiorite. On the south shore of Cochrane Bay a 1- to 2-m-thick regolith of angular granitoid clasts in a variably rusty, micaceous matrix, passing rapidly into fractured and altered plutonic rock, is preserved beneath the conglomerate. The contact is irregular and suggestive of a weathered surface at the other localities. Elsewhere quartz diorite and granodiorite intrude the Hayes River volcanics (*see* Delaney, below). Thus there is evidence that the base of the Island Lake sediments marks an erosional unconformity accompanied by a zone of weathering and that emplacement and unroofing of plutonic rocks began early in the history of the region. Relative to clasts of mafic and ultramafic rocks the conglomerates contain a large proportion of felsic to intermediate volcanic cobbles and porphyritic felsic material. Since mafic rocks predominate in the Hayes River volcanics, these observations could indicate substantial erosion of cyclical mafic to felsic volcanic piles, yielding largely felsic to intermediate debris from their tops.

The rare boulders of foliated granitoid rocks and layered gneisses seen within the Island Lake "Series" conglomerates indicate that deformation preceded the deposition. Layered gneisses also occur as blocks within conglomerate of the Bigstone Lake region. Ermanovics *et al.* (1975) have suggested that the blocks were derived from basement to the supracrustal rocks, exposed by the complete erosion of the lower volcanic sequence.

Excellent exposures of conglomerate in contact with leucocratic quartz diorite occur on the south side of the Stevenson Lake belt, east of Willow Lake. The conglomerate contains abundant boulders of leucocratic granitoid material, less abundant felsic and mafic volcanic cobbles, and minor ultramafic rocks, carbonate-bearing rocks, quartz-rich rocks, and iron-formation. Local irregularities in the contact and the brecciated and altered aspect of the quartz diorite (feldspar alteration observed up to 100 m from the contact) suggest that the conglomerate lies on a weathered surface.

Summary. All supracrustal/plutonic contacts in the Island Lake region have traditionally been considered intrusive (cf. Godard, 1963a, p. 31). Based on this season's observations the following alternative interpretation is suggested: A mainly volcanic sequence (Hayes River Group and equivalents) was deposited upon a gneissic basement. Intrusion of quartz diorite – granodiorite plutons followed. Substantial erosion of both volcanic and plutonic rocks (including regions of basement gneiss), and contemporaneous deposition of clastic sediments (Island Lake "Series" and equivalents) then occurred. More detailed work is required to expand this hypothesis, in particular by investigation of more supracrustal/plutonic contacts, and by consideration of possible lateral facies changes. In this regard the occurrence of volcanic rocks above the clastic sequence in the Bigstone Lake and Stevenson Lake belts (Ermanovics *et al.*, 1975) is an unresolved problem.

Other Supracrustal Rocks

Smaller outlying greenstone belts east and northeast of Oasis Lake and near Banksian Creek on the Manitoba-Ontario boundary (Quinn, 1960; Bennett and Riley *et al.*, 1967) cannot easily be correlated either with one another or with the larger belts (Fig. 83.1). Only metavolcanic rocks (including good pillowed basaltic rocks and coarse grained gabbroic sills east of Oasis Lake) occur in the smaller belts. The Cobham-Gorman rivers – Azure Lake belt in the south of the map-area is an extension of the Favourable Lake greenstone belt in Ontario, and contains calc-silicate assemblages which have not been found elsewhere in the map-area. Two additional, very small, metasedimentary belts southwest of Bigstone Lake, and southeast of Varveclay Lake in Ontario were discovered during 1975.

Metavolcanic rocks also form a component of migmatite zones that are present as horizons within granodiorite and quartz diorite batholiths, especially in the northern part of the map-area. Blocks of fine grained amphibolitic material of either intrusive or extrusive origin are rafted in a quartz diorite neosome along with coarser grained gabbroic and dioritic blocks, amphibolized dykes, and layered gneisses. Such migmatite zones may represent the eroded base of supracrustal piles, and their presence may be indicative of a greater original extent of cover rocks.

Plutonic Rocks

Regional Distribution. There are three distinct regions of plutonic rocks within the Island Lake map-area: 1) quartzofeldspathic gneisses and quartz monzonitic plutons lying north of the Stevenson Lake – Collins Bay supracrustal belt; 2) a similar gneissic complex of quartz monzonite – granodiorite plutons and quartzofeldspathic gneisses, surrounding the Cobham – Gorman rivers – Azure Lake paragneiss belt in the southwest and south; and 3) the central region of the map-area, containing the major preserved greenstone belts and their adjacent granodiorite – quartz diorite plutons, and regions of migmatite and layered gneisses.

Both gneissic complexes locally contain biotite- and hornblende-rich layers suggestive of assimilation of supracrustal material, but are dominantly composed of foliated to massive, pink, quartz monzonitic rocks, with local granodiorite, quartz diorite, and diorite. Development of these rocks in the south correlates with the relatively high metamorphic grade of the Cobham – Gorman rivers belt. In the northeast of the map-area, at McGowan Lake, conglomerate of the Island Lake "Series" is migmatized by quartz monzonite and granodiorite of the northern complex, indicating a relatively late development of that complex.

Central Region. Within the central region, the oldest rocks are probably the layered gneisses. They occur 1) adjacent to supracrustal belts, 2) as relics completely surrounded by granodiorite-quartz diorite material and migmatized by it, 3) on strike from supracrustal rocks, and 4) in migmatite zones with gabbroic, metavolcanic, and dyke rocks, that may represent the eroded base of supracrustal piles. They have previously been termed tonalitic gneisses (Ermanovics *et al.*, 1975). While their average chemical composition is probably quartz diorite, they contain amphibolitic and granodioritic layers. Some outcrops show concordant to semi-concordant amphibolite dykes within the layered gneisses. Near the margins of supracrustal belts their layering may be concordant with structural trends in the belt, but relatively large areas of layered gneisses occur with north-south layering, discordant to that of the supracrustal rocks.

Large areas of the central region are underlain by leucocratic quartz diorite – granodiorite plutonic rocks, with characteristic augen structure or cataclastic foliation trending east-southeast to southeast with steep dips. These rocks surround and are locally intrusive into the supracrustal rocks and layered gneisses; elsewhere contacts are faulted. They are probably the major source of granitoid boulders in the conglomerates. West of Island Lake and north of Bigstone Lake there are two distinct regions of these rocks, separated by layered gneisses at Begg Lake. South of Island Lake there are at least two large quartz diorite – granodiorite plutons, relatively coarse grained and indistinctly foliated. South and west of Bigstone and Cantin lakes there is a large, distinctive, coarse grained porphyritic quartz monzonite – granodiorite pluton with pink K-feldspar phenocrysts up to 10 cm by 2 cm. The relationship of these relatively coarse grained and undeformed bodies to the cataclastic granodioritic rocks is not clear, but they are probably younger intrusives.

Isolated plugs of gabbro, diorite, and ultramafic rocks are scattered throughout the central region. They are characterized by strong magnetic signatures. Two such bodies, migmatized by granodiorite, occur just west of Island Lake. They are zoned from hornblendite with relict pyroxene through gabbro and diorite to granodiorite.

Hypabyssal Intrusions

Several generations of diabase and gabbro dykes and their amphibolized equivalents are recognized. Older dykes in the granodiorite-quartz diorite plutons and in the layered gneisses and migmatite zones are variably sheared and altered, and trend northwest-southeast and northeast-southwest. The youngest mafic dykes trend approximately north-south and may be Proterozoic in age (cf. Molson dykes to the west, Ermanovics and Fahrig, 1975). They have unaltered chilled margins. A single north-south dyke cutting both basement and supracrustal rocks near Benson Bay on Island Lake, can be traced southward for some 45 km.

Dykes of intermediate composition, often porphyritic, cut both supracrustal and plutonic rocks.

Pegmatites and aplites are common, especially in the plutonic terranes.

Structure

Supracrustal belts in the map-area trend approximately northwest-southeast and northeast-southwest. This map pattern (Fig. 83.1) is a consequence both of the primary disposition of these belts, and of at least three phases of deformation (Ermanovics *et al.*, 1975, p. 314; Fyson, below).

Within the plutonic rocks, north-south layering and tight steeply dipping isoclinal folds in the gneisses probably represent the earliest structures. Cross-folds with east-west axial surfaces are then recognized in gneisses, and most other plutonic rocks are foliated in a west-northwest-east-southeast direction, often with development of augen structure.

Third-phase structures are represented by cataclastic foliation which develop near northwest-southeast and northeast-southwest faults. The faults are part of a large conjugate shear system, with right-lateral movement on the northwest-trending faults, and left-lateral movement on the northeast-trending faults. These faults are responsible on a regional scale for disruption of the greenstone belts in the northern half of the map-area. Movement on them may have continued for some time. Mylonite is locally developed. Late east-west fracturing is suggested by slight offsets of fresh diabase/gabbro dykes.

Metamorphic Grade

Studies of metamorphic grade in the Island Lake belt have now been initiated by G. Delaney. Grade in the map-area is generally greenschist to lower amphibolite facies, rising both north and south near the quartzofeldspathic gneiss complexes. Tiny garnets occur locally in aplite, and in granodiorite northeast of Island Lake. Epidote and chlorite are common along joints and foliation planes in plutonic rocks but primary assemblages in the granitoid rocks contain hornblende and biotite. Epidote, chlorite, hornblende, tremolite, and diopside variably characterize mafic rocks. Cordierite and andalusite occur locally within Island Lake "Series" sediments.

Economic Geology

Gold and copper mineralization occurs in association with porphyry intrusions. Copper and molybdenum mineralization occurs on northwest-southeast and northeast-southwest fractures within the Bella Lake pluton in the Island Lake belt. A fluorescent mineral in a single porphyry sample from Jubilee Island may be scheelite. Ultramafic rocks in the region have been prospected for copper and nickel. Several ultramafic localities are now known where high quality soapstone can be obtained for carving. Pleistocene sand and gravel deposits are relatively common.

Regional Correlation

Recently available high quality ERTS (LANDSAT) images, and aeromagnetic maps, greatly assist regional interpretation. Major topographic lineaments defining faults and supracrustal/plutonic contacts are easily seen on the ERTS photos; many lineaments are continuous for long distances outside the map-area. The four-mile aeromagnetic map of the Island Lake area (7274G) reinforces the identification of major faults and fracture systems and allows broad-scale correlation of magnetic anomalies with geology.

The map-area straddles a recognized major tectonic junction, that between the Berens River Batholithic Belt to the south, and the Sachigo River Volcanic Belt, which underlies much of the map-area (Ermanovics, 1973a, 1973b; Ermanovics and Davison, in press). A zone of high magnetic anomalies along the southern margin of the map-area, surrounding the Cobham — Gorman rivers paragneiss belt, correlates with the Berens River/Sachigo River junction, and coincides with the southern complex of quartzofeldspathic gneisses and quartz monzonitic plutons.

Along the northern margin of the map-area, there is also coincidence of high magnetization, quartzofeldspathic gneisses and plutons, relatively high metamorphic grade, and strong linear features. Another major tectonic junction may lie in this region, providing a means for subdividing the Sachigo River belt.

Geochemical and Metamorphic studies of the Island Lake Greenstone Belt

G. Delaney[1]

Introduction

The objectives of this work are: 1) to become familiar with stratigraphic and lithologic relations in the supracrustal rocks as a background for more detailed work; 2) to gain a representative view of the relative degree of metamorphism across the belt; and 3) to attempt a geochemical characterization of the volcanic members of the greenstone succession.

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94°
54°
53°
96°
McGowan Lake
York Lake
Kolman Lake
River
Oasis L.
Banksian
MANITOBA
ONTARIO
River
Varveclay Lake
LAKE
Benson B.
Bella Lake
Cochrane Bay
ISLAND
Wapus B.
Isbister
River
Cobham
Azure Lake
Collins Bay
Gorman Lake
Cordeau Lake
Cantin Lake
Dobbs Lake
Begg Lake
Willow L.
Knight Lake
Bigstone Lake
Stevenson Lake
Gunisao River
Hudwin Lake
20km
10
0
N

Six sampling traverses were completed, spaced across the whole strike-length of the greenstone belt. Other traverses examined: 1) the contact between the supracrustal rocks and the encompassing granitoid terrane; 2) the nature of small granitoid bodies within the greenstones; and 3) regions of anomalous metamorphic mineral assemblages such as the thin belt of cordierite schist which stretches southeastwards from the eastern end of Cochrane Bay (Fig. 83.1). Some 280 samples were collected. Godard's (1963a) maps were used as a control for the present work and were found to be quite reliable.

Stratigraphic Relations

The basal Hayes River Group is characterized by thick sequences of volcanic flows, fragmentals and pyroclastic material, with interlayered sediments. All rocks dip steeply. The volcanic rocks range from dark grey basalt with massive gabbroic phases, through drab greenish grey andesitic flows and tuffs, to white-weathering dacite to rhyodacite. Numerous primary volcanic features are preserved, including pillows, lava tubes, flow breccias and bombs. The sediments of the Hayes River Group consist of sequences of massive greywacke, pebble greywacke, subgreywacke, arkose, and laminated siltstones interbedded with mudstones. All sedimentary and volcanic rocks are weakly to strongly schistose. Zones of intense cataclasis are characterized by sericite and/or carbonate schists.

The contact between the surrounding granitoid terrane and the Hayes River Group is generally intrusive. This is defined by: 1) numerous xenoliths of mafic material (of varying sizes, shapes and degrees of recrystallization) within the quartz diorite – granodiorite bodies near their contact with the supracrustals, and 2) apophyses of granitoid material intrusive into the greenstones. An exception is the contact between andesitic lavas and the Loonfoot Island pluton at the eastern end of Island Lake. Here the granitoid material has an intense cataclastic foliation up to 150 m from the contact, and the lavas are extremely fractured, almost brecciated, for more than a kilometre from the contact. These relations suggest relatively cold diapiric emplacement of the granitoid mass (*see* Fyson, below).

(Legend for Figure 83.1 opposite)

- Supracrustal Rocks and Paragneisses
- (?) Proterozoic Diabase Dykes
- Position of Weathered Surfaces (?) on Plutonic Rocks
- Linear Features, from ERTS, Aerial Photographs, Aeromagnetic Maps
- Faults, Displacement Sense Indicated

The upper stratigraphic division in the belt, the Island Lake "Series", consists of a thick sequence of buff-weathering conglomerate which is overlain by a massive arkosic unit. Features of the Island Lake "Series" and of its contact with the Hayes River Group and plutonic rocks have been discussed above by Herd and Ermanovics.

Structural Development of the Island Lake Greenstone Belt

W.K. Fyson[1]

The succession of events suggested from (a) interpretation of existing maps (Godard, 1963a) and (b) additional observations by W. Fyson and T. Rivers. July 1975, mainly in the Collins Bay – Cochrane Bay and eastern portions of Island Lake, is as follows:

1) Mafic to felsic volcanics, tuffs, and flysch-type sediments of the Hayes River Group were deposited. No evidence of the nature of the basement was found.

2) Large granodioritic plutons were emplaced, with offshoots intruding the Hayes River Group. The minimal contact metamorphism in slates at one locality southwest of Cochrane Bay suggests 'cold' high-level emplacement, which in part may have followed later movements. Granitoid veins were observed with two foliations, equated with the two main phases of deformation (D_1 and D_2), and a steep intersection lineation, L_2, can be followed across one deformed contact of granodiorite with sedimentary rocks. Hence intrusion of the plutons was earlier than both the fabric-forming deformations. This view contrasts with that of Ermanovics *et al.*, 1975, p. 314, who interpret granodiorite intrusion between F_1 and F_2.

3) Deposition of coarse clastic rocks of the Island Lake "Series", accompanying differential uplift and erosion of plutons and marginal Hayes River rocks, then occurred. Proximal conglomerates contain abundant clasts up to boulder size of granitoid and mafic volcanic rocks, and in some localities, felsic volcanics. Internal fabrics within clasts are rare, as is to be expected if the plutons were intruded between D_1 and D_2 deformations. Any angular discordance of Island Lake "Series" rocks with Hayes River rocks has been obscured by deformation.

4) Further uplift of plutons and subsidence of marginal areas with formation of steep homoclinal successions of volcanics and clastics facing away from the plutons followed. Synclinal areas formed between the plutons. 'Cold' contacts of plutons with sedimentary rocks may be due to this uplift.

5) Regional deformation D_1 occurred, with development of F_1 folds with northwest- to west-trending steep axial surfaces, commonly conformable with margins of plutons. Individual folds extend several kilometres, but die out against marginal homoclines where

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these lie oblique to F_1 axial traces. F_1 folding may have directly followed uplift of plutons, but such oblique trends to margins indicate no simple causal relationship.

S_1 axial-surface foliation, apparently defined by aligned muscovite and chlorite (thin sections not yet examined) is only locally obvious in outcrop and distinct from subparallel S_2 foliation. Rarely exposed small-scale F_1 folds and axial lineations (trace of bedding on S_1) plunge at low angles, suggesting bedding was nearly horizontal prior to D_1.

6) Regional deformation D_2 followed, with small-scale folding about sub-vertical axial surfaces S_2 that curve from west-northwest to east-northeast. The S_2 axial-surface foliation is defined by crenulated S_1 schistosity where present, or by aligned micas or chlorite. S_2 passes into granitoid bodies as anastomosing cataclastic foliation with chlorite- and epidote-rich septa. As with D_1, D_2 was evidently during low-grade metamorphic conditions. F_2 axes and axial lineations are generally steeply plunging due to the intersection of S_2 with steep beds and S_1.

Although S_2 is the dominant foliation in many localities, as with S_1 it is variably developed. Both foliations are virtually absent from a zone, 3 to 4 km wide, which extends some 13 km between the Bella Lake pluton and the Loonfoot Island pluton at the eastern end of Island Lake. The zone transgresses lithologic boundaries and lies oblique to the strike of S_2 in adjacent rocks. Within the unfoliated zone pebbles in conglomerate and vesicles in volcanic rock appear undeformed. It thus seems that the zone between the plutons represents a large 'pressure-shadow' area which was protected from penetrative D_1 and D_2 strain. Elsewhere, deformed pebbles and vesicles indicate that the combined D_1 and D_2 strain was largely of the flattening type, with only locally a pronounced vertical elongation. The D_2 strain was presumably the result of regional horizontal compression directed approximately north-south.

7) Growth of cordierite and biotite, e.g., at the east end of Cochrane Bay, marked a metamorphic peak. S_2 foliation passes undeflected through cordierite crystals and biotites are randomly oriented, indicating that growth was after D_2. This metamorphism is possibly related to emplacement of late plutons (e.g., quartz monzonitic and granodioritic rocks north and northeast of Island Lake; *see* Herd and Ermanovics, above).

8) Local open F_3 folding occurred, about north-northeast vertical axial surfaces. Folds, some large-scale, associated with kinks, including northwest conjugate sets, and fracture cleavage were developed. F_3 structures are prominent along major north-northeast lineaments.

Intrusion of mafic dykes, some of which are pre-D_1 or D_2, and late faulting, have not been considered.

References

Bennett, G., Riley, and assistants
1967: Varveclay Lake sheet; Ont. Dept. Mines, Prelim. Geol. Map P430.

Ermanovics, I. F.
1973a: Precambrian geology of the Norway House and Grand Rapids map-area; Geol. Surv. Can., Paper 72-29.

1973b: Evidence for early Precambrian Mg-rich crust and 'pre-Kenoran' sialic crust in northwestern Superior Province of the Canadian Shield; *in* Volcanism and Volcanic Rocks; Geol. Surv. Can., Open File 164, p. 163-176.

1975: Island Lake map-area, Manitoba; Geol. Surv. Can., Open File 282.

Ermanovics, I. F. and Davison, W. L.
1975: The Pikwitonei granulites in relation to the northwestern Superior Province of the Canadian Shield; *in* The Early History of the Earth (ed. B. F. Windley), John Wiley and Sons. (in press)

Ermanovics, I. F. and Fahrig, W. F.
1975: The petrochemistry and paleomagnetism of the Molson Dikes, Manitoba; Can. J. Earth Sci., v. 12, no. 9, p. 1564-1575.

Ermanovics, I. F., Park, G., Hill, J., and Goetz, P.
1975: Geology of Island Lake map-area (53E), Manitoba and Ontario; *in* Report of Activities, Part A, Geol. Surv. Can., Paper 75-1A, p. 311-316.

Godard, J. D.
1963a: Geology of the Island Lake-York Lake area; Mines Br., Manitoba, Publ. 59-3.

1963b: Geology of the Sagawitchewan Bay area; Mines Br., Manitoba, Publ. 60-2.

Quinn, H. A.
1960: Island Lake Manitoba-Ontario; Geol. Surv. Can., Map 26-1960.

Quinn, H. A. and Meinert, R. J.
1959: The Island Lake Series, Island Lake, Manitoba; Precambrian, v. 32, no. 4, p. 15-22, 24-25, 34.

Wright, J. F.
1928: Island Lake area, Manitoba; Geol. Surv. Can., Sum. Rept. 1927, pt. B, p. 54-80 (Map 211A reissued 1935).

Project 740097

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Introduction

Regional stratigraphic and metallogenic studies (Ridler, 1975) of the Archean Abitibi Supergroup were continued in the summer 1975 (June 11 to September 10), principally in the Matachewan-Midlothian area (Fig. 84.1) and to a lesser degree in the Munro Township area of northeastern Ontario. The Ontario Division of Mines publications of Lovell (1967), Bright (1970) and Satterley (1952) have proven indispensable in carrying out this work. Brief field trips were completed in the Shiningtree and Timmins areas of northeastern Ontario, the Noranda area of northwestern Quebec, and the Chibougamau area of central Quebec. Sampling for isotopic ages in association with R.K. Wanless, and for primary paleomagnetism in association with J.H. Foster, was undertaken in support of regional stratigraphic correlation.

The author appreciates the generous co-operation of W. Karvinen, resident geologist, Ontario Division of Mines, Timmins; R. Walker, research geologist, Texas Gulf Sulfur, Timmins; H. Lovell, resident geologist, Ontario Division of Mines, Kirkland Lake; M. Carter, geologist, Shiningtree area, Ontario Division of Mines; G. McVeigh, geologist, Matheson; L. Cunningham and D. Lowe, prospectors, Kirkland Lake area; J. McIntosh, resident geologist, Q.D.N.R., Noranda; J. Cimon, resident geologist, Q.D.N.R., Chibougamua; E. Dimroth, professor, University of Quebec, Chicoutimi; G. Allard, professor, University of Georgia, and D. Fisher, graduate student, University of Toronto. The author was very ably assisted during the course of the field work by J. Sheen, undergraduate, University of Western Ontario.

A Reconnaissance Model for the Round Lake Batholith

The Round Lake Batholith (Fig. 84.2) is an elongate pluton (40 by 20 miles) with an average composition of a granodiorite. It has been interpreted as a magmatic intrusion (Lawton, 1954, 1957; Lovell, 1965, 1972; Moorehouse, 1944) of Algoman (i.e. post-kinematic, Kenoran) age. More recently, Ridler (1972, 1975) supported by Jolly (1974, and pers. comm.) has suggested that the batholith is basement to the Abitibi Supergroup, remobilized during the Kenoran orogeny and introduced into its current structural setting. Reconnaissance examination of the batholith, in association with sampling for U/Pb, zircon isotopic ages, was undertaken to test this hypothesis.

On a reconnaissance scale the batholith is comprised of three major litho-structural phases (Fig. 84.2). These are in decreasing age:

1) an interior annulus of relatively homogeneous granodiorite gneiss, in which the faint gneissosity trends east-west (Unit 1, Fig. 84.2),

2) a marginal zone of prominently compositionally layered, highly mylonitized and isoclinally folded gneiss (Unit 2, Fig. 84.2), and

3) an interior phase of massive granite, bearing melanocratic xenoliths (Unit 3, Fig. 84.2), a few of which appear to be derived from the border zone of Unit 2.

The contact of the batholith with the Supergroup may be "knife-sharp" (Ridler, 1972) or isoclinally infolded. A broad zone of "cool" contact metamorphism, characterized by the overprint of stilpnomelane-bearing assemblages on a pre-existing, probably diagenetic prehnite-pumpellyite assemblage, is found in the overlying volcanics (Jolly, 1974). The border zone rocks have undergone retrograde metamorphism (Jolly, pers. comm.). Randomly oriented amphibole porphyroblasts observed locally within a few thousand feet of the contact, suggest post-strain rehydration of the border zone from the volatile-rich adjacent volcanics. Boudins and septa of the relatively unmetamorphosed volcanics from the overlying Supergroup are found as pseudo-xenoliths near the contact. Analagous, physically incorporated fragments of border zone gneiss, suggest that intrusion of the batholith and consanguineous development of the border zone mylonite took place over a geologically long time. The incorporation of cobbles of the border zone in Timiskaming sediments at Kirkland Lake (Ridler, 1975) is consistent with this idea.

Large feldspar crystals which appear to overprint gneissosity were noted in Unit 1 immediately east of the eastern apex of the central intrusive phase (Unit 3). These may well be porphyroblasts associated with the intrusion of the core. The rare xenoliths of border zone in the core, its massive character, and its central position in the batholith, suggest that the core was cannibalized from the batholith and intruded late (post-strain) in the Kenoran orogeny.

The faint, but consistently east-west gneissosity of phase 1 is transposed on the eastern nose of the batholith by the gneissosity of the border zone, which everywhere parallels the contact of the batholith and the bedding of the overlying supracrustals (Fig. 84.2). The gneissosity of the interior annulus (phase 1) is thus older than that of the border zone (phase 2). On the north and south flanks of the batholith, where structural distinction is not possible, the banded border zone appears to grade imperceptibly into the "old gneiss". The border zone is thus a combination of cold-worked older gneiss with infolds and boudins of supracrustal rocks near the contact.

The Round Lake Batholith is intruded by the massive, granitic Crooked Creek Stock (Fig. 84.2), which, like the cannibalistic core (phase 3) carries xenoliths of the Border Zone (Ridler, 1975). Phase 3 and the Crooked Creek Stock are thus of the same relative age (i.e., late-Kenoran).

Zircons for isotopic dating have been obtained from: each of the major phases of the Round Lake Batholith along the section shown on Fig. 84.2, the cobble of Round Lake Batholith border phase in Timiskaming conglomerates at Kirkland Lake (Ridler, 1975), the Crooked Creek Stock, and the Otto Stock in Figure 84.2 (Lovell, 1972). Those obtained from the core zone and the Crooked Creek Stock are pristine. Zircons obtained from the border zone are zoned and abraded, while those obtained from the "old gneiss" are zoned but not abraded. The nature of the zircons can be explained by the preceding model for the development of the batholith.

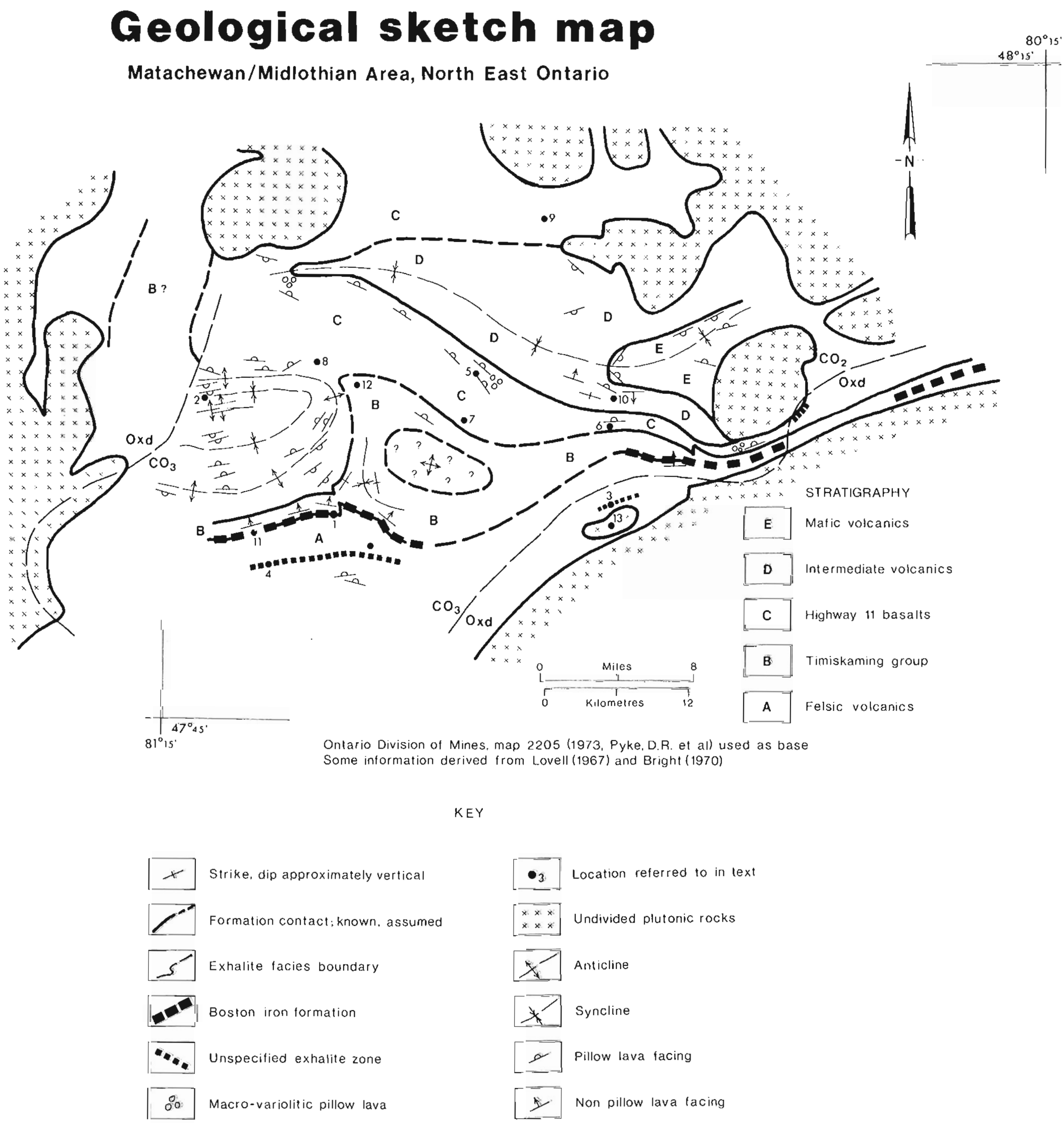

Figure 84.1.

Metallogeny West of Matachewan

Main Exhalite Zone

The main exhalite zone of the south margin of the Abitibi Basin, the Boston Iron Formation, has previously been traced westward from the Kirkland Lake area as far as Mistinikon Lake, west of Matachewan (Ridler, 1975). Flat-lying Aphebian rocks of the Cobalt Group discontinuously overlie the Archean for some 11 miles to the west of Mistinikon Lake, at which point the relevant strata reappear. Examination of outcrops in the vicinity of Stairs Gold Mine (location 1, Fig. 84.1) suggests the following (portions of which had previously been recognized by Marshall, 1947 and/or Bright, 1970):

1) The sequence is divisible into an older very felsic volcanic fragmental sequence and a conformably overlying sequence of extremely immature coarse clastic sediments of proximal volcanogenic derivation (Units A and B, respectively, Fig. 84.1).

2) Units A and B face north, dip steeply and strike east-west.

3) The contact is characterized by massive, siliceous, pyritic, fine grained carbonate with intermixed immature clastic sediments. The amount of carbonate in the sediments decreases gradationally above the felsic volcanics through several hundred feet. Cobbles of carbonate in the carbonate matrix were noted. Thus chemical sedimentation, erosion and clastic sedimentation overlap intimately.

4) The ore-bearing quartz veins inhabit a late, planar, discordant fracture cleavage which is apparently axial planar to mild undulations of the strata. The veins contain significant gold only where they intersect the main carbonate zone so the carbonate zone is probably the source of the gold via lateral secretion. The veins, though locally very rich in gold (Bright, 1970) are, by virtue of their small size, of little economic value. From a mineral exploration point of view they are chiefly useful as indicators.

The above stratigraphic relations are remarkably similar to those observed at Matachewan. An important difference is that Unit "A" has changed facies from mafic-intermediate tuffs and minor felsic breccia at Matachewan to an obvious felsic volcanic centre (Bright, 1970) in Midlothian Township, continuing the distal to proximal trend observed in the Matachewan area by Ridler (1975).

The probable correlation of the exhalite zone at Stairs with the Boston Iron Formation (the carbonate facies of which is usually called the Larder Lake Break), and its probable genetic connection with a major felsic volcanic centre, makes it a much more likely host for economically significant, large tonnage and low grade stratabound gold deposits.

The main exhalite zone continues to the west, where it becomes involved in the complexly refolded synformal structure indicated on Figure 84.1, and stratigraphic control is lost. An occurrence of classical Larder Lake-type carbonate exhalite (massive dolomite with chloritic septa) was discovered at location 2 (Fig. 84.1). Subjacent pillow lava contains identical interstitial carbonate-chlorite material, attesting to its exhalative origin. The stratigraphic relations of this exhalite zone are uncertain. Published data to the west and north (Pyke *et al.*, 1973) indicate a return to oxide facies exhalite (Fig. 84.1).

Other Exhalite Occurrences

The stratigraphic pile west of Matachewan is significantly thicker, and assuming that the structure is homoclinal, a lower exhalite zone has been identified (locations 3 and 4, Fig. 84.1). At location 3 the zone is at least 50 feet thick and comprised of intimately banded chert, pyrite, pyrrhotite(?), and magnetite (pyritic oxide facies). Very late quartz-carbonate veins bearing blebs of sphalerite and minor chalcopyrite have drawn development interest in the past. The zone is thus reminiscent of the oxide facies of the Boston Iron Formation in the Kirkland Lake area. It is possible that this zone correlates with the massive siliceous carbonate zone in felsic breccia examined at location 4 (Fig. 84.1).

Inadequate published information, and relatively poor outcrops, hinder interpretation of the metallogenetic relations at the Ashley Mine (location 5, Fig. 84.1). A series of basaltic, pillowed and massive lavas, some variolitic (Fig. 84.1), with minor interflow materials, strikes northwest and dips and faces northeast. This is the upper portion of Unit "C", the Highway No. 11 basalts (Fig. 84.1). The flow sequence is overlain by a thick sequence of micro-porphyritic intermediate breccias and crystal tuffs (Unit "D", Fig. 84.1). As at Stairs, economically insubstantial late, discordant quartz veins bear high gold values where they intersect a particular stratigraphic horizon in the basalt flows. Massive pyritic carbonate observed in the dump suggests that the source horizon may be an interflow exhalite. It is tempting to compare this setting with that of the Porcupine Camp, 35 miles to the northwest.

Metallogeny of the Matheson-Munro Townships Area

Metallogenetic studies of the Matheson-Munro townships area is in a very preliminary stage. Nevertheless, a few comments are justified.

Pipestone Fault

The "Pipestone Fault" which separates northeast-facing Hoyle Group sediments from northeast-facing mafic and ultramafic lavas and intrusions (Satterley, 1952; Pyke *et al.*, 1973), probably of the basal[1] Blake

[1] Basal Blake River Group is the "Kinojevis Group" of some authors.

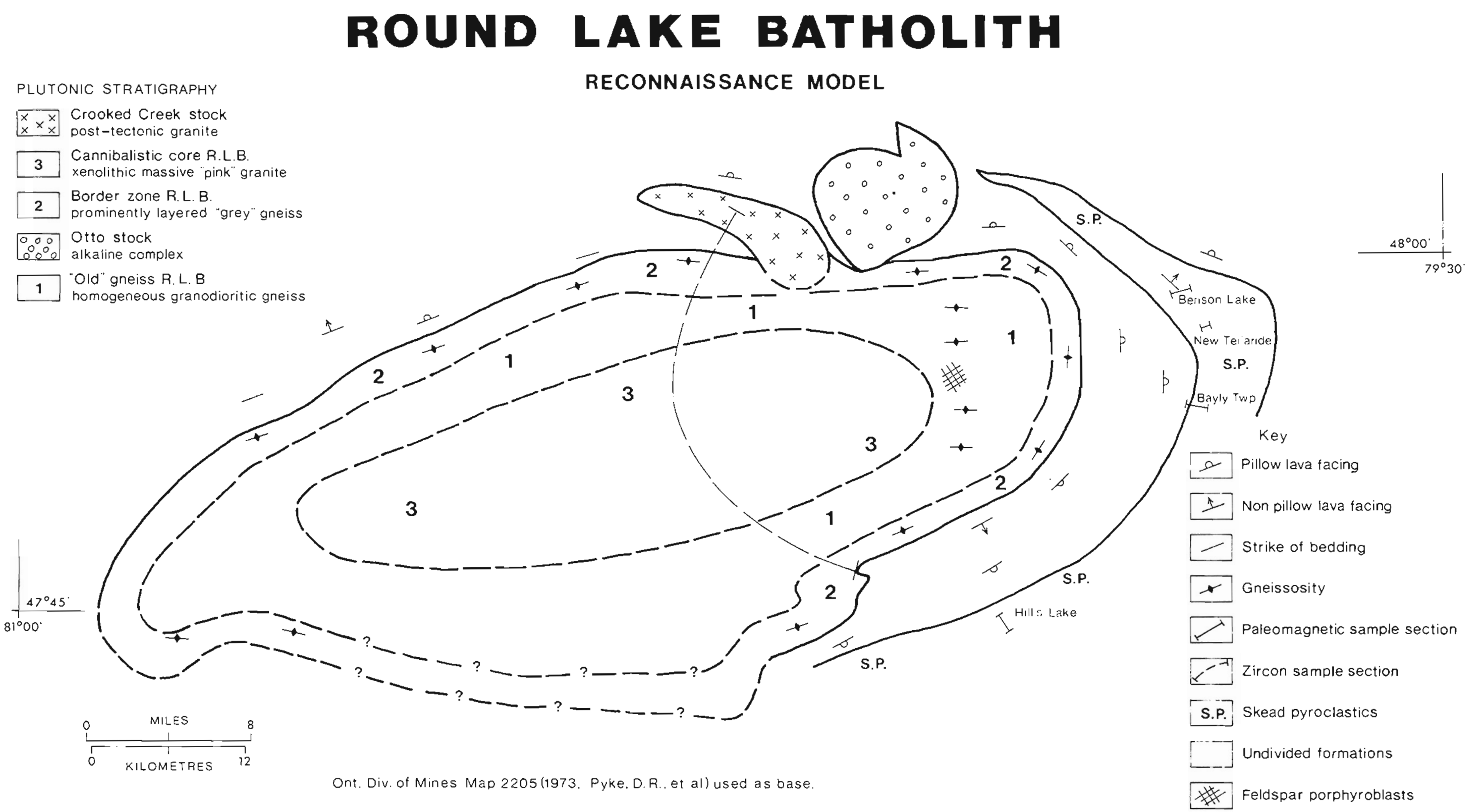
ROUND LAKE BATHOLITH
RECONNAISSANCE MODEL
PLUTONIC STRATIGRAPHY
Crooked Creek stock
post-tectonic granite
3
Cannibalistic core R.L.B.
xenolithic massive "pink" granite
2
Border zone R.L.B.
prominently layered "grey" gneiss
Otto stock
alkaline complex
1
"Old" gneiss R.L.B
homogeneous granodioritic gneiss
48°00'
79°30'
47°45'
81°00'
S.P.
Benson Lake
New Tel ande
Bayly Twp
Hills Lake
MILES
0
8
KILOMETRES
0
12
Ont. Div. of Mines Map 2205 (1973, Pyke, D.R., et al) used as base.
Key
Pillow lava facing
Non pillow lava facing
Strike of bedding
Gneissosity
Paleomagnetic sample section
Zircon sample section
S.P.
Skead pyroclastics
Undivided formations
Feldspar porphyroblasts

Figure 84. 2.

River Group, was examined in outcrop at several localities east of the road to the former Munro Asbestos Mine and north of Ontario Highway 101. In these localities, laminated turbidites with pronounced chert ribbing in places, some carbonate-rich laminae and one or two discontinuous beds up to one foot thick of rhyolitic tuff, are overlain by a relatively thin zone (less than two feet thick) of massive to weakly foliated carbonate exhalite. This sequence is directly overlain by pillow lava (the ambient volcanism), or by a discontinuous zone of quartz-eye rhyolite displaying fluidal flow banding and up to 100 feet in thickness. The foliation is in all cases oblique to bedding and is axial planar to small-scale folds observed in the outcrops. It is pervasive, though of greatly varying intensity on both sides of the contact.

The geological relations suggest a conformable contact between sediments and volcanics with no structural discordance, i.e., no fault. The paucity of chemical sedimentary deposition on the contact reflects: 1) a very low rate of exhalite deposition consistent with a distal turbidite environment far removed from source volcanism, 2) a relatively short time interval between the cessation of clastic sedimentation and the onset of effusive volcanism, or 3) both 1 and 2.

Regionally, the Hoyle sediments would be roughly equivalent to the Pontiac Group of the Noranda area. The geological relations described above, and exploration history, lower the economic potential of the contact. On the other hand, the potential of the overlying Blake River Group may be enhanced by the preliminary findings of field work in 1975.

Blake River Group

Although the area is undergoing active exploration, some conclusion can be presented without using specific mineral properties as examples.

The first few thousand stratigraphic feet of the Blake River Group north of the Hoyle sediments display little in the way of exhalative deposition, perhaps by virtue of an initial rapid rate of construction. Exhalative activity, however, did take place as shown by the presence of gold-rich, arsenopyritic, discordant, carbonate-rich, epizonal (epithermal?), pipes of diverse shape. A regional, relatively thick zone of carbonate exhalite (known as the Porcupine-Destor Break), does exist on the south limb of the presumed anticlinorium cored by the Hoyle sediments at what would be the appropriate stratigraphic level to be associated with the foregoing epigenetic exhalative activity. Above this stratigraphic level sulphide facies interflow exhalite associated with "rubbly" flow tops appears, perhaps reflecting the increased opportunity for deposition occasioned by a slackening in the rate of construction. The sulphide zones may be pyritic or pyrrhotitic. Sporadic gold mineralization has generated past development activity. Rarer occurrences bear economic concentrations of base metals (copper or copper-zinc). None of these zones can yet be recognized as an exhalite zone of regional extent, although this is a reasonable possibility, but they do coincide roughly with the appearance of discontinuous lenses of felsic volcanics in the sequence. Nevertheless, no direct genetic connection with felsic volcanism in the normal sense appears to be indicated by the geological relations and the association of unequivocal copper- zinc-bearing sulphide exhalite with basaltic or even ultramafic volcanism is an interesting example of the ubiquitous nature of volcanogenic chemical sedimentation.

Metallogeny of the Kirkland Lake Area

New Telluride

In the course of paleomagnetic sampling, some observations concerning the possible metallogeny of the New Telluride Mine (Hewitt, 1949) were made (Fig. 84.2). At the New Telluride, a stockwork of irregular, gold-rich, chalcopyrite-specularite veins cuts polymictic dacitic breccias and tuffs of the Skead Pyroclastics. A great deal of magnesian alteration is present in the vicinity of the showings, in particular, west of the New Telluride shaft, in what otherwise are extremely fresh volcanics. A zone of magnesian alteration with no known associated mineralization also was observed about 9000 feet along strike to the northwest (Fig. 84.2, Benson Lake section). If this is the same zone, albeit probably discontinuous, then significant but sulphur deficient exhalative activity must have been taking place. The apparent lack of an associated exhalite cap to this stringer zone and magnesian metasomatism may be an artifact of: 1) inadequate exploration, 2) non-deposition for a number of reasons, the most interesting being a possible subaerial site, or 3) decapitation by either primary causes (erosion, gravity sliding, etc.) or later faulting. The first possibility is most likely.

Main Exhalite Zone

Recent exploration activity (D. Lowe, L. Cunningham, pers. comm.), the results of which were examined briefly by the author, has confirmed the extension of the Larder Lake-type phase of the Boston Iron Formation well into the south limb of the Timiskaming Synclinorium southwest of Larder Lake (compare with Ridler, 1972).

Compilation of assessment data on file, coupled with field examinations by Lovell, Ploeger and Cunningham (pers. comm.) has extended the oxide facies of the Boston Iron Formation north of the Lebel Stock, bringing it into direct association with the carbonate facies, as documented in an analagous setting four miles to the west (Ridler, 1972).

Volcanic Stratigraphy

A shelf association east of Matachewan passes gradually into a basinal association in the Midlothian area (Fig. 84.1). A comparison of strata in these 2 areas is presented in Table 84.1.

Preliminary findings suggest the following.

Post-Timiskaming

The Highway No. 11 basalts formation (Unit "C", Fig. 84. 1) thickens dramatically west of Matachewan, into a classic mafic plate assemblage. A lowermost zone characterized by basaltic and ultramafic flows and intrusives (locations 6, 7, 8, 9 and others) is overlain by a predominantly basaltic sequence with a prominent macrovariolitic member (Fig. 84. 1), all of which is capped by a thin zone of probably andesitic composition.

Abruptly but conformably overlying the mafic plate is an unnamed, virtually unrecognized major accumulation (maximum thickness exceeds 10 000 feet) of intermediate to felsic, micro-porphyritic, unbedded breccia and massive crystal tuff (Unit "D", Fig. 84. 1). The unit is relatively unmetamorphosed and shows little evidence of deformation. This relatively unexplored proximal assemblage grades to the east through a heterogeneous zone of felsic fragmentals, carbonate-rich (Lovell, 1967), chloritic tuffs and extremely immature sediments (location 10, Fig. 84. 1) into distal turbidites north of Matachewan described by Ridler (1975) as "Post-Highway No. 11 Group Sediments".

Unit D appears to be overlain by the pillowed and massive lavas of the unnamed mafic plate, Unit E. Lack of information and the presence of a possible unconformity separating Unit E from the turbidites of "D" (Ridler, 1975) prevent a solution to this stratigraphic problem.

Timiskaming Group

West of Matachewan the Timiskaming Group thickens, diversifies, and becomes structurally complex (*see* discussion under metallogeny) changing its character rapidly a few miles west of the Stairs Mine (location 1, Fig. 84. 1). Immature, coarse clastic sediments at the mine give way to fragmental volcanic rocks. Nevertheless, the main exhalite zone outcrops at location 11 and a marker horizon of lithic tuff (Bright, 1970) bearing common massive sulphide and other exhalite fragments, as well as cobbles of spinifex-bearing ultramafics, is found at the same locality. Archean sedimentary formations commonly change facies along strike into fragmental volcanics. The Timiskaming Group exhibits similar behaviour. The author anticipates that the sequence of oxide exhalite-bearing tuffs (Pyke *et al.*, 1973) to the northwest (labelled B?, on Fig. 84. 1), are correlative, but the "nest" of refolded volcanics of great lithologic variety (Fig. 84. 1) must first be solved.

Table 84. 1

Comparison of Stratigraphy East of Matachewan and in the Midlothian Areas

East of Matachewan Shelf	Midlothian Basin
Distal oxide facies exhalite	Massive carbonate facies, in part proximal
A few thin formations	Numerous, varied, and thick formations
Characteristic distal tuffs, sediments and minor effusive lava	Proximal felsic volcanics; thick mafic plates, thick derived sediments
Relative structural simplicity	Structurally complex
Sequence rests on basement	Basement not known
Plutons common	Plutons rare (a few epizonal plutons present)
Intermediate metamorphism	Very low grade metamorphism

The Timiskaming Group southwest of location 7 is characterized by a proximal assemblage of coarse, intermediate polymict breccias bearing a high proportion of pumice fragments. No eutaxitic or welding phenomena were observed, so they are interpreted as shallow water deposits. The well bedded clastic sediments and laminated felsic tuffs south of location 6 (Ridler, 1975) are their distal correlatives. The relatively small, alkaline volcanic centre recognized in the Timiskaming Group at Matachewan (Ridler, 1975) is thus on the flank of the probably much larger volcanic structure in a basin to the west.

A series of folds and/or a pluton, as indicated by the question marks on Figure 84. 1, is required to account for: 1) the space between location 7 and location 1 on Figure 84. 1, 2) the known bedding orientations, and 3) the occurrence of typical Timiskaming conglomerate at location 12.

Miscellaneous

The small stock indicated at location 13 is a "late" syenite with a coarse grained trachytic border phase. It may be the westernmost member of the Kirkland Lake syenite suite. Barite mineralization in veins is spatially associated with the pluton.

A layered, ultramafic to gabbro sill, identified by the "top" indicator nearest location 2 (Fig. 84. 1) contains soapstone (probably after peridotite) at this locality.

Isotopic Dating in the Abitibi Supergroup

Sampling in association with R.K. Wanless continued. Zircon-bearing samples suitable for age determinations were secured from rhyolite at the Kidd Creek Mine of Texas Gulf Sulphur with the assistance of R. Walker, Texas Gulf Sulphur. In addition, samples were taken from five sites within potassic rhyolites in the Kam Kotia area west of Timmins. In this work the author was assisted by D. Fisher, a graduate student at the University of Toronto. M. Carter of Ontario Division of Mines collaborated in the sampling of rhyolitic volcanics at Shiningtree, 22 miles south of Stairs Gold Mine and at several localities in the vicinity of Stairs. These samples did not yield visible zircons nor did older Archean felsic volcanic sequences such as the Skead Pyroclastics and the Paymaster Porphyry of the Krist Fragmental. Lack of zircons in the older Archean units may reflect the less siliceous, and usually less potassic, nature of these rocks.

Paleomagnetic Program

In association with J. Foster, 40 sites (minimum 6 holes per site) were drilled to extend and further validate the primary paleomagnetic stratigraphy discovered by Ridler and Foster (1975). Sections drilled are indicated on Figure 84.2.

References

Bright, E.G.
1970: Geology of Halliday and Midlothian Townships; Ont. Dept. Mines Geological Report No. 79, 33 p.

Hewitt, D.F.
1949: Geology of Skead Township, Larder Lake area; Ont. Dept. Mines, Ann. Rept., v. 58, pt. 6, p. 1-43.

Jolly, W.T.
1974: Regional metamorphic zonation as an aid in study of Archean terrains: Abitibi Region, Ontario; Can. Mineral., v. 12, p. 499-508.

Lawton, K.D.
1954: The Round Lake Batholith and its satellitic intrusions in the Kirkland Lake area; unpublished Ph.D. thesis, University of Toronto.

1957: Geology of Boston Township and part of Pacaud Township; Ont. Dept. Mines, Ann. Rept., v. 66, pt. 5, p. 1-55.

Lovell, H.L.
1965: Flavelle-Sharpe area; Ont. Dept. Mines, Geol. Map P.264.

Lovell, H.L. (cont'd.)
1967: Geology of the Matachewan area; Ont. Dept. Mines, Geol. Rept. 51, 61 p.

1972: Geology of the Eby and Otto area, District of Timiskaming; Ont. Dept. Mines, Geol. Rept. 99, 34 p.

Marshall, H.I.
1947: Geology of Midlothian Township; Ont. Dept. Mines, v. 56, pt. 5, 24 p.

Moorehouse, W.W.
1944: Geology of the Bryce-Robillard Area; Ont. Dept. Mines, Ann. Rept., v. 50, pt. 4, 50 p.

Pyke, D.R. *et al.*
1973: Timmins-Kirkland Lake; Ont. Div. Mines, Map 2205.

Ridler, R.H.
1972: Volcanic stratigraphy of the Kirkland Lake area; *in* International Geological Congress, Precambrian Volcanism of the Noranda-Kirkland Lake-Timmins, Michipicoten and Mamainse Point areas, Quebec and Ontario; Goodwin, A.M. *et al.*, Guidebook A40-C40, p. 33-52.

1975: Regional metallogeny and volcanic stratigraphy of the Superior Province; *in* Report of Activities, Part A, Geol. Surv. Can., Paper 75-1A, p. 353-358.

Ridler, R.H. and Foster, J.H.
1975: Archean paleomagnetic stratigraphy of the Skead Group, Kirkland Lake area; Geol. Soc. Am., Abstr., v. 7, no. 6, p. 844.

Satterly, J.
1952: Geology of Munro Township; Ont. Dept. Mines, v. 60, pt. 8, 60 p.

ORIGIN AND DISTRIBUTION OF GOLD-BEARING CARBONATE ZONES OF THE KIRKLAND LAKE-LARDER LAKE AREA, ONTARIO

Project 740097

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It has long been recognized that practically all the important gold deposits found in the Larder Lake area occur either in or near bands of carbonate rock (Thompson, 1941). A study of the distribution of gold in the Larder Lakes area and relevant portions of the Kirkland Lake area, therefore, would be incomplete without an attempt to determine the origin of the carbonate rocks and the mechanism(s) of concentration of gold in these rocks. For this reason, the 1975 summer field season was spent investigating textures and field relations of the green carbonate rocks. Many of the known occurrences of these rocks in Eby, Otto, Teck, Lebel, Gauthier, McVittie and McGarry townships were plane-table mapped at a scale of 1 inch to 50 or 100 feet. This research was carried out as part of a Ph.D. thesis by Tihor, with the financial support of and in collaboration with the Geological Survey of Canada (R.H. Ridler).

The carbonate rock is principally a heterogeneous mixture of quartz, magnesite and dolomite according to X-ray diffraction studies of S.L. Tihor (pers. comm.). Generally, quartz predominates over the carbonates, however quartz may be virtually absent as in southwestern Teck Township. The green colour which characterizes most occurrences of this rock is caused by variable but generally minor amounts of fuchsite, a chromium-bearing muscovite. The rock is commonly very rich in pyrite, giving rise to rusty weathering outcrops. Locally fuchsite is absent and instead a black chlorite is found. Rarely, no ferro magnesian minerals are present and the rock is white or buff coloured.

Two main modes of occurrence of green carbonate rocks were recognized in the study area. In the western half from Highway 11 to the eastern edge of Gauthier Township, the rock typically occurs intimately intermixed with pink trachytic tuff (mode I). Some talc chlorite schist is also found in close association with the carbonates, especially in eastern Gauthier Township. Two ages of foliation are apparent in the green carbonates. The older foliation is expressed by the segregation of the constituent minerals into lensoid bands which parallel bedding in the enclosing rocks. The younger foliation generally cuts bedding at a variety of shallow angles reaching a maximum on fold noses.

In the eastern half of the study area, including the Kerr Addison Gold Mine, the green carbonate is associated primarily with mafic to ultramafic flows and schists. For example, at the west end of Bear Lake (eastern McVittie Township) the carbonates almost completely replace what appears to have been a flow breccia. The brecciated texture is well preserved and easily seen on weathered surfaces. On fresh surfaces, however, only massive green carbonate is visible. Every variation from completely replaced flow breccia to fresh mafic breccia free of green carbonate is present west of Bear Lake at the Old Cheminis Mine. Typically the green carbonate rocks of this type (mode II) are more massive and coarser grained than those of mode I. This may simply reflect the higher grade of metamorphism in the eastern portions of the study area (Jolly, 1974). On a regional basis the carbonate horizon overlies up to ten thousand feet of alkaline volcanic rocks (Ridler, 1970), interbedded and interfingered with shallow to deep water sedimentary rocks (Hyde, 1975). The carbonates are in turn overlain by turbidites and the Highway 11 Basalts (Ridler, 1970).

In southeastern Gauthier Township where Highway 66 crosses the Misema River, both types of green carbonate occur together. On the north side of the highway green carbonate is found intimately intermixed with pink trachytic tuff and gradations may be followed from pure carbonate rock to carbonate-free trachytic tuff. Narrow horizons of meta-argillite also grade into and out of this assemblage. The only sharp contacts are between small intrusive syenitic bodies and country rock. What may be an ultramafic flow with spinifex texture overlies these rocks on the south side of the highway. This grades upward into massive serpentinite which grades into talc-chlorite schist and, finally, the talc-chlorite schist into green carbonates. These are overlain by mafic flows.

Significantly, although no serpentine-bearing rocks had been previously reported in association with the carbonates at the Omega or the Cheminis Mines (Thompson, 1941), abundant serpentinite was found in the dumps of both mines. It seems that ultramafic rocks in contact with or near green carbonate rocks are rare west of Gauthier Township. A possible example is an occurrence along the Teck/Lebel Township boundary, near the Lebel Syenite stock. Here a large sill-like body of serpentinized peridotite is found about two thousand feet from the green carbonate occurrence. Because of lack of pertinent outcrop between these two units and intense folding in the area, it is impossible to determine the stratigraphic relationship between the peridotite and the carbonate rock.

Origin of the Green Carbonates

There is little agreement regarding the origin of the green carbonate rocks. Traditionally it was believed that they were formed by the alteration of pre-existing rocks through the action of solutions percolating upward through a persistent east-west

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shear zone called the "Larder Lake Break" (Thompson, 1941). The main difficulty with this hypothesis is that the carbonate zone appears to be stratigraphically controlled (Ridler, 1969, 1970) and the carbonate zone is folded along with enclosing strata. If faulting did have a role in metallogeny of this rock it must have preceded folding.

It has been suggested that the green carbonates are carbonate facies exhalites (i. e. chemical sediments of volcanic origin (Ridler, 1972)). The carbonate horizon has the physical configuration of a chemical sedimentary horizon -- great lateral extent (almost continuous for at least 60 miles and possibly up to 250 miles), but little vertical extent (maximum thickness about 600 feet). This origin is consistent with the fact that the carbonate is generally initimately associated with a variety of rock types, including alkalic pyroclastics and mafic and ultramafic flow rocks. In addition, across the entire study area the carbonate horizon follows with varying immediacy the end of alkalic volcanism (of the Timiskaming Group). In short, it seems to be an excellent stratigraphic marker horizon. However, the source of the huge volume of magnesium required to produce the horizon poses a problem. Volcanic exhalations are notably poor in magnesium, although often rich in carbon dioxide (White and Waring, 1963).

A number of field geologists familiar with the area (Lovell, pers. comm.) have suggested that the carbonates represent "altered ultramafic rocks". The close association of carbonates and ultramafic rocks and the similarity in magnesium content of these two rock types (Kerr Addison staff report, 1966) in the immediate Larder Lake area, lend some support to this theory. The carbonate zone, however, is virtually continuous across the Kirkland Lake-Larder Lake area and is not associated with ultramafic rocks in the immediate vicinity of Kirkland Lake. In fact, here it seems to be mixed on a microscopic scale with trachytic tuff.

The field evidence suggests the following sequence of events:

1. emplacement or deposition of a magnesite-dolomite rock, with or without syngenetic gold;
2. minor brecciation and quartz mineralization of part of the carbonate rock, possibly with some introduction or re-mobilization of gold;
3. intrusion of syenitic stocks and dykes with concurrent large-scale deformation. This probably included some brecciation of carbonates and introduction of more quartz, and possibly gold;
4. up to three more later stages of brecciation of carbonates and mineralization by quartz with or without introduction or re-mobilization of additional gold.

Laboratory study of distribution of immobile major, trace and rare-earth elements is planned for the near future, together with neutron activation analyses for gold to identify, if possible, the mineralogical site of gold in the carbonate horizon. It is hoped that this work, in addition to thin section and polished section microscopy, will shed further light on the origin of the gold-bearing green carbonate horizon.

References

Hyde, R. S.
1975: Depositional environments of Archean exhalites, Kirkland Lake-Larder Lake Area, Ontario; Geol. Soc. Am., Abstr., v. 7, no. 6, p. 789.

Jolly, W. T.
1974: Regional metamorphic zonation as an aid in study of Archean terrains: Abitibi region, Ontario; Can. Mineral., v. 12, p. 499-508.

Ridler, R. H.
1969: The relationships of mineralization to volcanic stratigraphy in the Kirkland Lake area, Ontario; Unpublished Ph. D. thesis, University of Wisonsin.

1970: Relationship of mineralization to volcanic stratigraphy in the Kirkland-Larder Lakes area, Ontario; Geol. Assoc. Can., Proc., v. 21, p. 33-42.

1972: Volcanic stratigraphy of the Kirkland Lake area; *in* Precambrian Volcanism of the Noranda-Kirkland Lake-Timmins, Michipicoten and Mamainse Point areas, Quebec and Ontario, by Goodwin *et al.*, International Geological Congress, Guide Book A40-C40, p. 33-52.

Thompson, J. E.
1941: Geology of McGarry and McVittie Townships, Larder Lake Area; Ont. Dept. Mines, Fiftieth Annual Report, v. 50, pt. 7, 97 p.

White, D. E. and Waring, G. A.
1963: Volcanic emanations; U. S. Geol. Surv., Prof. Paper 440K, p. K1-K29.

Sampling distribution and field description, 1975

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Introduction

In association with the "Metamorphic Map of the Canadian Shield" and "Regional Metallogeny and Volcanic Stratigraphy of the Superior Province" projects of the Geological Survey of Canada, a two-year sampling and thin sectioning program of the Abitibi greenstone belt and associated intrusive bodies has been undertaken. The aim is to gather closely spaced specimens from traverses along all available access routes, including roads, railroads, rivers, and lakeshores. Such sampling will permit determination of the areal distribution of metamorphic minerals from which a metamorphic history of the region may be developed. The purpose of this report is to summarize the distribution of samples collected in the first field season and to present preliminary results based on study of the hand specimens.

Geologic Setting

Lithologic subdivision. The rocks of the Abitibi belt may be divided for convenience into three broad groups:

1) Eruptive and associated subvolcanic rocks, with zircon ages of about 2.75 b.y. (Krogh and Davis, 1971), of great extent and thickness, are normally folded into narrow synclines between large granitic intrusions. The lavas are normally arranged in piles that become progressively more felsic upward (Goodwin, 1975; Jolly, 1975a; Gelinas and Brooks, 1975). The oldest lavas comprise ultramafic flows of Mg-rich komatiitic type, but these are mixed with moderately Fe-enriched tholeiitic types (Arndt, 1974). Then follows a thick sequence of strongly Fe-enriched tholeiites that display increasing breccia fractions concentrated at flow tops (Wilson and Morrice, 1975). The upper third of the typical stratigraphic section is composed largely of pyroclastic volcanic cones of mainly basaltic andesite composition (Jolly, 1975b).

2) Sediments, mostly of post-volcanic age but also apparently of pre- or synvolcanic age (Dimroth *et al.*, 1973), are normally exposed in synclinal troughs within the volcanic exposures. Most of the sediments are greywackes of deep water derivation. Flysch and iron-formation are common. However, in certain parts of the synclinal troughs, such as in the area near Kirkland Lake, Ontario, abundant textural evidence demonstrates that conglomerate was derived under subaerial conditions (Hyde, in prep.). Jolly (1974) reported that numerous clasts from the conglomerates carry a prehnite-pumpellyite mineralogy derived by erosion of the adjacent volcanic pile. These relations lead naturally to the conclusion that some of the sedimentary basins were formed atop the lava pile after or during its initial episode of folding.

3) Intrusive rocks, broadly circular in plan, are abundant at all edges of and within the lava-sediment pile. These are divided into four broad types:

a) Trondhjemitic bodies of layered or massive character, which are identical in composition to the tholeiitic and fragmental lavas near the top of the stratigraphic pile (Jolly, 1975b).

b) Granitic plutons (Fig. 86.1) of large size are commonly present within the lavas and the surrounding terrain. These rocks tend to display K/Na ratios higher than unity.

c) Granitic gneiss plutons (Fig. 86.1), commonly with igneous, equigranular granite interiors, are commonly found adjacent to and intruding volcanic rocks. These bodies invariably display K/Na ratios less than unity.

d) Late syenitic or alkaline intrusives, primarily of circular plan. Chemical data suggest that some of these intrusions are related to alkaline volcanic activity that occurred during development of the post-volcanic sedimentary piles (Cooke and Moorhouse, 1969).

Tectonic subdivision. The invasion of the thick volcanic pile by granitic diapirs was accompanied by doming and development of large scale, steeply dipping synclinal troughs in regions between the plutons; in addition, the volcanic sedimentary pile was everywhere gently folded during intrusion. The major synclinal features within the belt trend broadly east-west. Ages of the granitic materials average 2.4 b.y. (Kenoran, *see* Goodwin and Ridler, 1970). The degree of doming in the volcanics is proportional to the size of the individual plutons, so that larger intrusive bodies are surrounded by volcanic rocks from deep within the pile. Similarly, the stratigraphically highest lavas remained uneroded only in areas relatively remote from such intrusions. Thicker accumulations of lava, such as the pile in the area north of Kirkland Larder Lake area of Ontario, are not intruded by massive plutons, and behaved during Kenoran deformation as independent tectonic blocks of broadly synclinal character.

The interrelations between the plutons and synclinoria readily lend themselves to division into tectonic blocks (Fig. 86.2). Each such block is delineated by major synclinal features, commonly containing sediment, which surround uplifted volcanic

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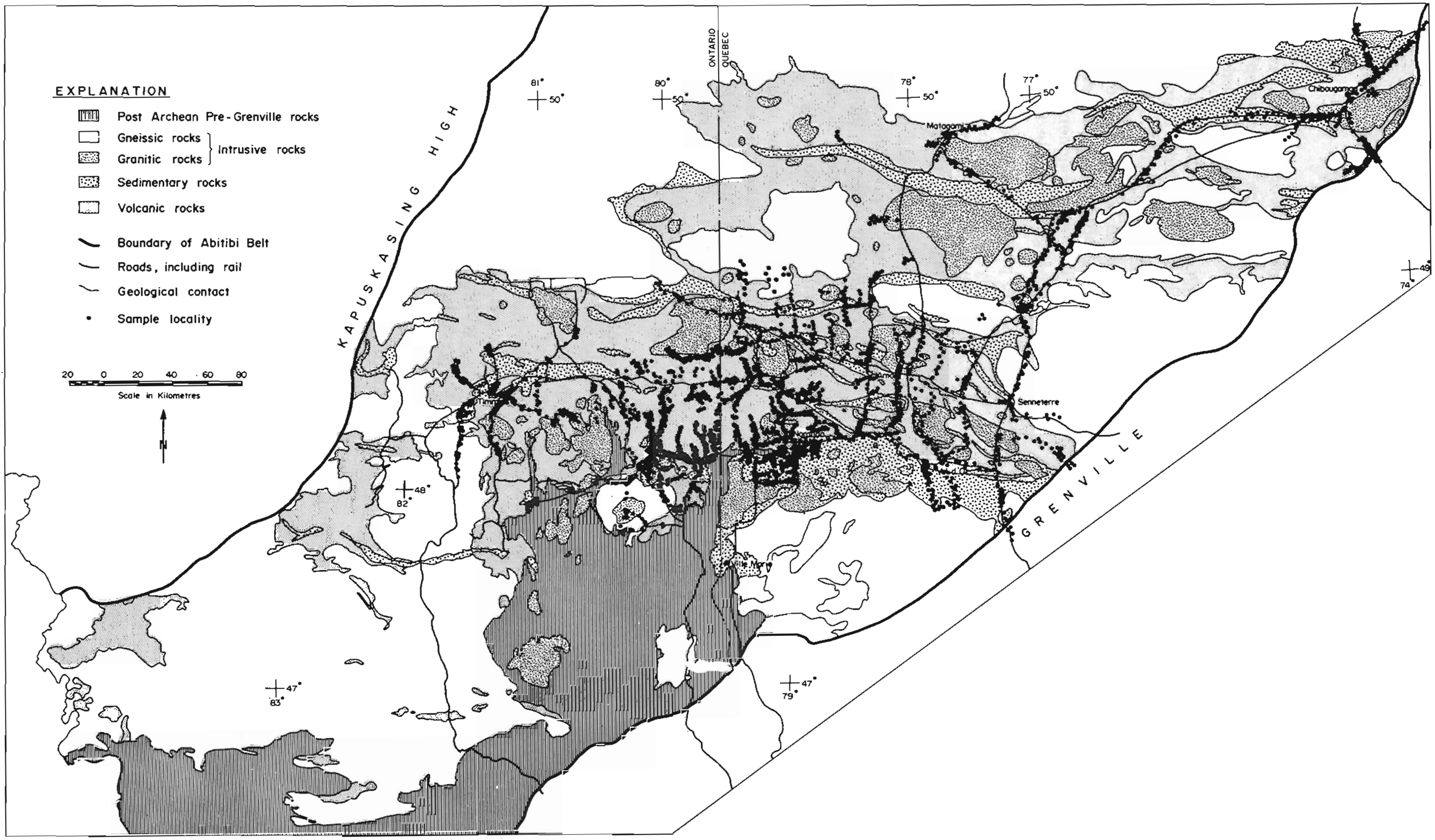

Figure 86.1. Geological sketch map of the volcanic part of the Abitibi as simplified from maps of the Quebec Department of National Resources and the Ontario Division of Mines. The terrane surrounding the volcanic piles are composed predominantly of gneiss-granitic intrusive materials. The sample localities for this study visited through 1975 are indicated.

rocks and the plutonic cores. At least 15 such blocks may be approximately delineated; each displays a dominant metamorphic rank which is dependent on the size or type of intruded plutonic core. The tectonic blocks are viewed as former centres of crustal convection in which the granitic cores rose to form anticlinal structures while the denser surrounding lavas sank into deep, steeply folded troughs.

Previous Metamorphic Studies

During township mapping of the Abitibi region by the provincial government surveys, gross metamorphic relations were summarized by each field party. Most such studies have classified the rocks into greenschist or amphibolite facies. Two studies of the area near the border between Ontario and Quebec have reported the presence of prehnite-pumpellyite-bearing secondary assemblages. Gelinas and Brooks (1974) mapped a pumpellyite "isograd" north of the Noranda-Rouyn mining camp. The isograd was considered a transition from relatively low rank metamorphism to the north to the greenschist facies southward toward Noranda-Rouyn. Jolly (1974) studied a strip of the belt centred on Larder Lake, Ontario and concluded 1) that the prehnite-pumpellyite alteration resulted from burial metamorphism during or closely postdating lava extrusion, and 2) that the "greenschist" or actinolite-bearing rocks resulted from intrusion of the granitic and trondhjemitic plutons. Thus, the latter was considered an over-printing of actinolite mineralogy onto the prehnite-pumpellyite assemblages. The latter was entirely obliterated in areas near intrusives. No zeolites have been reported from any of the Abitibi rocks.

Jolly (1974) and Ridler (1972) reported that gneissic intrusives, such as the Round Lake pluton south of Kirkland Lake, Ontario, are surrounded by a low-rank assemblage largely lacking hornblende, suggesting that intrusion took place at relatively low temperatures. Patches of pumpellyite-bearing rocks not over-printed by the actinolite assemblages are present in the metamorphic aureole surrounding the Round Lake pluton. Other non-gneissic granitic intrusions, however, display hornblende-garnet-bearing aureoles of relatively high temperature. Such observations will be extended in the present metamorphic map project, and may, in conjunction with other geologic data, help determine whether or not the volcanic rocks were erupted onto a granitic crust that was deformed by density instability during the Kenoran Orogeny, or if they represent a mafic crust erupted essentially on the Earth's mantle as oceanic crust that is forming at present in major oceanic basins. This is currently the most important problem to be solved by Archean research.

Results of 1975 Field Season

A total of about 2200 sample localities and over 2500 specimens have been collected from the Abitibi belt, thin sections are presently being made in the preparation lab at Brock University. Preliminary examination of hand specimens suggest the following conclusions:

1) Most of the granitic terrane surrounding the Abitibi belt is composed of a mixture of granitic igneous and gneissic material which is always intrusive into the lava pile. No clearly unconformable relations are known to occur, suggesting either that the density contrasts between the two rock types caused them to "pull apart" during the above mentioned convection phase (Kenoran), or that the intrusive bodies were introduced from igneous or other process at depth.

2) Contacts between volcanic material and gneissic rocks are generally actinolite-rich while those between volcanic material and equigranular, non-gneissic intrusions are of amphibolite rank. Significantly, amphibolite-rank rocks are absent from many areas near the borders of the volcanic-granitic terrane, contrary to previous assumptions (see, for example, Ont. Div. Mines, geological map of Ontario).

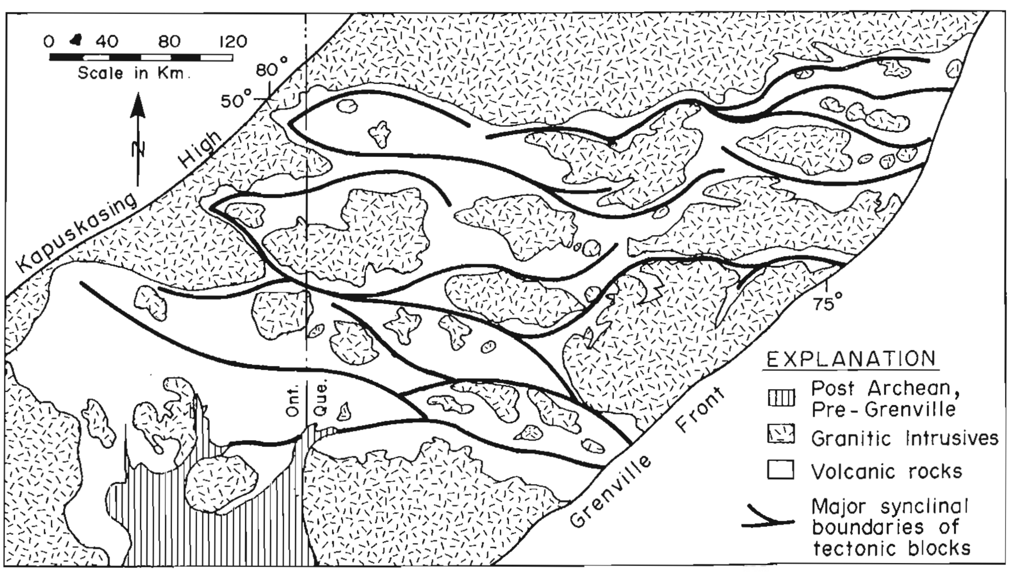

Figure 86. 2.

Tentative tectonic subdivision of part of the Abitibi belt, indicating approximate positions of major synclinal belts or structural breaks discussed in the text.

3) Prehnite-pumpellyite relics of the burial metamorphic phase of Jolly (1974) are widespread and concentrated in lavas relatively distant from the intrusive cores of the tectonic blocks previously described.

4) The rocks of the Superior Province are virtually unaffected by Grenvillian metamorphism until the contact with the Grenville Province (age about 1.0 b.y.) is reached.

Detailed analysis of metamorphic minerals already collected will be used to determine the distribution of the following key metamorphic zones: 1) prehnite-pumpellyite, 2) actinolite, 3) hornblende, 4) garnet, as well as the distribution of metamorphic Al-silicates in the sedimentary rocks of the belt. In addition, the field season of 1976 will be used for gathering samples from areas as yet uncovered (Fig. 86.1) in this project. Considerable effort will be spent sampling the areas between Lake Superior and Timmins, and the area south of Kirkland Lake in the vicinity of Timiskaming-Ville Marie. Boat traverses along the Harricanaw and Chibougamau rivers of Quebec will also be carried out to reach areas otherwise inaccessible. Finally, areas of interest delineated by thin section analysis will be more closely investigated to elucidate questions that might arise from the 1975-1976 thin section analysis.

Eventually, the samples gathered during the course of this work will be used to derive a chemical subdivision of the lavas and associated intrusions throughout the Abitibi belt (*see* Jolly, 1975).

Acknowledgments

The author wishes to express appreciation to R.H. Ridler for discussions in the field. This project is supported by the Geological Survey of Canada and the National Research Council of Canada.

References

Arndt, N.T.
1974: Ultramafic lavas in Munro Township: Economic and tectonic implications; Geol. Assoc. Can., St. John's, Nfld. meeting, abstracts vol., p. 2.

Cooke, D.L. and Moorhouse, W.W.
1969: Timiskaming volcanism in the Kirkland area, Ont.; Can. J. Earth Sci., v. 6, p. 117-131.

Dimroth, E., Boivin, P., Goulet, N., and Larouche, M.
1973: Tectonic and volcanic analytical studies in the Rouyn-Noranda area; Open file rep., Dep. Nat. Res. Quebec.

Gelinas, L. and Brooks, C.
1974: Archean quench texture tholeiites; Can. J. Earth Sci., v. 11, p. 329-341.

Goodwin, A.M.
1975: Archean volcanism in Superior Province, Canadian Shield; Geol. Assoc. Can. Special Paper, in press.

Goodwin, A.M. and Ridler, R.H.
1970: Abitibi orogenic belt; Geol. Surv. Can., Paper 70-40, p. 1-24.

Jolly, W.T.
1974: Regional metamorphic zonation as an aid in study of Archean terrains; Can. Mineral., v. 12, p. 499-508.

1975a: Subdivision of the Archean lavas of the Abitibi greenstone belt; Earth Planet. Sci. Lett., v. 27, p. 200-210.

1975b: Relations between Archean lavas and intrusive rocks of the Abitibi area; Geol. Assoc. Can. Special Paper, in press.

Krogh, T.E. and Davis, G.L.
1971: Zircon U-Pb ages of Archean meta-volcanic rocks in the Canadian Shield; Carnegie Inst. Wash. Yearbook 70, p. 241-270.

Ridler, R.H.
1972: Volcanic stratigraphy of the Kirkland Lake area; *in* Inst. Geol. Cong. Guide Book, A40-C40, Precambrian Volcanism of the Noranda-Kirkland Lake-Timmins, Michipicoten and Mamainse Point areas, Quebec and Ontario, Goodwin, A.M. *et al.*, p. 33-52.

Wilson, H.D.B. and Morrice, M.G.
1975: The volcanic sequence in Archean shields; Geol. Assoc. Can. Special Paper, in press.

GEOLOGY, DEEP RIVER MAP-AREA (31 K), ONTARIO AND QUEBEC

Project 750062

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Introduction

The Deep River map-area lies largely in southwestern Quebec. It is bounded by longitudes 76°W and 78°W and by latitudes 46°N and 47°N and covers a surface area of approximately 4600 square miles. The centre of the map-area is 100 miles northwest of the City of Ottawa. Various parts of the area have been investigated by the Ministère des Richesses Naturelles du Québec in recent years – Bourne (1970a, 1970b), Katz (1969), Kretz (1957), Lyall (1957), Marleau (1959, 1960), Rive (1970, 1972) and Sabourin (1963) have published maps at the scale of 1 inch to 1 mile or 1 inch to 2 miles for different portions of this region. In addition, Pourret (1968), Pourret and Bergeron (1970) and Wilson (1924) have studied the structural geology and economic geology of portions of this sheet. However, 30 per cent of this area has still not been examined on even a reconnaissance scale, and a geologic synthesis of the type undertaken by Wynne-Edwards (1966) for the area immediately to the east, has not been mapped.

During the summer of 1975 the writers visited most parts of the area, concentrating on unmapped areas in the north-central portions of the sheet. The purposes of the study were as follows:

1) To examine and correlate the geologic units defined by previous mappers;
2) To attempt to define and trace out any geologic boundaries of major significance;
3) To attempt a preliminary geological interpretation of the region.

Three major units have been recognized as a result of the field work. These are: rocks of the Grenville Supergroup, metamorphosed equivalents of Aphebian rocks, and metamorphosed equivalents of Archean rocks. The distribution of these units in map-area 31K is shown in Figure 87.1.

General Geology

Silver and Lumbers (1965) have dated the base of the Grenville Supergroup in southeastern Ontario at 1310 ± 15 m.y. (1285 ± 15, using modern decay constants). All other age determinations on Grenville Supergroup rocks are younger, with a minimum of about 1000 m.y. (Krogh and Hurley, 1968). The rock dated by Silver and Lumbers comes from the upper greenschist/lower amphibolite facies zone of the Hastings Basin (Hewitt, 1956), while the other reported ages are from higher-grade portions of the Grenville Supergroup. It is therefore possible that the older age is related to the time of formation of the metavolcanic rock dated, while the younger ages are related to the metamorphic episode known as the Grenvillian orogeny. Wynne-Edwards (1966, 1972) considers the Grenville Supergroup to be Aphebian or Paleohelikian.

The rocks herein classified as metamorphosed Archean rocks are so classified in order to be consistent with the scheme proposed by Wynne-Edwards (1966). There is no definitive proof that the rocks are of Archean origin; however, the discovery of east-west trends in these rocks and the broad correlation of rock types outlined by Fuh (1970), is convincing evidence.

The assignment of the remaining group of rocks to the Aphebian is speculative. No Aphebian Rb/Sr whole-rock isochrons have been reported from rocks of the map-area but an Aphebian age is considered likely for the following reasons:

1) The rocks are compositionally similar to those considered Aphebian by Lumbers (1971).

2) Rb/Sr whole-rock age determinations on the French River paragneiss by Krogh and Davis (1973), which is believed to belong to the same general group of rocks as those studied here, yielded ages of roughly 1850 m.y.

3) Dence *et al.* (1971) reported a K-Ar age of 1570 m.y. on hornblende (maximum possible age, 1700 m.y.) from samples taken in the vicinity of the Brent crater, immediately to the south of the present area. This suggests that no heating event sufficiently intense to result in the re-setting of the hornblende age, has occurred since this time. The rock dated by Dence *et al.*, is a pluton intrusive into metasedimentary rocks considered on lithologic grounds to belong to the same general group as those classified as Aphebian in this report.

4) A very strong northwest-southeast aeromagnetic trend is present throughout those parts of the map-area underlain by these rocks. In both the French River area and the Brent crater area, the aeromagnetic pattern of the rocks has a predominantly northwest-southeast trend, with local east-west perturbations.

5) Compositionally these rocks are very different from those classified as either Archean or Grenville Supergroup.

Description of the Rock Units

Grenville Supergroup

Rocks comprising the Grenville Supergroup are highly variable in composition. It is therefore impossible to state with certainty whether any given outcrop belongs to the Grenville Supergroup. However, two

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rock types found in abundance in the area underlain by the Grenville Supergroup rocks are found only sparingly elsewhere. These are marble and pelitic (aluminosilicate-bearing) gneisses.

Marble. Detailed descriptions of the marble units of the Grenville Supergroup abound in the literature and are not repeated here. They are in general snow white in colour and contain diopside, forsterite, graphite, pargasite, edenite, tremolite, and the minerals of the Humite group as accessory silicate phases. In the southeastern corner of the map-area, marble outcrops over many square miles and often appears massive, although compositional layering is usually visible on a fresh surface.

Pelitic Rocks. Pelitic units are intimately associated with the marble and possess a characteristic steel-blue colour on the fresh surface. The common diagnostic silicate mineral assemblage is garnet ± sillimanite (± cordierite) (no kyanite) + K-feldspar + melt. In rocks containing sillimanite, the garnets are commonly mauve in colour, whereas rocks lacking sillimanite have garnets that are generally red in colour. The reason(s) for this feature are under investigation. This relationship is general, but not invariable.

Metamorphic Equivalents of the Archean Rocks

Only one rock type, within the area mapped by Wynne-Edwards (1966) as Archean in age ('granite gneiss', map-unit 1) is both easily recognizable and widespread in this map area. Sabourin (1963) divided this unit into two sub-units – pink injection gneiss and grey injection gneiss – while Lyall (1957) called these rocks grey biotite granite gneiss. They are medium grained rocks which are quite homogeneous on the scale of the outcrop. They have been subjected to partial melting – in some of the rocks the melt rock is pink in colour, while in others it is white, hence the two-fold subdivision of Sabourin. Compositional layering is generally present but it is faint and overshadowed by compositional layering defined by the mobilizate/restite relationships.

In thin section they have two features which distinguish them from most of the rocks in the map-area:

1) They contain substantial amounts of K-feldspar (microcline).
2) They have been retrogressively metamorphosed, resulting in the development of significant amounts of muscovite and epidote, and small amounts of chlorite. Other rocks show this feature only locally.

The best exposure of this rock type (the "type locality") is located on highway 117 (former hwy. 58) at the south end of Lac Roland (mile 200) (former mile 58). This outcrop was investigated in detail by Pirie (1971). Excellent exposures of this rock type occur:

1) on highway 117 from the northern boundary of the map-area south almost to the gates of the Parc de la Verendrye;
2) west on C.I.P. route number 15 between miles 15 and 45.

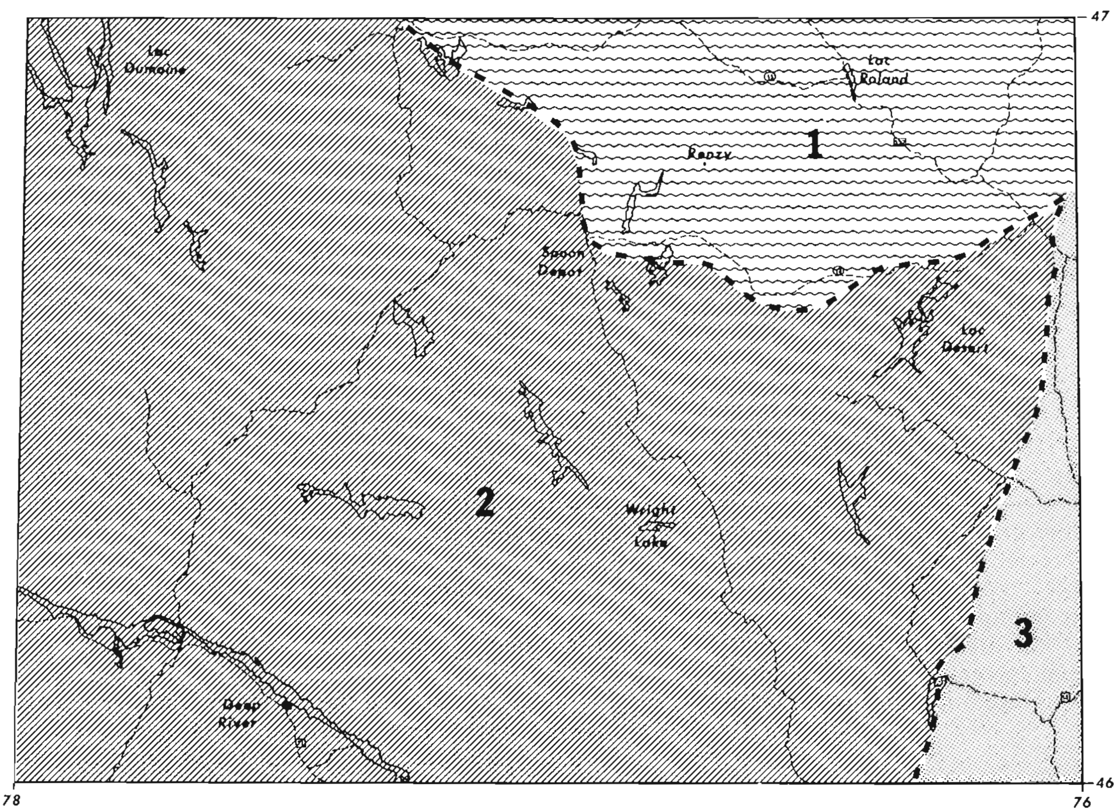

Figure 87.1

Distribution of major rock units.

Unit 1: predominantly metamorphosed equivalents of Archean rocks.

Unit 2: predominantly metamorphosed equivalents of Aphebian rocks.

Unit 3: rocks of the Grenville Supergroup.

The major access roads have been indicated by a light dashed line.

Associated rock types are described in detail in references cited above.

Metamorphic Equivalents of the Aphebian Rocks

This unit is a monotonous gneiss complex of probable sedimentary origin. It is the metamorphosed equivalent of a thick greywacke succession with minor shale and siliceous limestone members (Lumbers, 1971).

These rocks as a whole are poor in K-feldspar, and generally finer grained than the metamorphosed equivalents of the Archean rocks discussed above. In addition, the compositional layering is well defined in the outcrop. These rocks are probably equivalent to those shown on the Geological Map of Ontario as Middle to Late Precambrian (unit 11).

Near the boundary with the metamorphosed equivalents of the Archean rocks, the unit is much richer in potassium feldspar and its foliation is more irregular in direction. The rocks near this boundary are not typical of the rocks found throughout the rest of this zone to the west. Some outcrops appear to represent a "tectonic mixture" of both primary Aphebian and Archean rocks, and are probably a lateral extension of the migmatite zone mapped by Wynne-Edwards (1966). The location of the boundary between the metamorphic equivalents of the Archean and Aphebian rocks in Figure 87. 1 is only approximate for this reason.

Thin layers of siliceous marble occur throughout the area underlain by Aphebian rocks. Available evidence shows that there must be at least two primary ages for metacarbonate units in the Grenville Province (Frith and Doig, 1973; Krogh and Hurley, 1968), and it cannot be assumed that all metacarbonate rocks found within the geographical confines of the Grenville Province are a part of the Grenville Supergroup. This agrees with the authors' field observations. There is a close spatial correlation between the thin layers of siliceous marble mentioned above and the Aphebian metagreywacke succession. There is little doubt that they are related units.

Metamorphism

All of the rocks of the area, with the exception of a few diabase dykes, contain upper amphibolite and locally, granulite facies metamorphic mineral assemblages. Minerals such as andalusite and staurolite are unknown. Sillimanite is the predominant aluminosilicate and co-exists with K-feldspar and presumed melt throughout most of the area. In the northwest corner of the map-area, a zone of muscovite + quartz-bearing rocks was outlined by Rive (1972). The common aluminosilicate in this region is kyanite, which is replaced by sillimanite at approximately the same location as sillimanite + K-feldspar replace muscovite + quartz. No assemblages containing K-feldspar + kyanite were found.

Granulite facies rocks are only locally present. Limited data suggest that they are confined to a zone of higher aeromagnetic relief, which trends north-south, and bisects the map-area. Within the area of high magnetic relief, the granulites are situated around the margins of dome-like structures which contain cores of amphibolite facies rocks. The best known example of this feature found this summer is along the north shore of Wright Lake (31 K/7 west half).

Economic Geology

In recent years the only producing mine in the area has been the base metal deposit of Renzy Mines Limited. During the years 1969-1973, 12 943 527 pounds of both copper and nickel valued at $6 128 802 were produced (Fielder and Worobec, 1975). The mine is located at the margin of a significant aeromagnetic high, the reason for which is not clear. Immediately west of this region lies a much larger aeromagnetic high with a form similar to that of the Renzy Lake high, and hence the two might conceivably be related. The margins of the larger anomaly in the vicinity of Spoon Depot, might be worth investigation. There is good road access to this area.

Several uranium occurrences have been reported from the southeast corner of the map-area (Bourne, 1970a; Kretz, 1957). They appear to be associated with thinly bedded carbonate units (which have been metamorphosed into tactite/skarn deposits), of the type associated with the Aphebian rocks.

Geological Interpretation

It has been postulated that three major units are present in the map-area — metamorphosed rocks of originally Archean, Aphebian and Neohelikian age, respectively.

Those deemed to be Archean are considered to be a southeasterly continuation of the Archean rocks of the Superior Province. The presence of east-west trends within the Archean rocks in the Grenville Province supports this hypothesis. It is assumed that the Archean rocks acted as a stable craton upon which the Aphebian age metagreywacke suite was deposited. The two groups of rocks were then deformed and metamorphosed approximately 1850 m.y. ago, and subsequently intruded by a suite of plutons approximately 1700 m.y. ago. Presumably both the Archean and Aphebian rocks formed the stable craton upon which the rocks of the Grenville Supergroup were deposited.

There is no conclusive evidence that the rocks of the Grenville Supergroup were ever present outside a circle of approximately 150 mile radius centred at Renfrew, Ontario (Engel, 1966, p. 78). Small wisps of marble have been found in the Grenville Province outside of this circle, but their relationships to the marbles of the Grenville Supergroup are not known.

References

Bourne, J.H.

1970a: Geology of the Cayamant Lake area; Que. Dept. Nat. Resour., Prelim. Rept. 598, 20 p.

1970b: Geology of the Pythonga Lake area; Que. Dept. Nat. Resour., Open File rept. GM-28638, 22 p.

Dence, M.R., Hartung, J.B., and Sutter, J.F.
1971: Old K-Ar ages from the Grenville Province, Ontario; Can. J. Earth Sci., v. 8, p. 1495-1498.

Engel, A.E.
1956: Apropos the Grenville, *in* 'The Grenville Problem'; Roy. Soc. Can., Spec. Publ. 1, p. 74-96.

Fielder, F.M. and Worobec, A.
1975: Canadian Mines Handbook 1975-1976; Northern Miner Press, Toronto.

Frith, R.A. and Doig, R.
1973: Rb-Sr Isotopic ages and Petrologic studies of the rocks in the Lac St. Jean area, Quebec; Can. J. Earth Sci., v. 10, p. 881-899.

Fuh, Tsu-Min
1970: Correlation of rocks across the Grenville Front near Val d'Or, Quebec; unpubl. Ph.D. thesis, Queen's Univ., Kingston, Ontario.

Hewitt, D.F.
1956: The Grenville Region of Ontario, *in* 'The Grenville Problem', ed. J.B. Thomson; Roy. Soc. Can., Spec. Publ. 1, p. 22-41.

Katz, M.
1969: Geology of the Saint Patrice Lake and Portage-du-Fort areas; Que. Dept. Nat. Resour., Prelim. Rept. 578, 27 p.

Kretz, R.
1957: Pontefract-Gilles area, Pontiac Electoral District; Que. Dept. Mines, Prelim. Rept. 357, 7 p.

Krogh, T.E. and Davies, G.L.
1973: The effect of regional metamorphism on U-Pb systems in zircon and a comparison with Rb-Sr systems in the same whole-rock and its constituent minerals; Carnegie Inst. Yearbk., 72, p. 601-610.

Krogh, T.E. and Hurley, P.M.
1968: Strontium isotope variation and whole-rock isochron studies, Grenville Province of Ontario; J. Geophys. Res., v. 73, p. 7107-7125.

Lumbers, S.B.
1971: Geology of the North Bay area, District of Nipissing and Parry Sound; Ont. Dept. Mines and North Affairs, Geol. Rept. 94, 104 p.

Lyall, H.B.
1957: Hainaut-Champagne area, Pontiac Electoral District; Que. Dept. Mines, Prelim. Rept. 345, 8 p.

Marleau, R.-A.
1959: Perche-Poitou area, Pontiac Electoral District; Que. Dept. Mines, Prelim. Rept. 383, 8 p.

1960: Lorraine-Flandre area, Pontiac Electoral District; Que. Dept. Mines, Prelim. Rept. 420, 8 p.

Pirie, J.
1971: Mobilization in a migmatite from the Grenville Province, Quebec; unpubl. Ph.D. thesis, Queen's Univ., Kingston, Ontario, 163 p.

Pourret, G.
1968: Etude pétrographique et structural de la série crystallophyllienne de Grenville, Région du Lac McGillivray, Province de Québec, Canada; Thèse de Doctorat de 3ième cycle, Grenoble, France, 122 p.

Pourret, G. and Bergeron, R.
1970: Les schistes cristallins de la région du Lac McGillivray; Can. J. Earth Sci., v. 7, p. 1109-1116.

Rive, M.
1970: Geology of the Rowanton-Maganasipi east area; Que. Dept. Nat. Resour., Prelim. Rept. 586, 12 p.

1972: Geology of the Dumoine Lake area; Que. Dept. Nat. Resour., Prelim. Rept. 604, 19 p.

Sabourin, R.-J.-E.
1963: Geology of the Bourbonnais-Limousin area; Que. Dept. Nat. Resour., Prelim. Rept. 509, 7 p.

Silver, L.T. and Lumbers, S.B.
1965: Geochronologic studies in the Bancroft-Madoc area of the Grenville Province, Ontario, Canada; Geol. Soc. Am., Spec. Paper 87, p. 156.

Wilson, M.E.
1924: Arnprior-Quyon and Maniwaki areas, Ontario and Quebec; Geol. Surv. Can., Mem. 136, 152 p.

Wynne-Edwards, H.R.
1972: The Grenville Province, *in* 'Variations in Tectonic Styles in Canada', ed. R.A. Price, Geol. Assoc. Can., Spec. Paper 11, p. 263-334.

Wynne-Edwards, H.R., Gregory, A.F., Hay, P.W., Giovanella, C.A., and Reinhardt, E.W.
1966: Mont Laurier and Kempt Lake map-areas, Quebec; Geol. Surv. Can., Paper 66-32, 32 p.

BOREHOLE DENSITY LOGGING IN PERMAFROST, TUKTOYAKTUK, DISTRICT OF MACKENZIE

Project 730006

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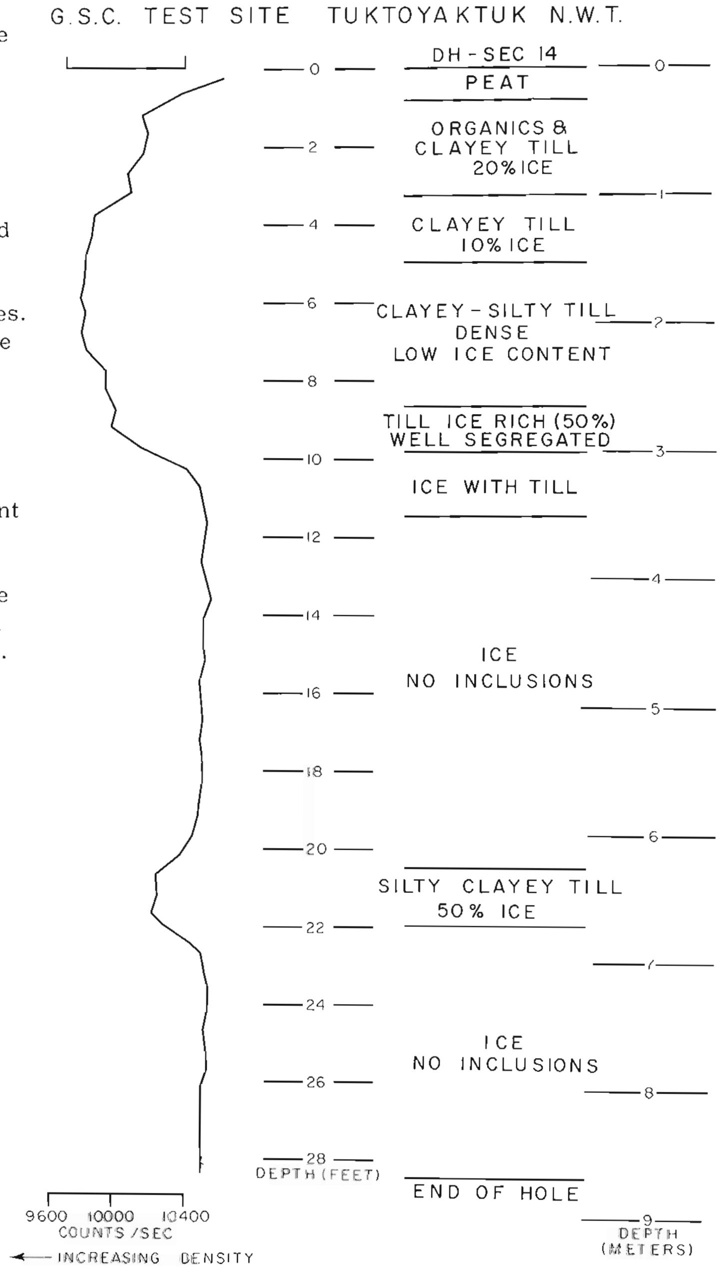
Recent Geological Survey shallow dry-hole drilling was completed at a test site near Tuktoyaktuk to provide detailed geological control for geophysical experiments. These holes provided an opportunity to test the density logging technique in ice-rich overburden materials.

Six shallow holes, with depths ranging up to 9.5 m were logged with a 1 15/16 -inch gamma-gamma tool using a Gearhart-Owen portalogger unit. A good example of the results is shown in Figure 88.1 compared with the geological description. To obtain better depth resolution the chart recorder was manually driven and measurements were made at 6-inch intervals up the holes. Since the tool has not been calibrated it was not possible to obtain absolute values of density; density variation is given in counts per second from the gamma-ray detector with high count rates indicative of low density. Good correlations exist between count rate variation and the change of ice content as given by the goelogy to a resolution exceeding 0.5 m. Variation in ice content of unconsolidated materials at low permafrost temperatures is probably the major factor responsible for variations in density in the Tuktoyaktuk area; hence it is suggested that the density logging tool can be used reliably to detect the presence of high ice content zones.



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ICE RECONNAISSANCE ON THE BEAUFORT SEA

Project 750029

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Since 1961, three attempts to define geology across the Canadian Arctic coastal and shelf regions by the seismic method have met with limited success. These attempts are described by Hobson and Overton, 1967; Overton, 1970, and Berry and Barr, 1971.

Each of these attempts has been hampered to some extent by problems of navigation and/or ice instability so that the Canadian Arctic coastal and shelf regions remain a seismic desert. It is the aim of this project to gain proficiency at seismic prospecting to study the regional and structural geological problems known to exist from reconnaissance magnetic and gravity surveys in this last Canadian frontier.

In view of recommendations originating from the Institute of Sedimentary and Petroleum Geology and the Earth Physics Branch favouring an extension of seismic techniques developed in the Sverdrup Basin onto the Beaufort Sea, an ice stability study using ERTS imagery and a helicopter reconnaissance in April, 1975, as conducted across the Beaufort Sea. These studies lead to the conclusion that ice instability in the shear zone of the Beaufort Sea precludes the use of the seismic methods used in the Sverdrup Basin. In April, 1975, an expanse of water at least 56 km wide was observed in the active shear zone. ERTS photographs show that wide leads in the shear zone are present in other years suggesting that highly unstable ice conditions prevail during the spring months every year.

References

Berry, M.J. and Barr, K.G.
1971: A seismic refraction profile across the Polar Continental Shelf of the Queen Elizabeth Islands; Can. J. Earth Sci., v. 8, no. 3, p. 347-360.

Hobson, G.D. and Overton, A.
1967: A seismic section of the Sverdrup Basin, Canadian Arctic Islands; Seismic refraction prospecting, Soc. Explor. Geoph., p. 550-562, ed. A.W. Musgrave.

Overton, A.
1970: Seismic refraction surveys, western Queen Elizabeth Islands and Polar Continental Margin; Can. J. Earth Sci., v. 7, no. 2, p. 346 365.

ARCTIC OCEAN: LOW-LEVEL AEROMAGNETIC PROFILES OBTAINED IN 1975

Project 650007

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The co-operative aeromagnetic project with the National Aeronautical Establishment was continued during 1975 and survey operations were carried out in the high Arctic during the period April 1-12, 1975. The North Star aircraft of the National Aeronautical Establishment which is equipped with an inboard digital-recording cesium magnetometer system was used as the survey platform. The flight elevation was maintained at 300-m (1000 feet) above sea level along the survey profiles flown. The primary navigation system used was a GNS 200 VLF navigation system and a Doppler radar whose outputs were fed to an Interdata minicomputer which calculated the best least-squares position and relayed the latitude and longitude of the aircraft to a Cypher digital magnetic tape recorder along with the total field values and time from a digital clock.

There were several objectives for the field operation in the high Arctic. The first was to carry out an aeromagnetic reconnaissance of the continental shelf of northern Ellesmere Island and Nansen Sound to ascertain whether a substantial thickness of sediments exists in those areas which would make them of interest for petroleum exploration. A second objective was to obtain magnetic survey evidence to evaluate the hypothesis that the Alpha Cordillera is part of the global midoceanic ridge system. In addition, the 1975 operation was an exercise which would provide experience to those involved in the logistical requirements for further airborne surveys in the Arctic Ocean using the new Convair 580 aircraft of the National Aeronautical Establishment, which is the replacement for the North Star aircraft.

Figure 90.1 shows the aeromagnetic profiles obtained in the Arctic Ocean between Ellesmere Island and the North Pole together with the actual track of the survey aircraft and the bathymetry.

The profiles obtained across the continental shelf of northeastern Ellesmere Island have the least magnetic expression of any recorded during the 1975 field operation. This is presumably due to a significant thickness of sedimentary rocks which occurs on this part of the polar continental shelf. The profiles with least magnetic expression occur near the entrance to Robeson Channel, and are evidence that the thick sequence of Lower Paleozoic rocks which strike in a northeast direction across northern Ellesmere Island extend for a considerable distance offshore. Depth determinations carried out on the profiles indicate that sediment thickness exceeds 20 000 feet (6 km) over wide areas of the Ellesmere continental shelf. Near Cape Fanshawe Martin a series of sharp anomalies occur which exceed 500 gammas from peak to trough. These anomalies may be caused by the presence of intrusive rocks similar to those onshore (norite and peridotite) which have been mapped by Christie (1957).

A distinctive double anomaly some 500 gammas or so in amplitude, occurs on a number of the profiles close to the edge of the continental shelf. Typically the peak of the anomaly is landward of the 500-m bathymetric contour and its associated trough is located in deep water at the bottom of the continental slope, usually near the 1000-m bathymetric contour. This diagnostic anomaly appears to be the well-known edge-effect anomaly which marks the transition from oceanic to continental crust. A similar type of anomaly occurs in the Labrador Sea and Baffin Bay (Hood and Bower, 1973).

Offshore from the edge of the continental shelf the wavelengths of the anomalies are quite variable, and some of the shorter wavelength anomalies are probably due to seamounts e.g. the anomaly at 85°10'N, 62°W on Profile IJ which was flown on April 8, 1975, and ends at the North Pole.

The profiles contained in the area MNOP were flown to obtain the magnetic expression of the Alpha Cordillera. In general, there is a central positive anomaly on each of the profiles, 600 gammas in amplitude, although the correlation from profile to profile is somewhat tenuous. It has been postulated by Vogt and Ostenso (1970) that the Alpha Cordillera is a fossil mid-ocean ridge that became inactive in middle Tertiary times. However Herron *et al.* (1974) pointed out that the depth of the Alpha Cordillera below sea level was inconsistent with the age versus depth curve of other sea-floor spreading ridges and they concluded that it was a fossil subduction zone-incipient island arc. There does not seem to be much evidence of symmetry in the profiles obtained over the Alpha Cordillera although the anomalies on the flank (Profile FP) are very striking indeed exceeding 1500 gammas in amplitude cf. the Reykjanes Ridge anomaly of about 2000 gammas. Such anomalies, which have a wavelength of about 15 km, are probably due to fairly recent intrusive activity in the area. This inference results from the fact that the natural remanent magnetization of the mid-Atlantic ridge anomaly is known to decay fairly rapidly with time.

An anomaly some 75 km wide and 500 gammas in amplitude occurs on Profile IJ in the central part of the Lomonosov Ridge. Churkin (1973) postulated that the Lomonosov Ridge was a narrow fragment of Eurasia that has drifted into the centre of the Arctic Ocean as the Eurasian Basin spread. The short wavelength character of the profile across the Ridge certainly indicates that the feature has a continental origin, and is in striking contrast to the rest of the profile to the pole which is relatively flat especially over the deeper water of the Eurasian Basin.

The large (1500 gamma) long-wavelength anomaly on Profile KL at 88°N is an intriguingly-shaped anomaly

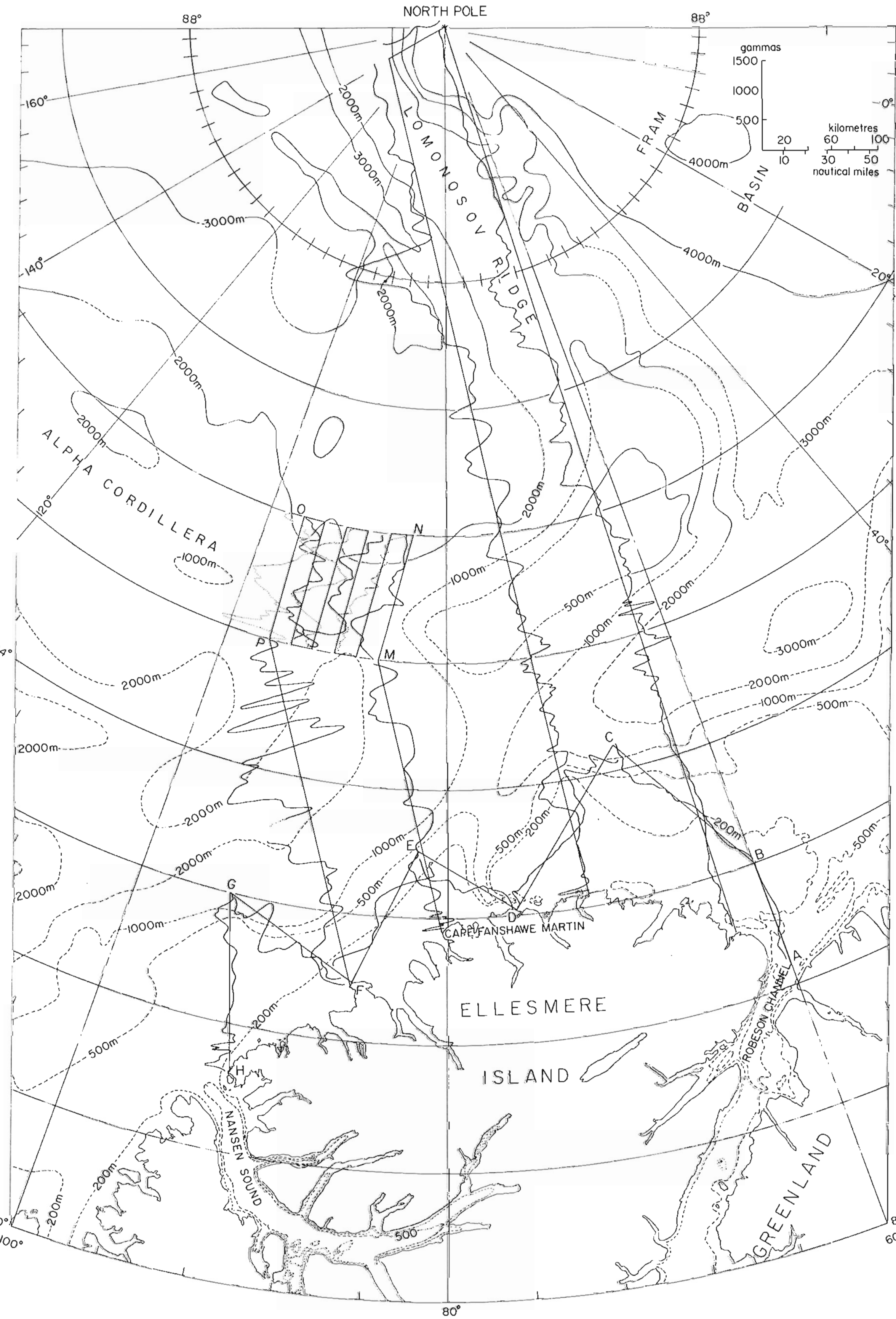

Figure 90. 1. Aeromagnetic profiles obtained in the Arctic Ocean during 1975. The bathymetric contours are in metres.

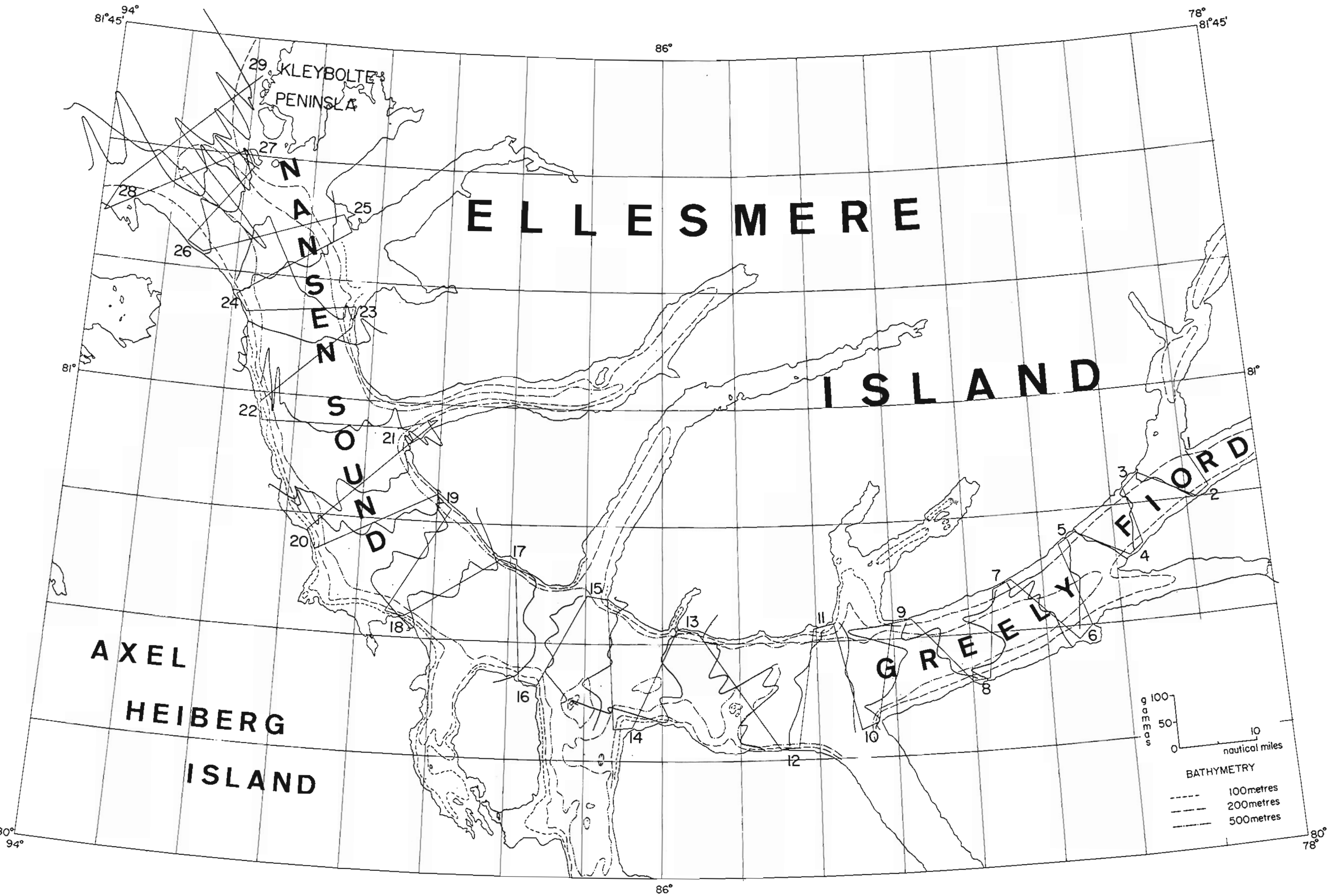

Figure 90. 2. Aeromagnetic profiles obtained in Nansen Sound and Greely Fiord during 1975. The bathymetric contours are in metres.

being located on the flank of the Lomonosov Ridge. Its apparent wavelength may however be exaggerated due to the fact that the profile is subparallel to the strike of the Lomonosov Ridge.

Nansen Sound

The low-level aeromagnetic profiles obtained in Nansen Sound (and Greely Fiord) are shown in Figure 90. 2, and these have been plotted with the regional gradient removed. The bathymetry has also been plotted from the information contained on maps published by the Canadian Hydrographic Service using contour intervals of 100, 200 and 500 m. In general the water depth increases rapidly from the shoreline of the Sound, particularly on the northern side where the land rises sharply usually as a cliff face which is in contrast to the much flatter southern shoreline. From a consideration of this morphological evidence, coupled with the arcuate shape of the Sound, it appears probable that Nansen Sound is a half-graben. There is a considerable amount of support for this hypothesis from the aeromagnetic profiles specifically Profiles 16-17-18-19-20-21 in the steep gradient at the north shore which appears to be produced by downfaulted rock formations. However only one profile has the diagnostic U-shaped signature indicative of a graben and this is Profile 21-22 cf. the magnetic signature over the Melville Bay graben (Hood and Bower, 1973).

Nansen Sound and Greely Fiord are bounded on either side by sedimentary rocks of Cenozoic, Mesozoic and Paleozoic age, so it is to be expected that sedimentary rocks underlie the waters of the sound itself. Depth determinations on the anomalies indicate that the thickness of sediments exceeds 10 000 feet (3 km) under Nansen Sound.

At the entrance to Nansen Sound, the magnetic signature along Profile 28-29 contains sharp, large amplitude anomalies and the maximum peak to trough excursion recorded is 4250 gammas. Onshore on Kleybolte Peninsula, Ordovician volcanics have been mapped by Trettin (1969), and it seems probable that these rocks extend across the entrance to the sound and produce the distinctive anomalies in question. However in view of the fact that the depth of water approaches 500 m, and the survey aircraft flew at 300 m, i.e. a minimum separation between magnetometer and causative body of 800 m, there is a distinct possibility that iron-formation underlies the entrance to Nansen Sound.

Acknowledgments are made to the aircrew of the North Star aircraft who carried out their duties in their usual conscientious way: - Pilot: Captain W.T. Chevrier; co-pilot: Benjamin Budgeon; navigator: Captain T.R. Brownley, and flight engineer: O.H. Stevens. Acknowledgment is also made to B.W. Leach, J. Jordan and N. McPhee of the National Aeronautical Establishment who took part in the field operation; C.D. Hardwick also made a significant contribution in the necessary preparations for the 1975 aeromagnetic survey operation.

References

Christie, R.L.
1957: Geological reconnaissance of the north coast of Ellesmere Island, District of Franklin, Northwest Territories; Geol. Surv. Can., Paper 56-9, 40 p.

Churkin, M.
1973: Geologic concepts of Arctic Ocean basin, in Arctic Geology; Am. Assoc. Pet. Geol., Mem. 19, p. 485-499.

Hood, P. and Bower, M.E.
1973: Low-level aeromagnetic surveys of the continental shelves bordering Baffin Bay and the Labrador Sea, *in* Earth Science Symposium on offshore Eastern Canada; Geol. Surv. Can., Paper 71-23, p. 573-598.

Trettin, H.P.
1969: Pre-Mississippian geology of northern Axel Heiberg and northwestern Ellesmere Islands, Arctic Archipelago; Geol. Surv. Can., Bull. 171, 82 p.

Vogt, P.R. and Ostenso, N.A.
1970: Magnetic and gravity profiles across the Alpha Cordillera and their relation to Arctic sea-floor spreading; J. Geophys. Res., v. 75, p. 4925-4937.

91. GEOLOGICAL INVESTIGATIONS IN NORTHERN ELLESMERE ISLAND (Reports 91 to 96) INTRODUCTION

Project 730051

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Location and Accessibility of Project Area

The project area (Fig. 91.1) comprises all those parts of northern Ellesmere Island lying north and east of the region covered by recent geological maps at the 1: 250 000 scale (Thorsteinsson, 1972; Thorsteinsson and Kerr, 1972; Thorsteinsson and Trettin, 1971, 1972a, b; Kerr, 1973) and also marginal parts of the recently mapped areas where major revisions are necessary. Northwestern and eastern Judge Daly Promontory (Christie, 1974), key areas that require a special investigation, however, are excluded from the present study.

The Defence Research Board base at Tanquary Fiord, used in 1975, lies about 840 km (520 miles) north of Resolute Bay, the nearest airline terminal, and 225 km (140 miles) north of Eureka weather station, a terminal for sealift shipments.

Previous Geological Investigations

Initial reconnaissance work by a British naval expedition under Sir George Nares (Feilden and de Rance, 1878), the Danish geologist J.C. Troelsen (1950), R.G. Blackadar (1954), and R.L. Christie has been incorporated in comprehensive reports by Christie (1957, 1964). Geological studies in adjacent parts of Ellesmere Island, carried out by R. Thorsteinsson (1974), E.T. Tozer (1963 and *in* Douglas, 1970, p. 574-589), J.W. Kerr (1967, 1968), and H.P. Trettin (1969a) mainly in 1961-62 established a stratigraphic framework that is, to some extent, applicable to the present area.

A series of special reports resulted from Operation Grant Land, a reconnaissance project of the Geological Survey organized by R.L. Christie in 1965-66 with some follow-up work in later years. It includes:

- a study of lower Paleozoic and Proterozoic metamorphic and plutonic rocks in northernmost Ellesmere Island by T.O. Frisch (1974) supplemented by recent isotope determinations by A.K. Sinha and T.O. Frisch (1974, 1975);
- studies of lower Paleozoic (and Pennsylvanian) rocks in various parts of the project area by H.P. Trettin (1969b, 1971, 1972);
- a study of Paleozoic and Mesozoic strata at Yelverton Pass by W.W. Nassichuk and R.L. Christie (1969); and
- a study of Mesozoic and Tertiary rocks at Lake Hazen by A.A. Petryk (1969).

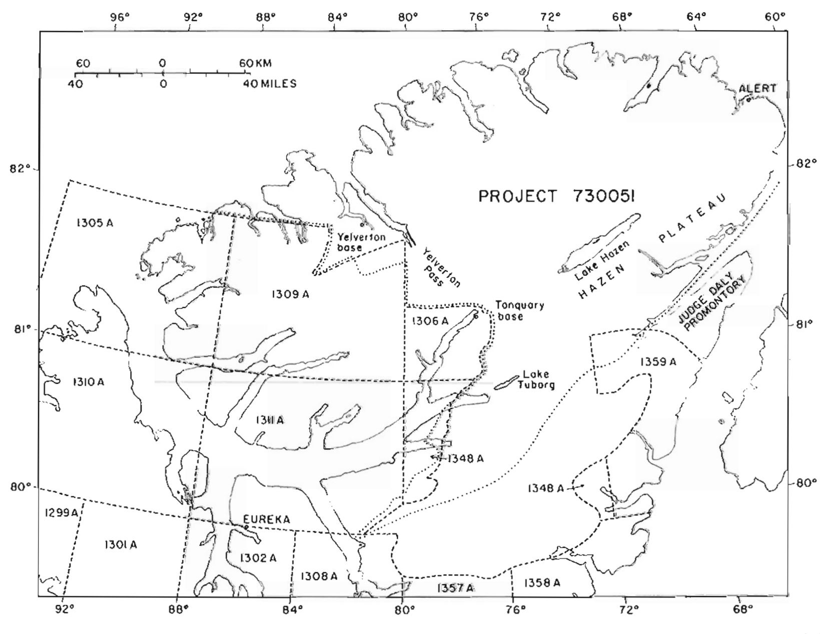

Figure 91.1.

Index showing project area and recently published, A-series geological maps. (Only those maps adjacent to project area are listed under references.)

These reports are accompaned by specialized geological maps at various scales. Information obtained prior to 1973 was summarized on preliminary drafts of 1: 1 000 000 atlas sheets with accompanying notes, correlation charts, etc. (Trettin, 1973). The data, however, were insufficient to prepare standard geological maps and a coherent stratigraphic framework for this very complex region.

Objectives and Method of Current Project

The objectives of Project 730051 are to produce standard geological maps at 1: 250 000 or more detailed scales where required, to establish a coherent stratigraphic framework, and to produce terminal reports dealing with most aspects of the bedrock geology. Some parts of the project area will require complete study or revision whereas others will require only minor revisions. Much of the work necessarily will remain of a reconnaissance nature but detailed research will be done wherever required to solve basic problems of the regional geology.

It is hoped that this can be achieved by four staff geologists in two seasons using one helicopter and some fixed-wing support with subsequent follow-up work at lesser expense. The first main phase of field work, concerned with the western parts of the project area, took place in 1975; the second, to deal with eastern parts, is scheduled for 1977.

Geological Responsibilities

The writer is responsible for the project as a whole. The responsibility for different aspects of the regional geology has been divided as follows:

Igneous intrusions and metamorphic terranes .... T.O. Frisch

Geology of lower Paleozoic sedimentary and volcanic formations .... H.P. Trettin

Overall responsibiltiy for upper Paleozoic, Mesozoic, and Tertiary geology and special studies of upper Paleozoic stratigraphy... U. Mayr

Mesozoic and Tertiary stratigraphic sections .... D.G. Wilson

The standard geological maps will be produced jointly but reports dealing with the four major topics outlined above will be prepared independently by the four participants who planned and executed their work independently.

General Account of 1975 Field Work

In addition to the staff geologists mentioned, the party included John Calvert, graduate student assistant and mountaineering expert, Mark Oliver, student assistant, Bradley Bossort, camp manager, and Jerry Diakowicz, cook.

The field work was done with a Bell G 4A helicopter leased from Viking Helicopters Ltd., Ottawa, flown by Earl Lozo with David Lamaanen as engineer and alternate pilot. The machine was adequate at low and intermediate altitudes but useful at higher altitudes only owing to Lozo's skill and experience.

The party left Calgary on June 3rd and returned on August 22nd, the geological field work having extended from June 6th to August 19th. Base camp was at the head of Tanquary Fiord from June 4th to July 7th, southwest of Yelverton Bay from July 8th to August 10th (Fig. 91.2), and again at Tanquary Fiord from August 11th to 21st. At Tanquary Fiord, facilities of the Defence Research Board were made available to us through the courtesy of H. Serson.

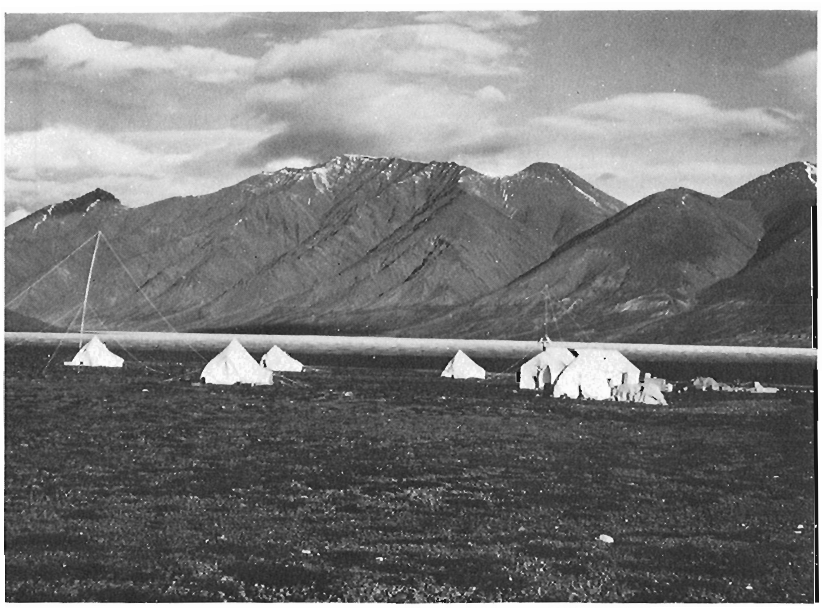

Figure 91.2.

Base camp at Kulutingwak Fiord south of Yelverton Bay; view to the northeast. In the background flysch sediments of Imina Formation (Silurian and (?) uppermost Ordovician) metamorphosed in greenchist facies.

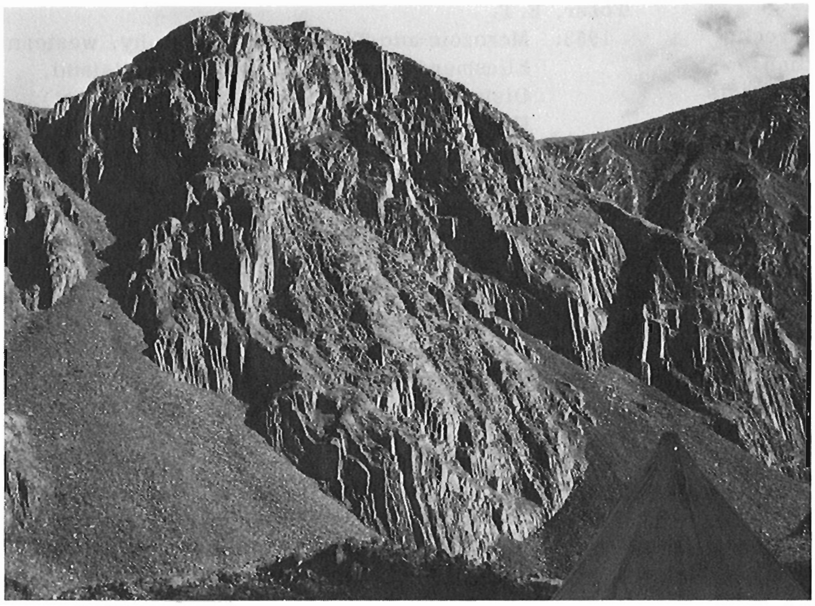

Figure 91. 3.

Fly camp on Hazen Plateau, northeast of Lake Tuborg; view to the east. Steeply dipping strata of unmetamorphosed Imina Formation (Silurian) were studied for paleocurrent directions.

Moves from Resolute or Eureka to Tanquary Fiord were carried out by DC 3 and flights to the Yelverton base camp by Twin-Otter, single-engine Otter and Beaver. Fifty hours of flying time on the single-engine Otter was allotted to the project by G. D. Hobson, Coordinator, Polar Continental Shelf Project; the P. C. S. P. also is thanked for expediting services at Resolute.

The field work was done both out of the base camp with the helicopter and on foot out of fly camps (Fig. 91. 3); fly camping was especially important in the Yelverton Inlet region where the weather was very unstable.

The geological results obtained are summarized by Frisch, Mayr, Wilson, and the writer in separate contributions to this volume.

References

Blackadar, R. G.
1954: Geological reconnaissance, north coast of Ellesmere Island, Arctic Archipelago, Northwest Territories; Geol. Surv. Can., Paper 53-10.

Christie, R. L.
1957: Geological reconnaissance of the north coast of Ellesmere Island, District of Franklin, Northwest Territories; Geol. Surv. Can., Paper 56-9.

1964: Geological reconnaissance of northeastern Ellesmere Island, District of Franklin; Geol. Surv. Can., Mem. 331.

1974: Northeastern Ellesmere Island: Lake Hazen region and Judge Daly Promontory; *in* Report of Activities, Part A, Geol. Surv. Can., Paper 74-1A, p. 297-299.

Douglas, R. J. W. (ed.)
1970: Geology and economic minerals of Canada; Geol. Surv. Can., Econ. Geol. Rep. no. 1.

Feilden, H. W. and de Rance, C. E.
1878: Geology of the Coasts of the Arctic lands visited by the late British expedition under Captain Sir George Nares; Quart. J. Geol. Soc. London, v. 34, p. 556-567.

Frisch, T. O.
1974: Metamorphic and plutonic rocks of northernmost Ellesmere Island, Canadian Arctic Archipelago; Geol. Surv. Can., Bull. 229.

Kerr, J. W.
1967: Stratigraphy of central and eastern Ellesmere Island, Arctic Canada. Part I. Proterozoic and Cambrian; Geol. Surv. Can., Paper 67-27, Part I.

1968: Stratigraphy of central and eastern Ellesmere Island, Arctic Canada. Part II. Ordovician; Geol. Surv. Can., Paper 67-27, Part II.

1973: Geology, Kennedy Channel, District of Franklin; Geol. Surv. Can., Map 1359A.

Nassichuk, W. W. and Christie, R. L.
1969: Upper Paleozoic and Mesozoic stratigraphy in the Yelverton Pass region, Ellesmere Island, District of Franklin; Geol. Surv. Can., Paper 63-39.

Petryk, A. A.
1969: Mesozoic and Tertiary stratigraphy at Lake Hazen, northern Ellesmere Island, District of Franklin; Geol. Surv. Can., Paper 68-17.

Sinha, A.K. and Frisch, T.O.
1974: Whole-rock Rb/Sr ages of metamorphic rocks from northern Ellesmere Island, Canadian Arctic Archipelago; Geol. Soc. Am., abstracts with programs, v. 6, no. 7, p. 1064-1065.

1975: Whole-rock Rb/Sr ages of metamorphic rocks from northern Ellesmere Island, Canadian Arctic Archipelago, I. The gneiss terrain between Ayles Fiord and Yelverton Inlet; Can. J. Earth Sci., v. 12, p. 90-94.

Thorsteinsson, R.
1971: Geology, Greely Fiord west, District of Franklin; Geol. Surv. Can., Map 1311A.

1972: Geology, Cañon Fiord, District of Franklin; Geol. Surv. Can., Map 1308A.

1974: Carboniferous and Permian stratigraphy of Axel Heiberg Island and western Ellesmere Island, Canadian Archipelago; Geol. Surv. Can., Bull. 224.

Thorsteinsson, R. and Kerr, J.W.
1972: Geology, Greely Fiord east, District of Franklin; Geol. Surv. Can., Map 1348A.

Thorsteinsson, R. and Trettin, H.P.
1971: Geology, Tanquary Fiord, District of Franklin; Geol. Surv. Can., Map 1306A.

1972a: Geology, Cape Stallworthy, District of Franklin; Geol. Surv. Can., Map 1305A.

1972b: Geology, Otto Fiord, District of Franklin; Geol. Surv. Can., Map 1308A.

Tozer, E.T.
1963: Mesozoic and Tertiary stratigraphy, western Ellesmere Island and Axel Heiberg Island, District of Franklin (preliminary account); Geol. Surv. Can., Paper 63-30.

Trettin, H.P.
1969a: Pre-Mississippian geology of northern Axel Heiberg and northwestern Ellesmere Islands, Arctic Archipelago; Geol. Surv. Can., Bull. 171.

1969b: Geology of Ordovician to Pennsylvanian rocks, M'Clintock Inlet, north coast of Ellesmere Island, Canadian Arctic Archipelago; Geol. Surv. Can., Bull. 183.

1971: Geology of lower Paleozoic formations, Hazen Plateau and southern Grant Land Mountains, Ellesmere Island, Arctic Archipelago; Geol. Surv. Can., Bull. 203.

1972: Reconnaissance of lower Paleozoic geology, Phillips Inlet region, north coast of Ellesmere Island, District of Franklin; Geol. Surv. Can., Paper 71-12.

1973: Preliminary draft of 1: 1 000 000 geological atlas sheets, Eureka Sound and Robeson Channel area, Canadian Arctic Islands (NTS 560, 340 and Canadian part of 120); Geol. Surv. Can., Open File 174.

Troelsen, J.C.
1950: Contributions to the geology of northwest Greenland, Ellesmere Island, and Axel Heiberg Island; Medd. om. Grønland, v. 149, no. 7.

Project 730051

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Nearly four weeks were spent sampling gabbroic sills and dykes in upper Paleozoic and Mesozoic rocks of the area around Tanquary Fiord. Sills are particularly abundant but their cliff-forming nature, gentle dips and commonly strong alteration in an area of high relief combine to hamper sampling. The chilled margins and centres of 35 sills and 10 dykes were sampled for petrographic study, chemical analysis and K-Ar age determination. Little or no differentiation was seen in any sill or dyke and, except for variations in the abundance of plagioclase, the rocks appear to be remarkably uniform.

A further four weeks were taken up by studies of the chiefly metamorphic terrane between Ayles Fiord and Phillips Inlet on the north coast of Ellesmere Island (Fig. 92.1). Special emphasis was laid on mapping those areas least known geologically, such as Milne Fiord and the head of Petersen Bay. Figure 92.1 is a simplified sketch-map showing the gross geological features of the region investigated.

Granitoid gneisses and metasediments of late Precambrian age (Sinha and Frisch, 1975) form a curvilinear belt stretching from the west coast of Ayles Fiord to south of Phillips Inlet. The gneisses commonly

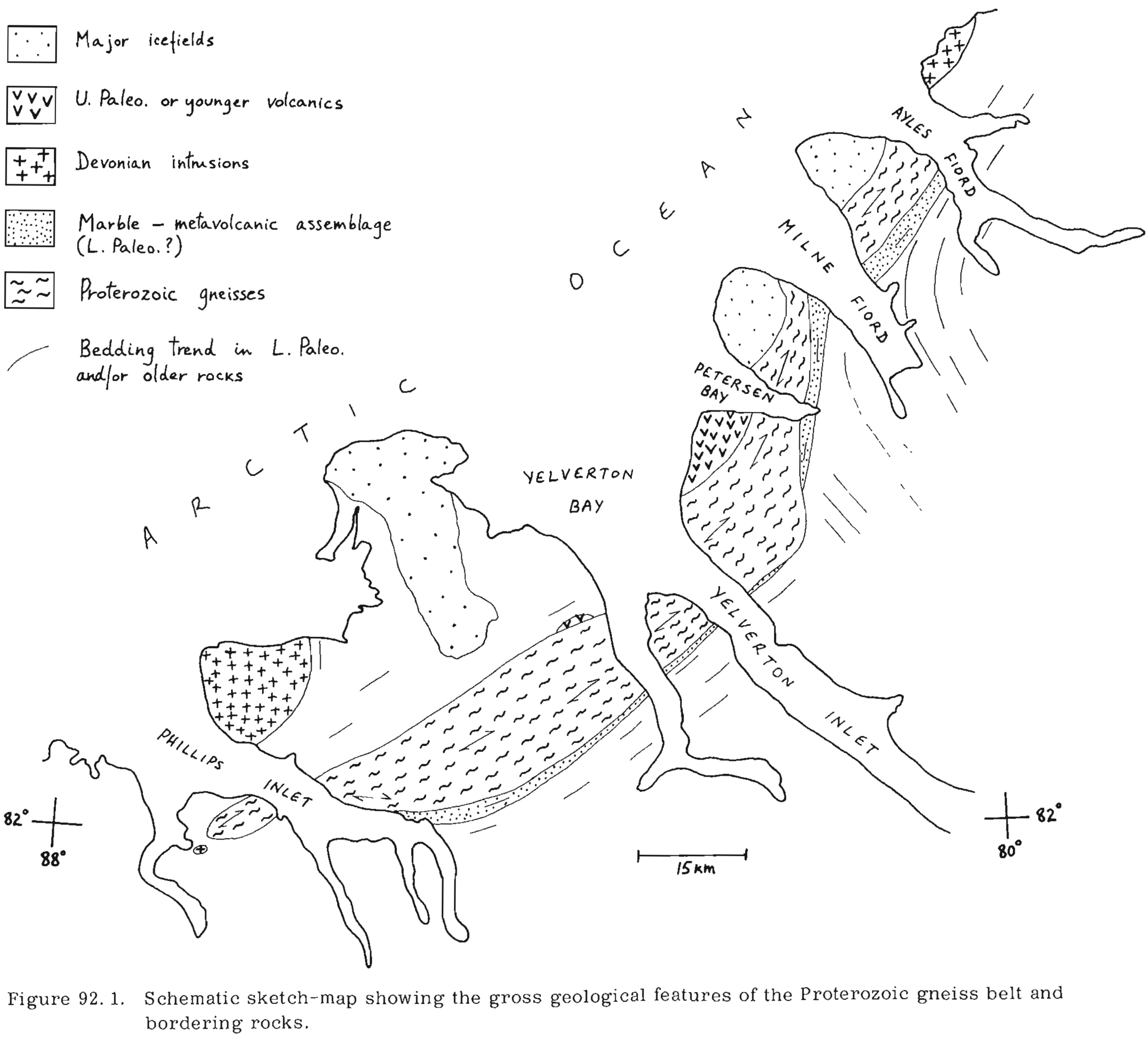

Figure 92.1. Schematic sketch-map showing the gross geological features of the Proterozoic gneiss belt and bordering rocks.

are either (garnet-) biotite-feldspar-quartz rocks with augen of feldspar or more leucocratic and homogeneous granitic rocks, at least some of which appear to be meta-intrusive. A younger, leucocratic, grey gneiss forms layers and veins in the augen gneiss. Concordant layers of amphibolite, 10 cm to several metres thick, are abundant locally in the gneisses. Fine grained margins and zonal distribution of recrystallized plagioglase phenocrysts in a number of amphibolites suggest an igneous intrusive origin for these rocks. Brown-weathering mica schists, some of which carry staurolite, kyanite and/or garnet occur among the gneisses and are interpreted as metasediments.

Bordering the gneissic belt to the south is a mixed marble and metavolcanic assemblage in which the metamorphic grade changes from amphibolite to greenschist southward over a distance of a few hundred metres. Near the contact, the gneisses are sheared, even mylonitized, and exhibit retrograde metamorphism. The adjacent metavolcanics include garnet amphibolites, cordierite-bearing schists and metasediments (some with staurolite or kyanite) indistinguishable from schists occurring within the gneisses. Associated marble on Milne Fiord contains tremolite. Farther south, the metavolcanics are represented by greenschist-grade chlorite schist, phyllite and metagabbro, and the marble is a weakly recrystallized limestone without tremolite. The marble-metavolcanic assemblage on Petersen Bay and Milne Fiord is correlated with similar rocks in the Yelverton Inlet and Phillips Inlet areas, suggested by Trettin (preceding paper) to be Ordovician in age. If this age assignment is correct, amphibolite-facies metamorphism occurred in early or middle Paleozoic time over a large area of northern Ellesmere Island.

Along the upper reaches of Milne Fiord and southeast of Petersen Bay, steeply-dipping, greenschist-grade carbonates, phyllites, slates and quartzites trend northwest before curving northeast to conform with the gneiss belt and the marble-metavolcanic assemblage (Fig. 92.1). South of Petersen Bay, the northwest trend is terminated abruptly by a northeast-striking fault. Repetition of lithologically similar strata suggests isoclinal folding about northwest-trending axes. These rocks are considered to be lower Paleozoic and/or older, pending isotopic age determinations and detailed lithologic comparisons with lower Paleozoic formations in the Yelverton Inlet and M'Clintock Inlet regions.

The area around Hansen Point is underlain by unmetamorphosed, chiefly intermediate to acid volcanics intruded by numerous basic dykes. The volcanics are grey on fresh surfaces, feldspar- and quartz-phyric, and locally vesicular to amygdaloidal. Alteration (typically producing zeolites) is common and exposure poor, rendering the recognition of primary structures difficult. However, columnar jointing was seen at one locality. The volcanics are correlated with those west of Yelverton Bay (Frisch, 1974) and tentatively suggested to be of late Paleozoic or younger age. At both localities, gneisses are thrust over the southern margin of the volcanic complex.

Aerial reconnaissance and sampling showed that, west of Yelverton Bay, the gneiss belt is flanked on both sides by low-grade lower Paleozoic and older (?) rocks. Gneissic and associated metamorphic rocks mapped by Trettin (1972) as units lPh (lower Paleozoic hybrid rocks) and Dqm (Devonian? granitic rocks) are now regarded as extensions of the Proterozoic gneiss belt.

References

Frisch, T.O.
1974: Metamorphic and plutonic rocks of northernmost Ellesmere Island, Canadian Arctic Archipelago; Geol. Surv. Can., Bull. 229.

Sinha, A.K. and Frisch, T.O.
1975: Whole-rock Rb/Sr ages of metamorphic rocks from northern Ellesmere Island, Canadian Arctic Archipelago. I. The gneiss terrain between Ayles Fiord and Yelverton Inlet; Can. J. Earth Sci., v. 12, p. 90-94.

Trettin, H.P.
1972: Reconnaissance of lower Paleozoic geology, Phillips Inlet region, north coast of Ellesmere Island, District of Franklin; Geol. Surv. Can., Paper 71-12.

93. RECONNAISSANCE OF LOWER PALEOZOIC GEOLOGY, AGASSIZ ICE CAP TO YELVERTON BAY, NORTHERN ELLESMERE ISLAND

Project 730051

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Institute of Sedimentary and Petroleum Geology, Calgary

In 1975, the lower Paleozoic geology in a strip about 260 km (161 miles) long and 90 to 130 km (56-81 miles) wide, extending from Agassiz Ice Cap to Yelverton Bay was investigated (Figs. 93. 1, 93. 10). Outcrop of lower Paleozoic strata in this belt is nearly continuous, except for a 32-km-(10-mile) wide belt between Tanquary Fiord and Yelverton Inlet that is covered by upper Paleozoic and Mesozoic strata (*see* Nassichuk and Christie, 1969, Fig. 4) and a 19-km-(12-mile) wide belt south of Yelverton Bay that is underlain by Proterozoic metamorphic rocks (Fig. 93. 11). Some additional work was done north and east of the head of Otto Fiord (areas 5 and 6 in Fig. 93. 10).

Stratigraphy and depositional history

One of the main objectives of the present project is to build a coherent stratigraphic framework that reflects the depositional history of the region; some progress in that direction was made in 1975 (*see* correlation chart, Fig. 93. 1). In the following brief summary, observations made in 1975 are integrated with previous work, both published and unpublished.

Proterozoic crystalline basement

The oldest unit in the present area comprises amphibolite-grade granitic gneisses and metasediments on the north coast of Ellesmere Island (Frisch, 1974 and this publication, report 92) that have yielded a whole-rock Rb/Sr isochrone age of 742 ± 12 m. y. regarded as a minimum by Sinha and Frisch (1975). The boundaries of the crystalline complex with the lower Paleozoic strata are faulted in the present area.

Grant Land Formation

The Grant Land Formation was established to comprise a widely distributed succession of variably feldspathic quartzite and green, red and minor grey slate with lesser amounts of granule and pebble conglomerate. Previously reported limestone and chert now are regarded as fault slices or synclinal keels of the Hazen Formation.

Outcrops discovered west of Yelverton Bay demonstrate that the Grant Land Formation overlapped the Proterozoic basement but the base of the Grant Land Formation has not been seen. The upper contact, with the Hazen Formation, placed at the lowermost carbonate beds within a slaty succession (Fig. 93. 2), is exposed at various localities and clearly is conformable.

A well-exposed section north of the head of Tanquary Fiord, and six detailed intervals in a very long section (to be measured photogrammetrically) north of the head of Otto Fiord were measured and sampled bed by bed in order to determine depositional environments and processes. Numerous fining-up sequences and medium-scale trough crossbeds confirm that most sandstones are fluvial in origin. Abundant tabular beds, which in Recent settings are characteristic of catastrophic floods in arid belts (e. g. McKee *et al.*, 1967) are attributed primarily to lack of vegetation in the early Paleozoic, the climate being unknown. Several units of dark grey slate, on the other hand, especially those underlying the Hazen Formation, are regarded as marine.

The unfossiliferous formation is considered tentatively as Lower Ordovician and possibly Upper Cambrian because it underlies strata of the Hazen Formation that are probably not older than Arenigian (late Early Ordovician).

The quartzites and conglomerates, which contain coarse grained microcline, probably were derived from gneissic terranes in the northern regions because correlative units to the southeast (Copes Bay and Parrish Glacier formations) are composed mainly of impure carbonate rocks. A marked coarsening to the northwest, however, was not observed. Paleocurrent studies in trough crossbedded strata require relatively soft, hoodoo-forming sediments that show the orientation of trough axes; the strongly indurated Grant Land quartzites, unfortunately, were unsuited for that purpose.

Hazen Formation

The name Hazen Formation was applied (Trettin, 1971) to a Lower to upper Middle Ordovician succession in the Hazen Plateau region that consists of carbonate strata, bedded chert, and dark grey slate with minor amounts of calcareous and dolomitic siltstone and lies stratigraphically between Grant Land and Imina formations (*see* correlation chart, Fig. 93. 1). Lithology and fauna (graptolites, radiolarians) indicated that all sediments, including the carbonates, were deposited in a relatively deep, starved basin, the Hazen Trough.

Current studies suggest that the definition of the formation should be expanded to include: (1) deep-water carbonate rocks, chert, and slate of Ordovician to late Early Silurian age underlying the Hazen Formation between Hazen Plateau and northwestern limit of the miogeocline (defined by the northwestern facies boundary of Cornwallis Group and Allen Bay and Read Bay formations); and (2) similar beds at Yelverton Inlet occurring stratigraphically between the Grant Land Formation and volcanic unit of Yelverton complex.

The outcrop area of the (expanded) Hazen Formation is divisible into four facies.

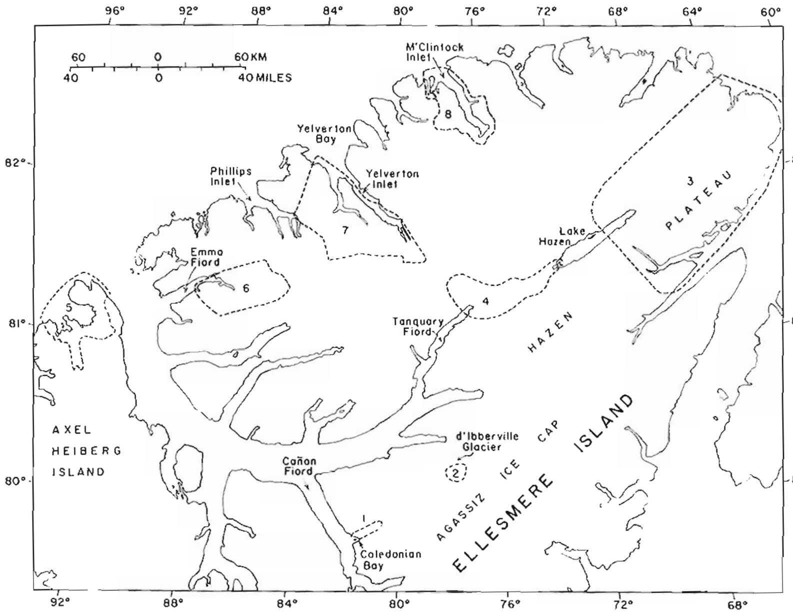
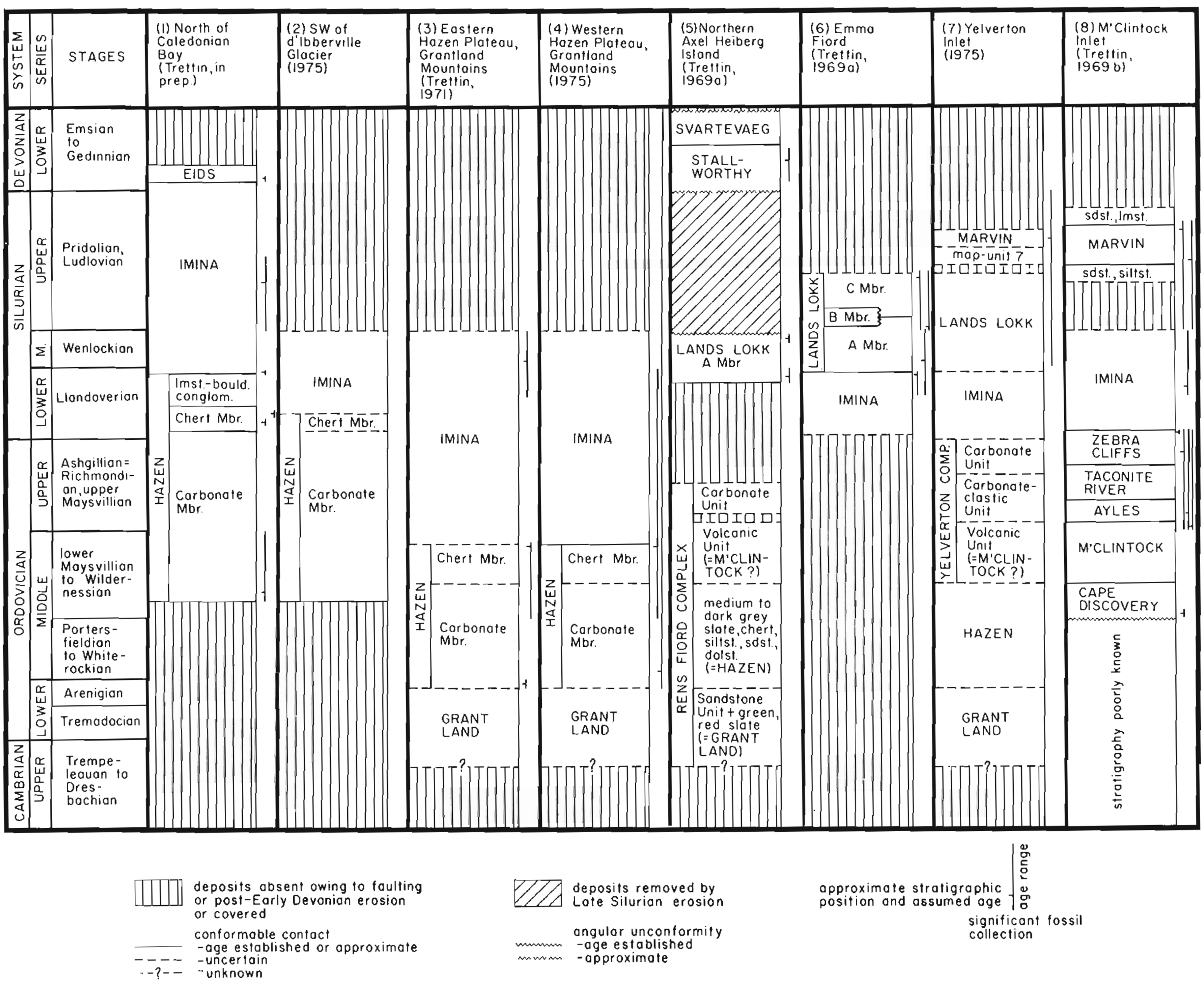


Figure 93. 1

Provisional correlation chart and index.



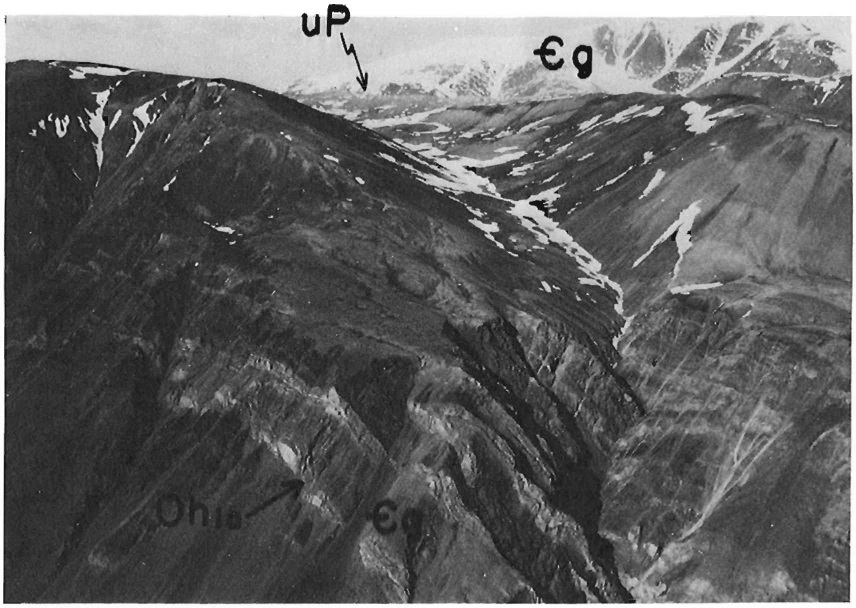


Figure 93.2

Foreground: slates of upper Grant Land Formation (Єg) and basal sub-unit of Hazen Formation (OHla) involved in complex folds and faults. Background: resistant Grant Land quartzites (Єg) thrust over upper Paleozoic strata (uP) that unconformably overlie lower Paleozoic units.

Facies A is represented by a section north of Caledonian Bay (Fig. 93.4, loc. 1) measured in 1972 (Trettin, in prep.). Three major units are recognized from top to base: (1) a limestone boulder conglomerate containing shelf- or reef-derived fossils of contemporaneous age (27 m); (2) radiolarian chert with minor amounts of thinly interstratified slate and redeposited carbonate sediments (120 m); and (3) a lower carbonate member composed of thinly interlaminated lime mudstone and claystone, laminated and graded lime wackestone and packstone, two major limestone boulder conglomerates one of which contains shelf-type fossils, primary and replacement chert, and dark grey, graptolitic slate (minimum thickness 260 m).

The conglomerates are interpreted as mass flow deposits, the graded and laminated wackestones and packstones as turbidites *sensu stricto*, and the laminated lime mudstones as deposits of very dilute turbid flows. All these carbonate sediments must have been derived from the adjacent miogeoclinal shelf; the associated, very fine grained terrigenous sediment probably came from the Canadian Shield.

The Hazen Formation is overlain by upper Llandoverian beds of the Imina Formation; its base is not exposed. The oldest diagnostic fossils, occurring about 40 m above the base of the section, are late Middle Ordovician conodonts (identification by T.T. Uyeno).

Facies B is known from a section southwest of d'Iberville Glacier (Fig. 93.4, loc. 2). This section, also overlain by the Imina Formation, is divisible into an upper chert member (about 22 m thick) and a lower carbonate member (minimum thickness 275 m), the base of which is not exposed. The carbonate member consists mainly of thinly laminated, slightly silty and argillaceous lime mudstone (carbonate particles are mostly of fine silt grade) with small amounts of thinly interlaminated calcareous siltstone or mudstone. Very subtle graded bedding on a millimetre scale is apparent in some laminae. Sand-grade skeletal packstone, composed mostly of echinoderm fragments, is rare. The phenoclasts in two limestone cobble conglomerates near the base of the section are composed of unfossiliferous dark grey lime mudstone. The lime mud in this facies also was derived from the miogeocline, but the scarcity of associated sand-grade carbonate sediment and the absence of shelf-derived conglomerate indicate deposition at a greater distance from the shelf margin than in facies A: the cobble conglomerates probably developed from slumped slope deposits.

Facies C underlies the Hazen Plateau and is represented by the type section (Fig. 93.4, loc. 3; Trettin, 1971, p. 40) and two sections between Lake Hazen and Tanquary Fiord (Fig. 93.1 and Fig. 93.4, locs. 4, 5). Throughout the Hazen Plateau region, an upper chert member and a lower carbonate member can be distinguished (Fig. 93.5). The upper member consists mainly of chert with lesser amounts of dark grey slate and relatively few carbonate turbidites and breccias. Chert is more abundant in the northwest (locs. 4, 5) than in the southeast (loc. 3). The member is 216 m thick at locality 3, 183 m at locality 4, and between 184 and 216 m at locality 5. The carbonate member is composed chiefly of limestone with lesser quantities of thinly interstratified slate, chert, dolostone, calcareous and dolomitic siltstone, and very fine grained sandstone. The limestone is variably silty and argillaceous and shows horizontal lamination (Fig. 93.3), small-scale cross-lamination and, less commonly, graded bedding. Some rocks are brecciated. A basal sub-unit (designated Ohla in Fig. 93.2), 12 to 19 m thick, is recognized between Tanquary Fiord and Lake Hazen. It is thicker bedded (beds up to 30 cm), more resistant to weathering, and structurally more competent than the rest of the member. The thickness of the carbonate member is difficult to establish owing to poor

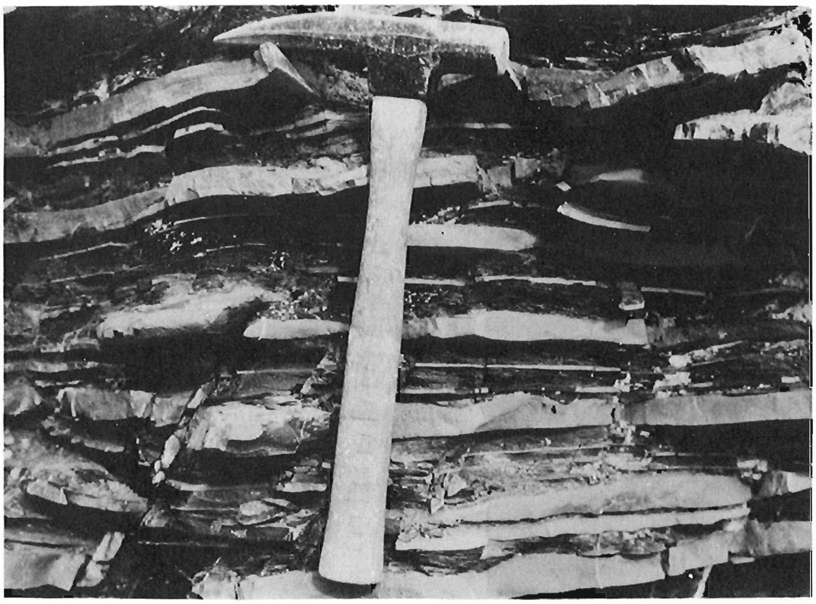


Figure 93. 3

Thinly interstratified limestone and mudstone in carbonate member of Hazen Formation at loc. 4, Figure 4.

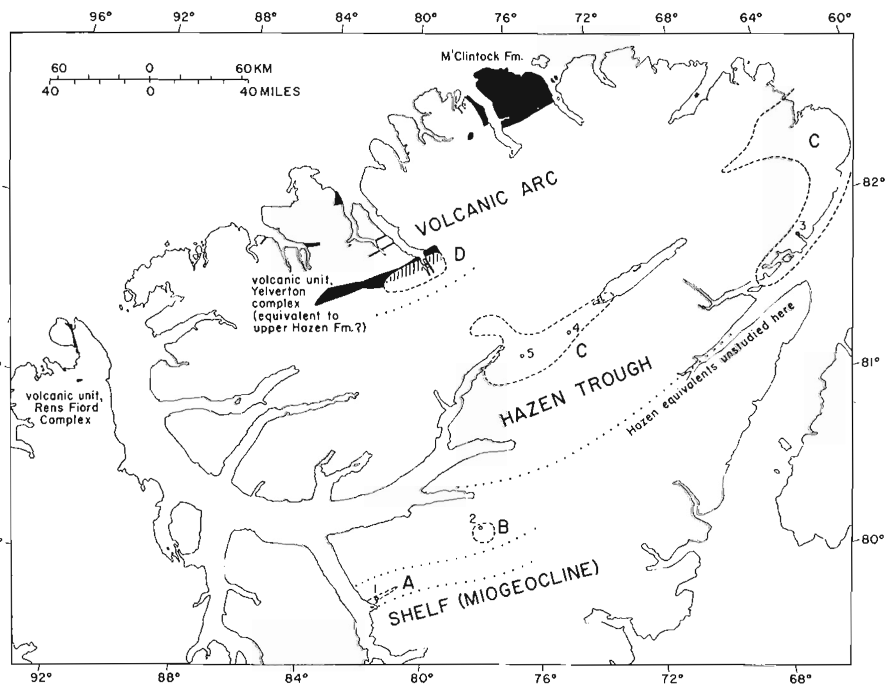


Figure 93. 4

Generalized outcrop area of facies in the Hazen Formation (A to D), outcrop areas of dated or presumed upper Middle to lower Upper Ordovician volcanic rocks (black) and localities referred to in text. Lined pattern indicates overlap of Hazen Formation and volcanic rocks.

exposure and extreme deformation. A section at locality 4, composed of several incomplete partial sections, is about 92 m thick.

Late Middle Ordovician graptolites occur near the top of the type section, and late Early Ordovician (latest Arenigian) graptolites 70 m or more above the base. Poorly preserved corals, probably not older than Middle Ordovician (B. S. Norford), were found in the middle of the carbonate member between localities 4 and 5 (Fig. 93. 4).

Facies C represents the axial region of the Hazen Trough, which probably received sediments from both sides. Terrigenous silt and sand, scarce in facies A and B, were derived from the northwest and the same probably applies to carbonate breccias that appear too coarse to have been derived from the remote southeastern margin. Lime mud of the miogeocline, on the other hand, could well have been transported into this area.

Facies D occurs adjacent to the head of Yelverton Inlet where it overlies the Grant Land Formation and underlies the volcanic unit of the Yelverton complex. All three formations are exposed in two synclines on the southwest side of the inlet (*see* Fig. 93. 11) but stratigraphic sections could not be measured owing

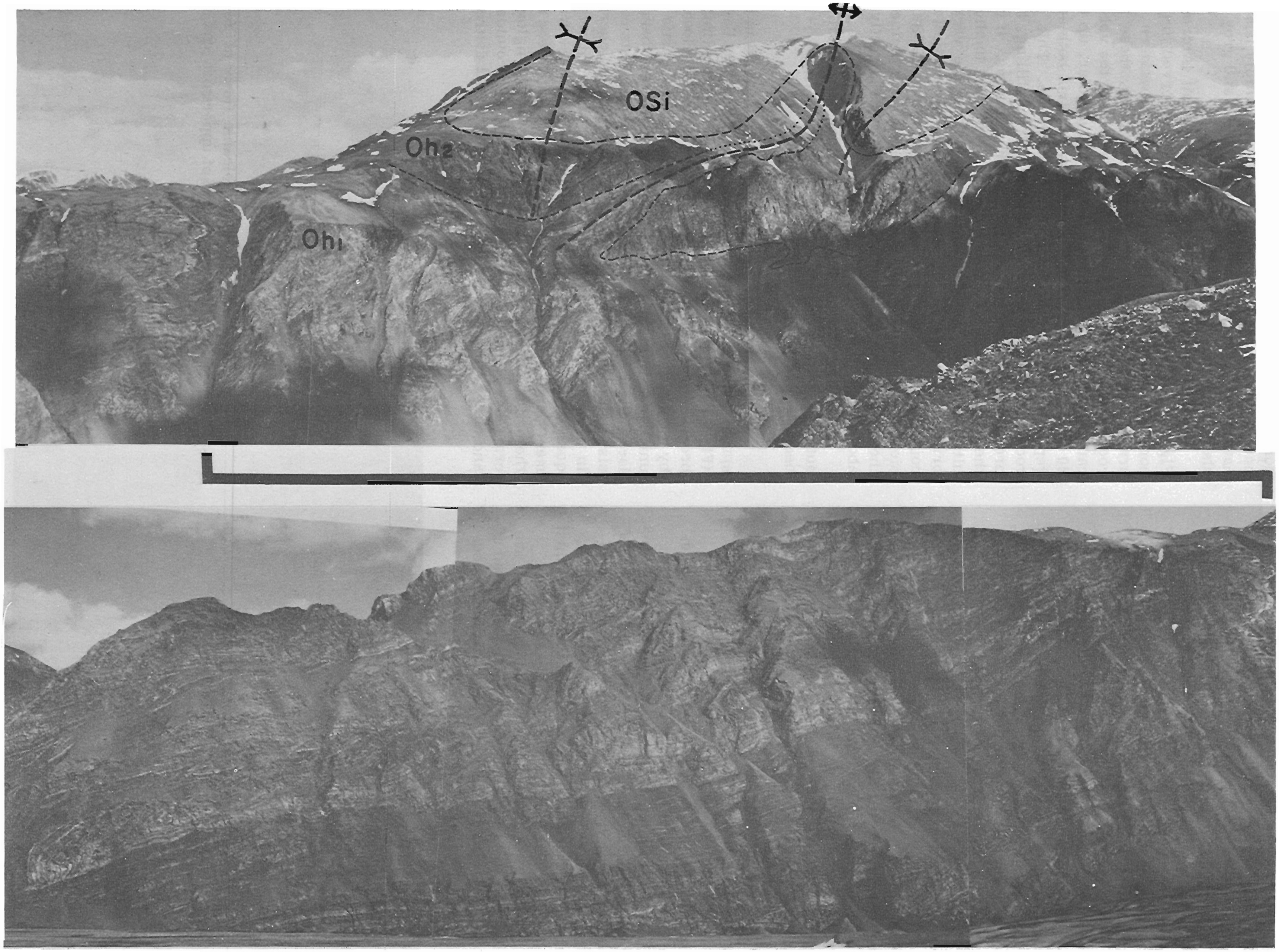


Figure 93. 5. Major folds, formed in competent chert member of Hazen Formation (OH_2) terminate near base of member; underlying incompetent carbonate member (OH_1) shows chaotic minor folds and faults. Imina Formation (OSi) is represented mainly by felsenmeer. South branch of Macdonald River, view to the east from locality 5, Figure 4.

to partial cover and structural complexity. The facies consists mainly of medium dark grey, variably argillaceous limestone with thinly interlaminated slate showing horizontal and small-scale cross-lamination. Thin-bedded chert is associated with the carbonate strata, but the conspicuous, compact chert member of the Hazen Plateau is not recognized here. It is tentatively assumed that volcanic rocks of the Yelverton complex occupy the stratigraphic position of the chert member in this region (*see* Fig. 93.1, correlation chart).

The Yelverton complex directly overlies the Grant Land Formation in a belt lying 30 to 72 km (19-45 miles) southwest of the head of Yelverton Inlet and, also, in an anticline east of the head of the inlet. It is uncertain whether the absence of the Hazen Formation is due to faulting or an unconformity.

Yelverton volcanic sedimentary complex

The provisional, informal name, Yelverton complex, is here used for a structurally very complex succession of volcanic, carbonate, and clastic rocks that occurs stratigraphically between the Hazen and Imina formations. It is divisible into three major units, a lower predominantly volcanic unit, and two overlying sedimentary units. The three units could not be separated in some parts of the area and satisfactory stratigraphic sections are rare.

Volcanic unit. This formation is at least 1000 m thick in a syncline east of the upper part of Yelverton Inlet. It consists mainly of pyroclastic rocks and breccias with lesser amounts of unbrecciated flows, some of which show pillows (Fig. 93.6). The rocks appear to be mainly intermediate in composition (between basalt and rhyolite) and are metamorphosed in the greenschist facies; all plagioclase identified so far is albite.

Relatively thin limestone units of limited extent are associated with the volcanics. Southwest of the head of Yelverton Inlet, for example, an intensely deformed oolitic grainstone, 10 to 20 m thick, occurs about 75 m above the base of the unit. At another locality, a carbonate member, 175 m thick, in the upper part of the unit consists mainly of medium dark grey, thinly laminated, aphanitic limestone with some green and grey slate.

A massive serpentinite breccia (map-unit 4Abx of Fig. 93.11), more than 10 m thick, was observed in the core of a complex anticlinorium south of the western arm of Kulutingwak Fiord. Stratigraphically, these rocks appear to occupy the uppermost part of the volcanic unit. The fragments are round to angular, up to 20 cm or more long, and fragments and matrix both are replaced by minerals of the serpentine group. This rock is tentatively interpreted as a diatreme filling, the brecciation having been caused by erupting gas. If so, the original rock may have been kimberlite but unaltered ultramafic rock was not discovered during the very brief examination of this locality. The serpentinite could, alternatively, represent a minor ultramafic intrusion (e.g. dunite) but this would leave unexplained the occurrence of an unusual conglomerate directly above it (*see* below).

Air observations suggest that other serpentinite outcrops are present in the area but they have not yet been mapped.

Carbonate-clastic unit. This unit is exposed in the anticlinorium mentioned, and west of Yelverton Bay (Fig. 93.11). In the anticlinorium, the serpentinite breccia (included in the volcanic unit) is overlain locally by a conglomerate (assigned to the carbonate-clastic unit), consisting of poorly sorted, rounded to subangular phenoclasts up to 30 cm long. The phenoclasts consist mainly of dolostone with some quartzite and chert in a sand-grade matrix of carbonate grains, quartz, quartzite, chert and minor andesitic volcanic rock fragments and muscovite. Areal extent and thickness of the bed are uncertain. It is thought that this conglomerate may represent country rock, fractured, rounded, and mixed by gas erupting from the postulated diatreme.

Figure 93.6

Pillows formed in volcanic flows of Yelverton complex west of uppermost part of Yelverton Inlet.

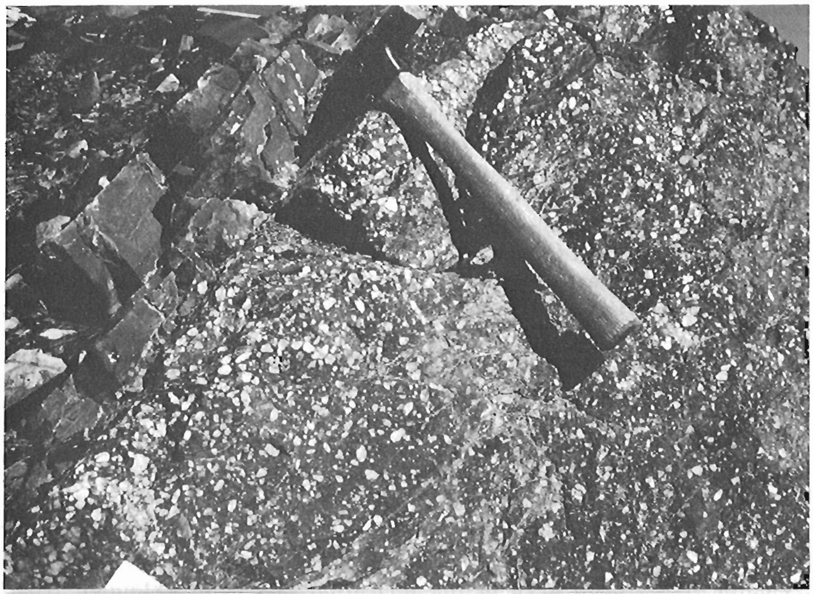


Figure 93.7

Pebble conglomerate with abundant sandy and muddy matrix intercalated between sandstones and siltstones that form Bouma sequences. Matrix content increases and pebble grade decreases slightly stratigraphically upward in a section that is upside down. Lands Lokk Formation, Yelverton Inlet region, loc. 75TM152 of Figure 93.11.

Strikingly similar in field appearance is a conglomerate occurring in the lower part of the carbonate-clastic unit west of Yelverton Bay. The phenoclasts also are dolomitic, but are smaller (a few centimetres in length) and dispersed in an abundant, poorly sorted groundmass of sand- to predominantly silt-grade carbonate and quartz grains with some muscovite and feldspar. This bed occurs throughout an area about 15 km in diameter and is clearly not confined to a diatreme. However, the phenoclasts may have been blown out of the diatreme(?) discussed or out of other vents in the vicinity.

Carbonate unit. This unit consists of probably not more than 100 m of original limestone, dolomitic limestone, and argillaceous limestone, in part metamorphosed to marble or calcareous phyllite. It underlies the Imina Formation in several anticlines (Fig. 93.11).

Amphibolite-grade metasedimentary and metavolcanic rocks (map-unit 4m). Amphibolite and quartz-biotite schist form a sinuous belt along the southeastern, apparently faulted boundary of the Proterozoic crystalline complex that is aligned with volcanic and carbonate rocks at Phillips Inlet (map-units Sv and Sc of Trettin, 1971) but separated from them by glaciers (*see* Frisch, preceding report 92). Some specimens examined in thin section are comparable to sediments of the carbonate-clastic unit; others, plagioclase-rich rocks, could be altered tuff. A marble lying adjacent to the faulted contact with the Proterozoic rocks (map-unit 4ml) locally contains what appear to be metamorphosed remnants of echinoderm columnals. These metamorphic rocks are correlated tentatively with the Yelverton complex (*see* legend, Fig. 93.11).

Age, correlation, and regional significance of Yelverton complex. The volcanic unit is comparable in lithology, order of thickness, and stratigraphic position to the fossiliferous upper Middle to lower Upper Ordovician M'Clintock Formation of northern Ellesmere Island (Fig. 93.1 and Trettin, 1969b) and this name will be used should the proposed correlation be supported by conodonts. It also is comparable to the volcanic unit of the Rens Fiord Complex in northern Axel Heiberg Island (Trettin, 1969a). All three formations consist of pyroclastic deposits, flows and intercalated carbonates. They probably represent a volcanic arc, extending from northern Ellesmere Island to northern Axel Heiberg Island (or farther) that fringed the Hazen Trough on the northwest during late Middle Ordovician to early Late Ordovician time. The carbonates formed in limited shelf areas on volcanic islands or seamounts, in part by inorganic precipitation in shallow, agitated water (ooids).

The combined sedimentary units of the Yelverton complex are comparable in stratigraphic position and lithology to the combined Upper Ordovician Ayles, Taconite River, and Zebra Cliffs formations of M'Clintock Inlet, but a one-to-one correlation of individual units is not applicable owing to facies changes. Both packages of strata lie stratigraphically between volcanics below and flysch of the Imina Formation above, and both are composed of mixed carbonate and clastic sediments. The sedimentary units of the Yelverton Inlet area, however, are thinner, finer in grade, and lack the boulder conglomerate and red beds present at M'Clintock Inlet. The carbonate unit of the Rens Fiord Complex probably also belongs to this Upper Ordovician succession.

Following the late Middle to early Late Ordovician event, a continuous shelf area appears to have developed in northern Ellesmere and Axel Heiberg islands while the Hazen Trough persisted farther to the southeast. This shelf received carbonate sediments as well as clastic detritus derived from a source in the present offshore region (Pearya Geanticline; Trettin, 1971). A major dispersal centre seems to have lain adjacent to the M'Clintock Inlet region where the coarsest clastic sediments occur.

Imina Formation

The Imina Formation is a very thick succession of alternating siltstone and sandstone with local conglomerate and breccia. The sandstone and siltstone are calcareous and dolomitic and show flysch-like structures such as: graded and massive bedding; horizontal, convolute, and small-scale cross-lamination; and sole markings. The terrigenous sediment consists mainly of quartz with lesser amounts of feldspar, chert, muscovite, chlorite and low-grade metamorphic rock fragments. The unit is widely distributed in northern and central Ellesmere Island, having been mapped in northwestern Ellesmere Island, at M'Clintock Inlet, on the Hazen Plateau, east of Cañon Fiord, and west of Trold Fiord. Diagnostic fossil collections demonstrate changes in age that reflect the initial expansion and subsequent southeastward shift of the Hazen Trough. In northern Ellesmere Island, only partial sections are exposed, but complete and relatively undisturbed sections are exposed east of Cañon Fiord that have been studied in some detail (Trettin, in prep.). More than two thousand paleocurrent determinations from the Hazen Plateau demonstrate that the Silurian sediments were transported by turbid flows that entered the trough transversely, from the northwest, and were deflected southwestward, parallel with strike, in the axial region. These sediments undoubtedly were derived from a northerly source area in the present offshore region (Pearya Geanticline). Upper Lower Devonian flysch sediments at Trold Fiord, on the other hand, are now known on the basis of paleocurrent determinations (Trettin, 1974) to have been derived from southeasterly to easterly sources. Biostratigraphy demonstrates correlation of the Trold Fiord outcrops with the partly nonmarine Vendom Fiord Formation of central Ellesmere Island (unpubl. conodont identifications by T.T. Uyeno and brachiopod identifications by R.E. Smith, J.G. Johnson, and D. Perry).

During the present investigation, the known outcrop area of the formation has been extended from Phillips Inlet eastward into the Yelverton Inlet region, and from the Hazen Plateau southwestward toward the Agassiz Ice Cap. At Yelverton Inlet, the Imina Formation occurs stratigraphically between carbonates of the Yelverton complex and siltstone and slate of the Lands Lokk Formation and is tentatively considered as latest Ordovician and Early Silurian in age. Although metamorphosed in the greenschist facies, primary structures are generally preserved, but the outcrops seen were not suitable for paleocurrent studies. Southwest of the Hazen Plateau, the Imina overlies the Hazen Formation and is probably Early Silurian and(?) younger. More than one hundred paleocurrent determinations at a locality 14 km northeast of Lake Tuborg (Fig. 91.3) indicate southward (transverse) and southwestward (longitudinal) flow as in other parts of the Hazen Plateau.

Lands Lokk Formation

The name Lands Lokk Formation was given (Trettin, 1969a) to a Silurian succession of predominantly clastic sediments with lesser amounts of volcanic rocks and very small amounts of limestone that conformably overlies the Imina Formation in northwestern Ellesmere Island and is unconformably overlain by the Lower Devonian Stallworthy Formation in northern Axel Heiberg Island. Three members are distinguished in the type area south of Emma Fiord, northwestern Ellesmere Island (Figs. 93.1, 93.8): member A, composed mainly of siltstone and slate; member B, composed of volcanic and clastic rocks with some limestone lenses; and member C, composed mainly of siltstone and sandstone with lesser amounts of pebble conglomerate and slate. The volcanic rocks occur only in two areas; elsewhere member C lies directly on member A. Diagnostic graptolites indicated that, in the type area, member A is Wenlockian to early Ludlovian, and that members B and C are early Ludlovian in age.

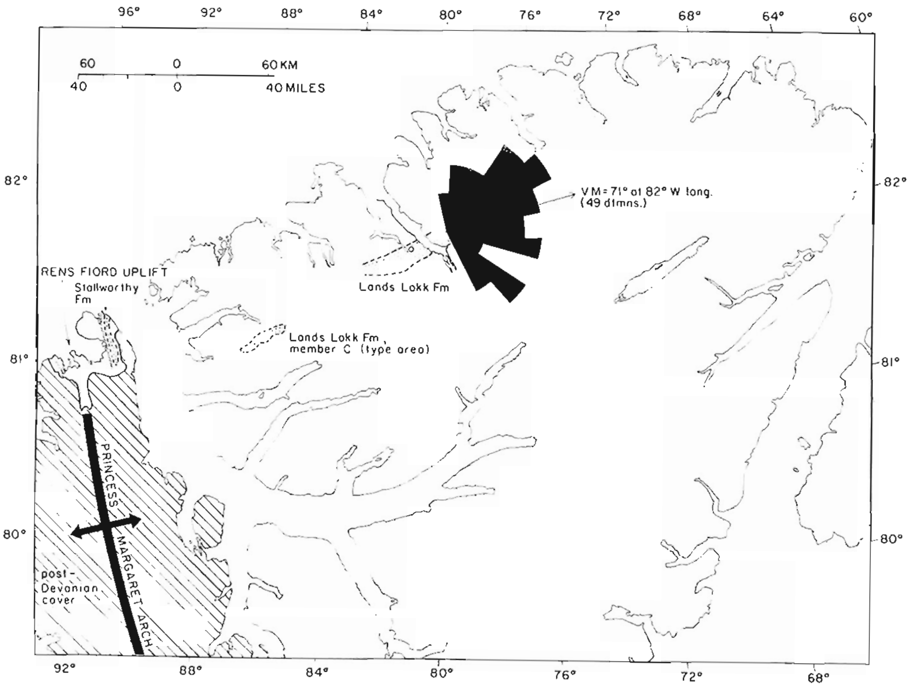


Figure 93.8

Paleocurrent data from the Lands Lokk Formation; outcrop areas of Lands Lokk and Stallworthy Formations; and their inferred source, the Rens Fiord Uplift. This lower and middle Paleozoic tectonic element probably extends some distance under the Tertiary Princess Margaret Arch formed in the upper Paleozoic and Mesozoic cover.

An upward-coarsening succession of slate, siltstone, and sandstone with minor amounts of pebble conglomerate overlying the Imina Formation in the Yelverton Inlet region is assigned to the Lands Lokk Formation (members A and C) on the basis of stratigraphic position, lithology, and mineral composition. Repetitive graded bedding, massive bedding, sole markings, and other features characteristic of turbidites and mass flow deposits are well developed in the sandstones and siltstones. A typical pebble conglomerate, intercalated between Bouma sequences of sandstone and siltstone and showing an upward increase in matrix content (Fig. 93.7) also is of deep-water origin. These features, as well as the graptolitic fauna of the type area, indicate that the Lands Lokk Formation represents a prograding submarine fan complex rather than a prograding delta as originally (Trettin, 1969a) suggested.

The Lands Lokk differs from the Imina Formation in containing a markedly greater content of dark grey slate and siltstone (especially in the lower part) and in the composition of sandstone and siltstone. The Lands Lokk sandstone consists mainly of quartz and chert with less carbonate particles, feldspar, muscovite, chlorite, and rock fragments than the Imina sandstone.

Two lines of evidence suggest that the Lands Lokk Formation was not derived from the Pearya Geanticline but from the Rens Fiord Uplift (Fig. 93.8) of northern Axel Heiberg Island (including its extension beneath the Tertiary Princess Margaret Arch). (1) The sandstone and conglomerates of member C are similar in composition to those of the Lower Devonian Stallworthy Formation of northernmost Axel Heiberg Island; the latter undoubtedly was derived from Grant Land quartzite and Hazen chert exposed in the core of the adjacent Rens Fiord Uplift. (2) Paleocurrent determinations at locality 75TM192 (Figs. 93.8, 93.11) indicate northeastward, rather than southward flow.

Map-unit 7

This unit has been observed only in an area southwest of the head of Yelverton Inlet where it is overlain by chertified limestone of the Marvin Formation; the lower contact is not exposed (Fig. 93.11). It consists of a variety of medium to dark grey, quartz-silty and sandy, variably dolomitic limestones and related dolostones with interbedded calcareous and argillaceous, fissile siltstones. Laterally and upward in the section, the carbonate sediments pass into replacement chert that has retained colour and gross stratification of the original rocks. The total thickness of the unit is unknown owing to excessive structural deformation but probably exceeds 100 m.

The most common limestone is a silt- to very fine sand-grade packstone that shows thin horizontal lamination or small-scale cross-lamination and contains considerable amounts of silt- to sand-grade detrital quartz and scattered microcrystalline euhedral dolomite. Some graded beds show Bouma sequences. A typical 6.5-cm-thick specimen, for example, which decreases in grain size from very coarse sand-grade at the base to silt grade at the top, shows (miniature) massive bedding, small-scale cross-lamination, and horizontal lamination in vertical succession. These beds are interpreted as distal carbonate turbidites.

Marvin Formation

The Marvin Formation was established (Trettin, 1969b) for about 183 m of shelf-type limestone and sandy limestone at M'Clintock Inlet that are intercalated between

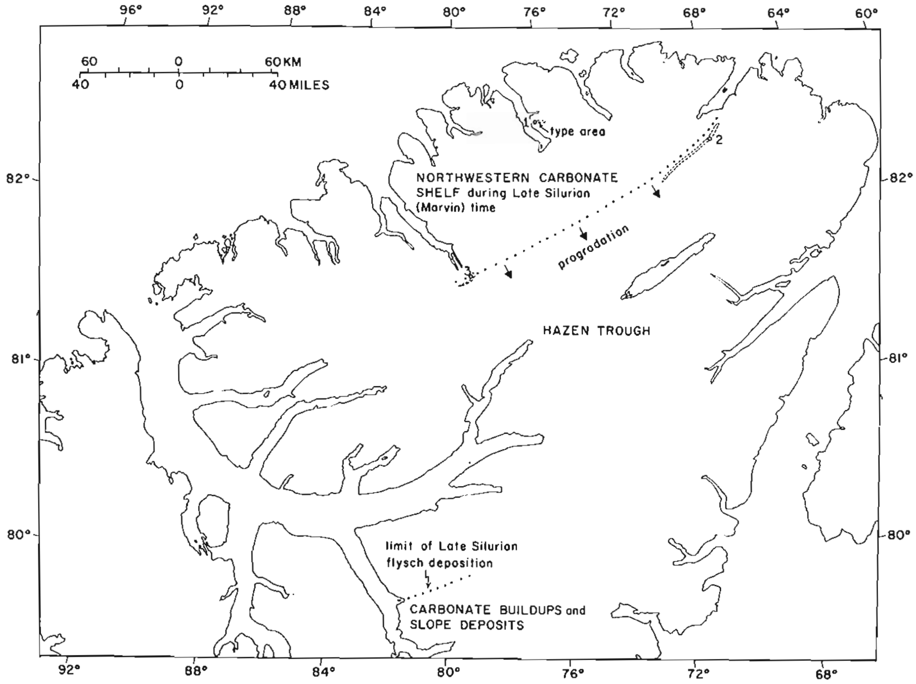


Figure 93.9

Outcrop belts of Marvin Formation, localities discussed, and inferred paleogeography during part of Late Silurian time.

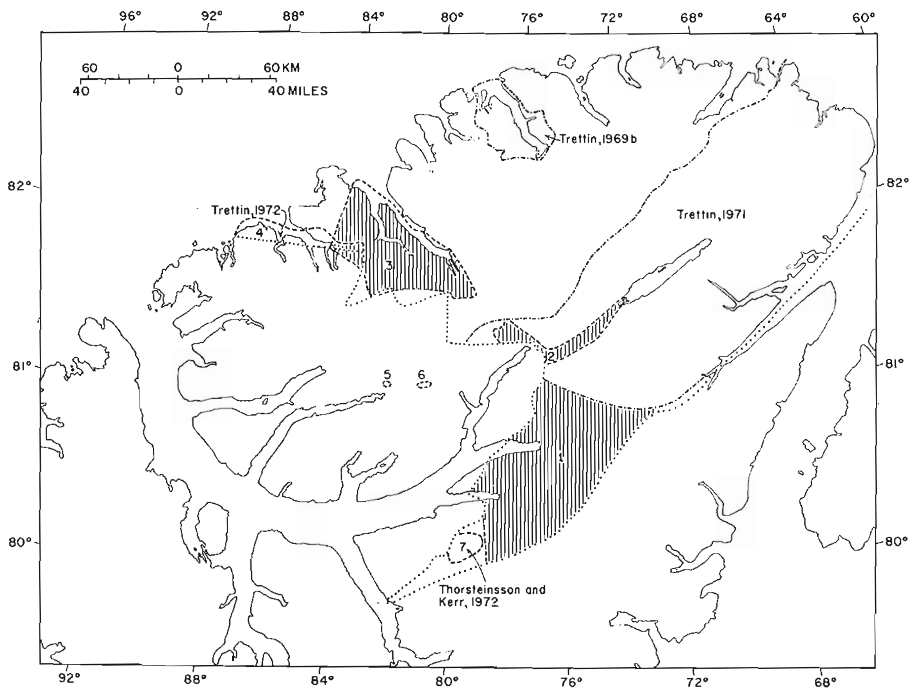


Figure 93.10

Index map showing areas mapped or revised in 1975.

unnamed clastic formations. In the type area, the formation contains a Late Silurian fauna characterized by a very large species of *Kirkidium*, ?*Atrypella*, and various corals. Tabular stromatoporoids form patch reefs at some stratigraphic levels.

Another outcrop belt, extending from northeast of Piper Pass to Barrier Glacier, was examined briefly by the writer at Piper Pass in 1967 (Trettin, 1971; Fig. 93.9, loc. 2 of this report). There, it consists of medium dark grey, slaty limestone and slate, dolomitic limestone, dolostone, and skeletal breccia. This belt requires more detailed study, but the dark grey limestone and slate as well as the breccia are suggestive of a slope rather than shelf environment. Fossils from this locality and from Barrier Glacier (Christie, 1964) are comparable to those from M'Clintock Inlet.

An outcrop belt of comparable limestone, discovered by Nassichuk (Nassichuk and Christie, 1969) southeast of the head of Yelverton Inlet (Fig. 93.9, loc. 3) was re-examined briefly by the writer in 1975. At the base of the outcrop (the underlying strata are concealed by overburden) there is a breccia composed of lithic and skeletal fragments, mainly stromatoporoids with lesser corals and brachiopods. The breccia is overlain by 50 m or more of medium grey, massive limestone with relatively scarce fossils, including favositid corals. The fossils from this locality are regarded as Middle or Late Silurian by B. S. Norford. The same unit seems to continue southwest of a major glacier flowing into Yelverton Inlet, where it is totally replaced by chert and underlain by chertified beds of map-unit 7.

The skeletal breccia probably is a submarine talus slope deposit derived from a stromatoporoidal and coralline reef complex, and the overlying limestone probably represents a somewhat shallower shelf margin environment. Map-unit 7 and the Marvin Formation combined are a regressive succession that indicates southeastward progradation of the northern Ellesmere carbonate shelf into the Hazen Trough.

Geological mapping

This section contains brief comments on three areas mapped or remapped completely in 1975 and a few other previously mapped areas for which some revisions are suggested (*see* Fig. 93.10).

Area 1, previously unmapped, is underlain mainly by tightly folded strata of the Imina Formation, but the Hazen Formation outcrops in an anticlinorium southwest of d'Iberville Glacier (*see* Hazen Formation, facies B, Fig. 93.4).

Area 2, although included in a previous reconnaissance map (Trettin, 1971, Fig. 3), was poorly known and required complete remapping. It is underlain by Grant Land, Hazen, and Imina formations. The contact between Grant Land and Hazen formations, lies in incompetent, complexly folded and faulted strata and is difficult to recognize on aerial photographs.

The geology of the previously unmapped area 3 is shown in Figure 93.11.

A reconnaissance map of area 4 (Trettin, 1971, Fig. 2) can be adapted for present compilations by reassigning map units Sv and Sc to the volcanic and carbonate units of the Yelverton complex respectively, and map-unit 1Ph, as well as map-unit Dqm southwest of the entrance of Phillips Inlet, to the Proterozoic crystalline complex (*see* Frisch, Fig. 92.1).

Some corrections have to be made in the lower Paleozoic geology of areas 5 and 6 (included in Thorsteinsson and Trettin, 1972). In area 5, a fault slice of Hazen chert occurs above the Grant Land Formation and other such slices probably are present in the general area. The limestone, shown as map-unit Sc in area 6, now is known to be a fault block of Nansen Formation on the basis of fossil identifications by B. S. Norford.

Area 7, mapped by J. Wm. Kerr in 1962 (Thorsteinsson and Kerr, 1972), was not reached in 1975 owing to weather conditions. The geology of this area probably can be adapted to the present stratigraphy by re-assigning the Cape Phillips Formation to the chert member, and the Cornwallis Group to the carbonate member of the Hazen Formation.

Some structural observations

Vertical changes in structural style, between Lake Hazen and head of Tanquary Fiord

Two major features characterize the structure of the remapped belt between Lake Hazen and the head of Tanquary Fiord: the Lake Hazen fault zone, which has thrust the Grant Land Formation southeastward upon rocks ranging in age from Ordovician to Mesozoic; and a fold belt, bordering the fault zone on the southeast. The fold axes in this belt, which involves Grant Land, Hazen, and Imina formations, trend regularly to the southwest, and most axial planes are inclined steeply to the northwest. Competent and incompetent units within this succession differ markedly in structural style. Doubly-plunging anticlines - presumably flexural slip folds - are formed in competent quartzite and minor interbedded siltstones just northwest of Very River that can be traced for distances of up to about 2.7 km. The slaty upper part of the Grant Land Formation and thinly stratified carbonate member of the Hazen Formation, on the other hand, are characterized by chaotic minor folds and faults (Fig. 93.5) that are mostly unmappable. Only the thicker bedded basal sub-unit of the Hazen Formation forms flexural slip folds with half-wave lengths in the order of tens of metres that can be recognized locally (Fig. 93.2). Major doubly-plunging anticlines formed by slippage of chert beds along thinly interstratified argillaceous laminae characterize the chert member of the Hazen Formation (Fig. 93.5). These folds are traceable for distances up to 12 km. The immediately overlying beds of the Imina Formation conform with this pattern but, higher in the section, very complex minor folds take over. The folding in the Imina Formation is not "similar" as previously stated (Trettin, 1971), but approaches the chevron-type characteristic of many flysch formations (Ramsay, 1974; cf. Figs. 44 and 45 *in* Trettin, 1971).

Major structural features of Yelverton Inlet region

The structure of the Yelverton Inlet region is exceedingly complex and poorly understood; the brief time available was used to establish a stratigraphic framework and map major units rather than for detailed structural studies. Only some of the most important features are pointed out here.

Structural trends are mainly southwesterly but swing into northerly directions west of Yelverton Bay. There, imbricate faults have thrust the Grant Land Formation upon the Yelverton complex in eastward directions.

Proterozoic gneisses at Yelverton Bay dip mainly southeast at low to moderate angles and must be involved in folds that are overturned toward the northwest or in thrust faults with movement in that direction (Fig. 93.12). Major thrust faults, formed in upper Paleozoic and Triassic strata at the head of Yelverton Inlet, also are

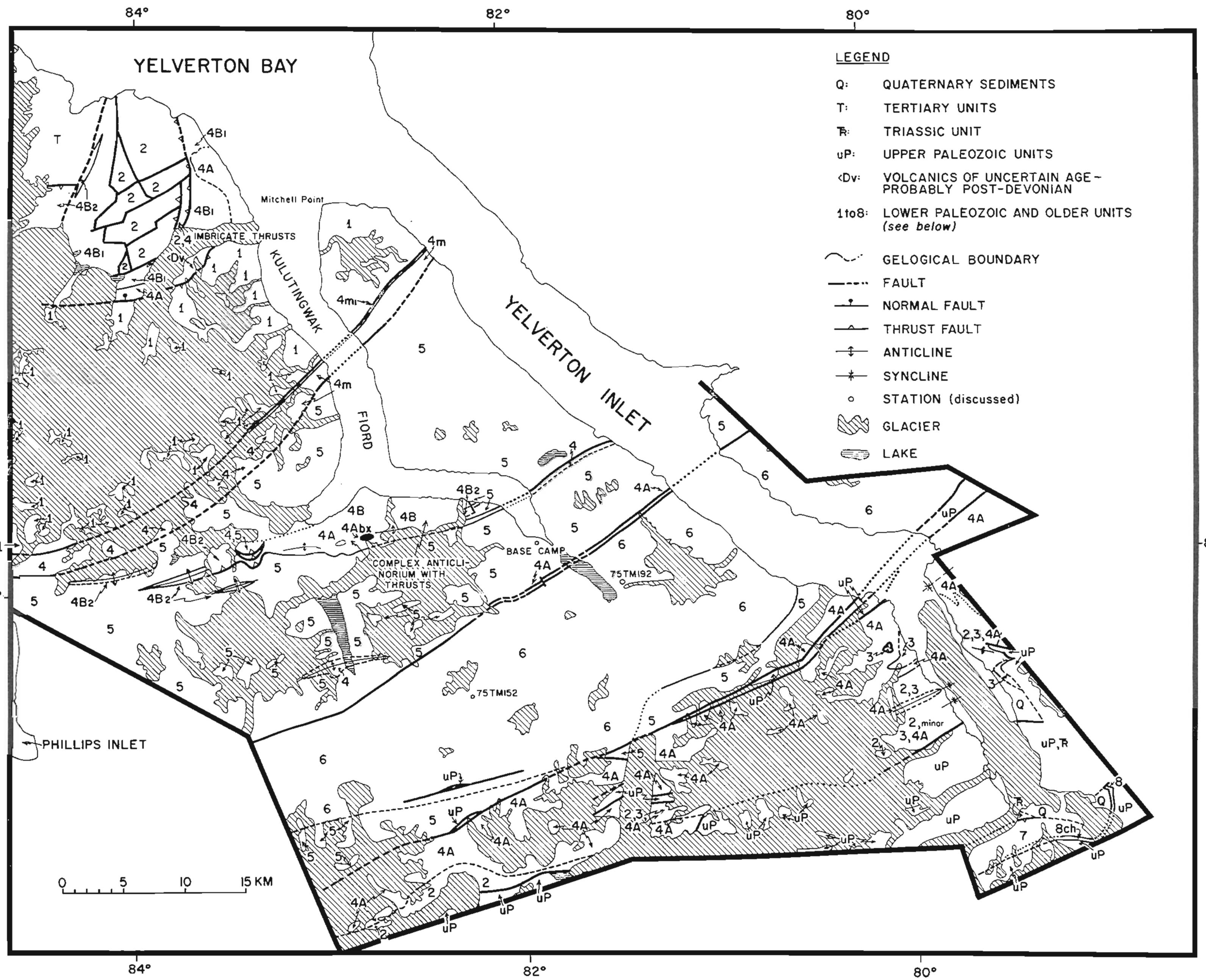


Figure 93.11. Provisional geological map, Yelverton Inlet region.

LEGEND FOR LOWER PALEOZOIC AND OLDER ROCKS

(Most age assignments are based on lithological correlation and provisional)

SILURIAN

Upper Silurian

8 Marvin Formation: massive limestone: skeletal and lithic limestone breccia; 8 ch: replaced by chert

7 limestone, variably silty, sandy, and dolomitic: calcareous and dolomitic siltstone; minor dolostone: all rock types partly replaced by chert

Upper and Middle Silurian

6 Lands Lokk Formation: siltstone, sandstone, slate: minor pebble conglomerate

SILURIAN AND (?) ORDOVICIAN

Lower Silurian and (?) Upper Ordovician

5 Imina Formation: calcareous and dolomitic siltstone and sandstone (partly metamorphosed in greenschist facies)

ORDOVICIAN

4 Yelverton complex (undivided)

4B sedimentary units (undivided)

4B2 carbonate unit: limestone, in part dolomitic and argillaceous, variably metamorphosed

4B1 carbonate-clastic unit: sandstone and siltstone, variably calcareous and argillaceous, argillaceous quartz sandstone, limestone, conglomerate (mostly metamorphosed in greenschist facies)

4A volcanic unit: pyroclastic rocks and volcanic flows, mostly of intermediate composition, metamorphosed in greenschist facies; minor limestone, dolomitic limestone, slate

4Abx serpentinite breccia (probably diatreme-filling in upper part of 4A)

3 Hazen Formation: limestone, chert, slate

ORDOVICIAN AND (?) CAMBRIAN

Lower Ordovician and (?) Upper Cambrian

2 Grant Land Formation: quartzite, variably feldspathic; red, green and grey slate (red slate absent west of Yelverton Bay); minor conglomerate

PROTEROZOIC

Lower Proterozoic or Older

1 amphibolite-grade granitic gneiss and metamorphosed sediments and basic intrusions

4m amphibolite-grade metamorphic rocks, probably equivalent to Yelverton complex (especially unit 4B1)

4m1 marble (locally crinoidal)

Figure 93.12

Southeast-dipping Proterozoic gneisses and metasediments at Mitchell Point, Yelverton Bay, view to the southeast.

directed to the northwest. By contrast, the folds in the lower Paleozoic rocks are overturned mainly to the southeast. Most axial planes are steeply inclined but nearly recumbent folds occur on the ridge east of the upper part of Yelverton Inlet. Nearly recumbent folds also are indicated by upside-down strata of the Lands Lokk Formation (Fig. 93.7) at locality 75TM152 of Figure 93.11.

The folds in the extensive clastic terrains are generally too small and complex to be mapped but several major, doubly-plunging anticlines, involving the upper part of the Yelverton complex, have been outlined; they seem to form a left-hand en echelon pattern.

References

Christie, R.L.
1964: Geological reconnaissance of northeastern Ellesmere Island, District of Franklin; Geol. Surv. Can., Mem. 331.

Frisch, T.O.
1974: Metamorphic and plutonic rocks of northernmost Ellesmere Island, Canadian Arctic Archipelago; Geol. Surv. Can., Bull. 229.

McKee, E.D., Crosby, E.J., and Berryhill, H.L., Jr.
1967: Flood deposits, Bijou Creek, Colorado, June 1965; J. Sediment. Petrol., v. 37, p. 829-851.

Nassichuk, W.W. and Christie, R.L.
1969: Upper Paleozoic and Mesozoic stratigraphy in the Yelverton Pass region, Ellesmere Island, District of Franklin; Geol. Surv. Can., Paper 63-39.

Ramsay, J.G.
1974: Development of Chevron folds; Geol. Soc. Am. Bull., v. 85, p. 1741-1754.

Sinha, A.K. and Frisch, T.O.
1975: Whole-rock Rb/Sr ages of metamorphic rocks from northern Ellesmere Island, Canadian

Sinha, A.K. and Frisch, T.O. (cont'd.)
Arctic Archipelago. I. The gneiss terrain between Ayles Fiord and Yelverton Inlet; Can. J. Earth Sci., v. 12, p. 90-94.

Thorsteinsson, R. and Kerr, J. Wm.
1972: Geology, Greely Fiord east, District of Franklin; Geol. Surv. Can., Map 1348A.

Thorsteinsson, R. and Trettin, H.P.
1972: Geology, Otto Fiord, District of Franklin; Geol. Surv. Can., Map 1309A.

Trettin, H.P.
1969a: Pre-Mississippian geology of northern Axel Heiberg and northwestern Ellesmere islands, Arctic Archipelago; Geol. Surv. Can., Bull. 171.

1969b: Geology of Ordovician to Pennsylvanian rocks, M'Clintock Inlet, north coast of Ellesmere Island, Canadian Arctic Archipelago; Geol. Surv. Can., Bull. 183.

1971: Geology of lower Paleozoic formations, Hazen Plateau and southern Grant Land Mountains, Ellesmere Island, Arctic Archipelago; Geol. Surv. Can., Bull. 203.

1972: Reconnaissance of lower Paleozoic geology, Phillips Inlet region, north coast of Ellesmere Island, District of Franklin; Geol. Surv. Can., Paper 71-12.

1974: Stratigraphy and sedimentology of Silurian and Devonian clastic formations, central Ellesmere Island; *in* Report of Activities, Part A, Geol. Surv. Can., Paper 74-1A, p. 357-360.

Middle Ordovician to Lower Devonian deep-water succession east of Cañon Fiord, Ellesmere Island: sedimentation and tectonic development at southeastern margin of Hazen Trough; Geol. Surv. Can., Bull. (in prep.)

94.

UPPER PALEOZOIC SUCCESSION IN THE YELVERTON AREA, NORTHERN ELLESMERE ISLAND, DISTRICT OF FRANKLIN

Project 730051

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This report deals with the upper Paleozoic succession in the area between the head of Tanquary Fiord and the northern part of Yelverton Inlet, northern Ellesmere Island (*see* index map in Fig. 94.1). The geology of parts of the area has been described by Christie (1965) and by Nassichuk and Christie (1969). The sequence discussed is on the Tanquary structural high, along its western or northwestern flank, and to the northwest and north of the high. This report is based on field descriptions of lithology and on field observations of the physical relationships of the various rock units. Microscopic study of lithologies and identification of fossils are in progress. Figure 94.1 shows a generalized stratigraphic cross-section compiled from selected localities.

Borup Fiord Formation

The Borup Fiord Formation, which consists of dark red sandstone, siltstone, and shale, is the lowest upper Paleozoic unit in the area. It lies unconformably on the strongly folded lower Paleozoic rocks. The thickness of the formation is variable; the measured maximum thickness is about 900 m (3000 ft.) at location 2. At other localities it is absent.

At location 5 and in the lower part of the formation at location 2 the sandstone, siltstone, and shale occur as parts of upward-fining, generally less than 10-m-(33 ft.) thick cycles. Three subdivisions can be distinguished in each cycle. The lower one is composed of resistant, thick bedded to massive sandstone, which is coarse grained and conglomeratic in the basal part, and has a sharp, erosional lower contact. This unit commonly contains planar crossbeds of 30 cm (1 ft.) or more thickness per set. Trough crossbedding is rare. The lower subdivision grades upward into the middle subdivision of medium- and fine-grained sandstone and siltstone. The sandstone and siltstone are thin to very thin bedded and contain small-scale crossbeds. The upper part of a cycle consists of shale and silty shale which are, in places, interbedded with fine grained sandstone and siltstone. Mudcracks occur in this part of the cycle, but are rare.

The upper part of the formation at location 2 contains considerably less shale and siltstone than is present elsewhere and cyclicity is not apparent. Plant fragments and one bone fragment are the only megafossils recovered from the Borup Fiord Formation. The formation is probably a flood plain deposit of a meandering stream.

Unit A

In the Yelverton Pass area and in the nunataks about 25 km (15 miles) to the north (locality 10), the Borup Fiord Formation is overlain by resistant, white sandstone and sandy conglomerate. Although some thin intervals of red shale are present in the lower part of unit A, the basal contact of the unit is sharp. At Yelverton Pass unit A is 310 m (1017 ft.) thick, and is separated from the overlying units by an angular unconformity.

The sandstone consists of angular quartz grains and relatively large amounts of white chert fragments. Grain size becomes finer upward, in rhythmical intervals which are between 3 m and 5 m (10 ft. and 16.5 ft.) thick. The basal contacts of individual intervals are sharp and, in places, formed by channels. Massive or thick bedded sandy conglomerate and coarse grained conglomeratic sandstone form the lower part of the intervals. The upper part consists of medium grained, thin- to medium-bedded sandstone with conglomeratic lenses. Large scale crossbeds, both of planar and trough type, are present throughout the intervals, particularly in the upper part.

No megafossils were found in unit A, which may have been deposited in a braided stream environment.

Unit B

In the area at the head of Yelverton Inlet, the Borup Fiord Formation is overlain by unit B with a gradational contact; laterally the Borup Fiord interfingers with unit B. Unit B consists almost entirely of black or very dark grey, medium and coarsely crystalline dolomite. Bedding may be either massive or uniformly thin. Rare micritic limestone beds are black and thin bedded and contain abundant brachiopods. The limestone beds are slightly nodular with partings of black, calcareous shale. At location 1 the thickness of unit B is estimated to be 450 m (1500 ft.).

At locality 2 the basal transition zone is 179 m (587 ft.) thick and the carbonates in the zone consist of winnowed skeletal limestone and light coloured, medium and coarsely crystalline dolomite. Shallow marine, nearshore conditions are a likely environment for the transition zone, while euxinic conditions prevailed for the black limestone and dolomite in the higher part of unit B.

Unit C

Unit C was observed only in the Yelverton Inlet area. At locality 1 it is estimated to be 500 m (1600 ft.) thick. It thins toward the northwest and is not present to the southeast. Because of rugged terrain and poor weather conditions the section at locality 1 was not examined in detail. The lower part of the unit (C_1) consists of gypsum and anhydrite interbedded with

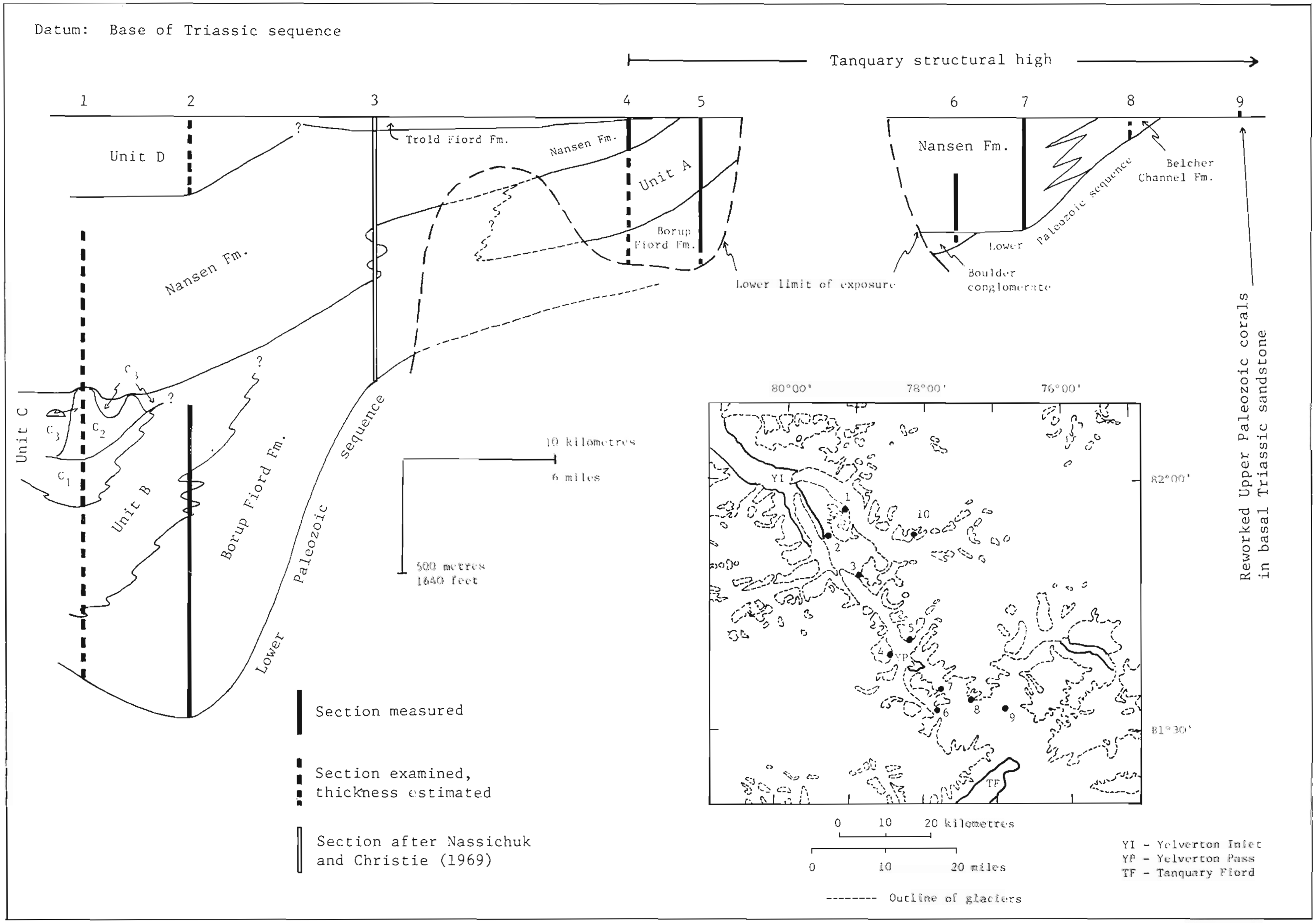

Figure 94. 1. Generalized stratigraphic cross-section, Yelverton Inlet to Tanquary Fiord, northern Ellesmere Island.

Figure 94. 2.

Nansen Formation overlying unit C at locality 1.

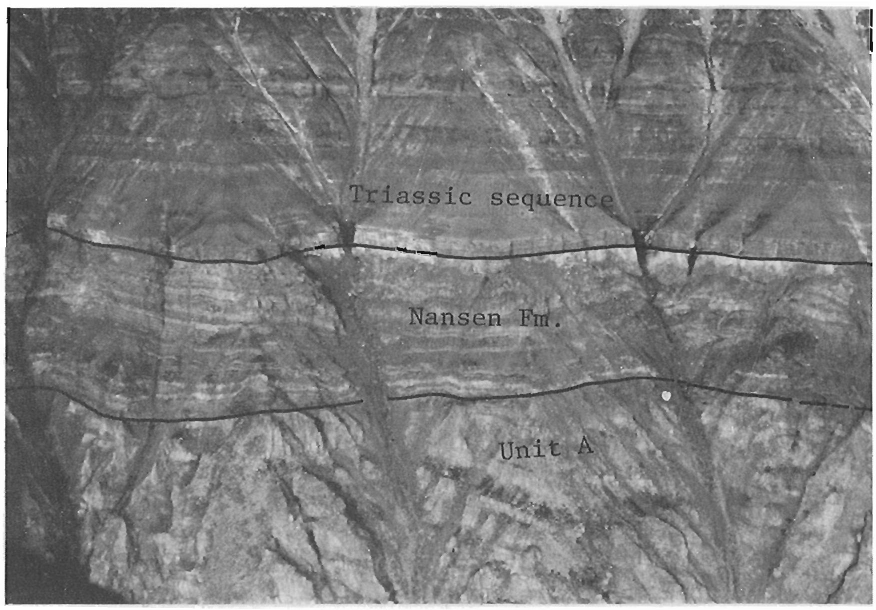

Figure 94. 3.

Pinch-out of Nansen Formation in vicinity of locality 4. The Nansen Formation here is approximately 40 m (130 ft.). Upper and lower boundaries are angular unconformities.

very dark grey lime mudstone. Sub-unit C_2 is formed by carbonate mounds of an estimated maximum thickness of 300 m (1000 ft.). At the base of the mounds very fossiliferous, massive dolomite occurs. Sub-unit C_3 comprises dark grey shale and thin-bedded dark grey siltstone.

Unit B and sub-unit C_1 are probably both correlative with the Otto Fiord Formation. Evaporites in a stratigraphic position similar to that of sub-unit C_1 also were reported by Trettin (1969) from the M'Clintock Inlet area about 95 km (59 miles) to the northeast. Lithology and stratigraphic relationships of sub-units C_2 and C_3 are similar to those in the lower and middle part of the Hare Fiord Formation, as discussed by Thorsteinsson (1974, p. 29, and 31).

Nansen Formation

The Nansen Formation varies in form depending on its position relative to the Tanquary structural high. It occurs as out-liers on the Tanquary structural high, and as pinch-outs at the flank of the Tanquary structural high in the Yelverton Pass area, but is fully developed in the Yelverton Inlet area.

At the locations on the Tanquary structural high (6 and 7 in Fig. 94. 1) the formation unconformably overlies the folded lower Paleozoic sequence. The thickness at locality 7 is close to 400 m (1300 ft.) and the formation consists almost entirely of limestone. Locally it is underlain by red boulder conglomerate. The limestone is generally thick bedded or massive and varies from light grey lime mudstone to well-winnowed skeletal grainstone. Several zones of sandstone and siltstone and of red and green, nodular and thin bedded limestone occur in the sequence. The red and green limestone is similar to that of the Belcher Channel Formation exposed farther southeast at the head of Tanquary Fiord and at locality 8. Brachiopods and corals are present throughout the formation.

In the Yelverton Pass area the Nansen Formation rests with angular unconformity on unit A and is overlain disconformably by the Trold Fiord Formation (locality 4). The greatest angle of unconformity was seen in the nunataks 25 km (15 miles) north of Yelverton Pass (locality 10), where the value is about 25 degrees. The formation pinches out rapidly toward the southeast and is not present at locality 5. There, unit A is overlain directly by Triassic shale. The Nansen Formation comprises interbedded light- and medium-grey lime mudstone and skeletal grainstone. Corals and brachiopods are present in the upper part. The thinning and disappearance of the Nansen Formation is caused by two phases of erosion; the first and major phase occurred prior to deposition of the Trold Fiord Formation, and is indicated by karst features in the uppermost 10 m (33 ft.) of the formation at locality 4. The second erosional event occurred subsequent to Trold Fiord deposition and, at locality 5, removed the Trold Fiord Formation as well as the entire Nansen Formation. In addition to erosion, the thinning of the Nansen Formation also may have been caused by onlap of the carbonates onto the tilted surface of unit A.

No complete sections of the Nansen Formation were examined in the area around Yelverton Inlet, owing to the nature of terrain and weather. The basal part of the formation overlies unit C with sharp, probably disconformable contact and consists of rubbly, massive limestone of probable algal origin but without macroscopically discernible texture. The uppermost part consists of medium- to coarse-grained calcirudite and is profusely fossiliferous. In the Yelverton Inlet area the formation exceeds 700 m (2300 ft.) in thickness and is overlain with sharp, probably disconformable, contact by unit D.

Belcher Channel Formation

At locality 8 the thickness of the Belcher Channel Formation does not exceed 100 m (330 ft.). Only the lower part is well exposed and consists of nodular dark grey and maroon limestone. At this locality the Belcher Channel Formation is overlain by white, fine grained sandstone which has been tentatively assigned to the Triassic.

Unit D

Unit D has been established for mapping purposes in the Yelverton Inlet area and includes a variety of rocks between the Nansen Formation and the Triassic sequence. This unit is estimated to be between 300 m and 400 m (1000 ft. and 1300 ft.) thick in most areas but, locally, it may be considerably thinner.

Three laterally separate, lithological sequences make up unit D. Sequence 1 consists of very dark grey calcareous shale, interbedded with limestone. The limestone varies from silty and argillaceous lime mudstone to skeletal packstone and grainstone. Sequence 2 comprises siltstone. The siltstone is calcareous and of dark grey or dark brown colour. Sequence 3 is a succession of limestones, mostly skeletal grainstone. Conspicuous in the latter sequence is a bright red interval. All three sequences are profusely fossiliferous. The stratigraphical relationships of the three sequences to each other are not known at present.

Unit D probably corresponds to Thorsteinsson's (1974, p. 17 and 21) stratigraphic sequences 4 (Lower Permian marine sequence) and 5 (Upper Permian marine sequence).

Trold Fiord Formation

At locality 4 the Trold Fiord Formation is 1.5 m (5 ft.) thick and consists of thin-bedded, green siltstone with abundant corals. The relationship between the Trold Fiord Formation and unit D is not known for certain and interpretation depends on identification of the faunas collected from unit D.

References

Christie, R.L.
1965: Triassic disconformity in the Tanquary Fiord Yelverton Pass region, Ellesmere Island, N.W.T.; Geol. Surv. Can., Paper 65-2, p. 67.

Nassichuk, W.W. and Christie, R.L.
1969: Upper Paleozoic and Mesozoic stratigraphy in the Yelverton Pass region, Ellesmere Island, District of Franklin; Geol. Surv. Can., Paper 68-31.

Thorsteinsson, R.
1974: Carboniferous and Permian stratigraphy of Axel Heiberg Island and western Ellesmere Island, Canadian Arctic Archipelago; Geol. Surv. Can., Bull. 224.

Trettin, H.P.
1969: Geology of Ordovician to Pennsylvanian rocks, M'Clintock Inlet, north coast of Ellesmere Island, Canadian Arctic Archipelago; Geol. Surv. Can., Bull. 183.

STUDIES OF MESOZOIC STRATIGRAPHY, TANQUARY FIORD TO YELVERTON PASS, NORTHERN ELLESMERE ISLAND, DISTRICT OF FRANKLIN

Project 730051

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The upper Paleozoic and Mesozoic geology north of Tanquary Fiord was investigated first in 1963 and 1966 by Nassichuk and Christie (1969) and they briefly described the Triassic Bjorne, Schei Point, and Heiberg formations, an unnamed succession of Jurassic-Lower Cretaceous shales and sandstone, and the Cretaceous Isachsen Formation. They recognized and named the Tanquary Structural High (Figs. 95.1, 95.2), a belt in which the upper Paleozoic to lower Upper Triassic succession is absent and in which the Upper Triassic-Lower Jurassic Heiberg Formation lies directly on the lower Paleozoic basement. Near the southeastern limit of the high there is a major fault (here informally referred to as the Ekblaw fault -- for Ekblaw Lake) that subsequently was included in the Lake Hazen Fault Zone, a belt of imbricated, southeastward-directed thrust faults extending from north of Hare Fiord to north of Alert. The Lake Hazen Fault Zone places the Lower Ordovician and/or Cambrian Grant Land Formation on strata ranging in age from Ordovician to Tertiary (Trettin, 1971). The frontal thrust of that zone lies south of the Viking Ice Cap and trends in a more easterly direction than the Ekblaw fault.

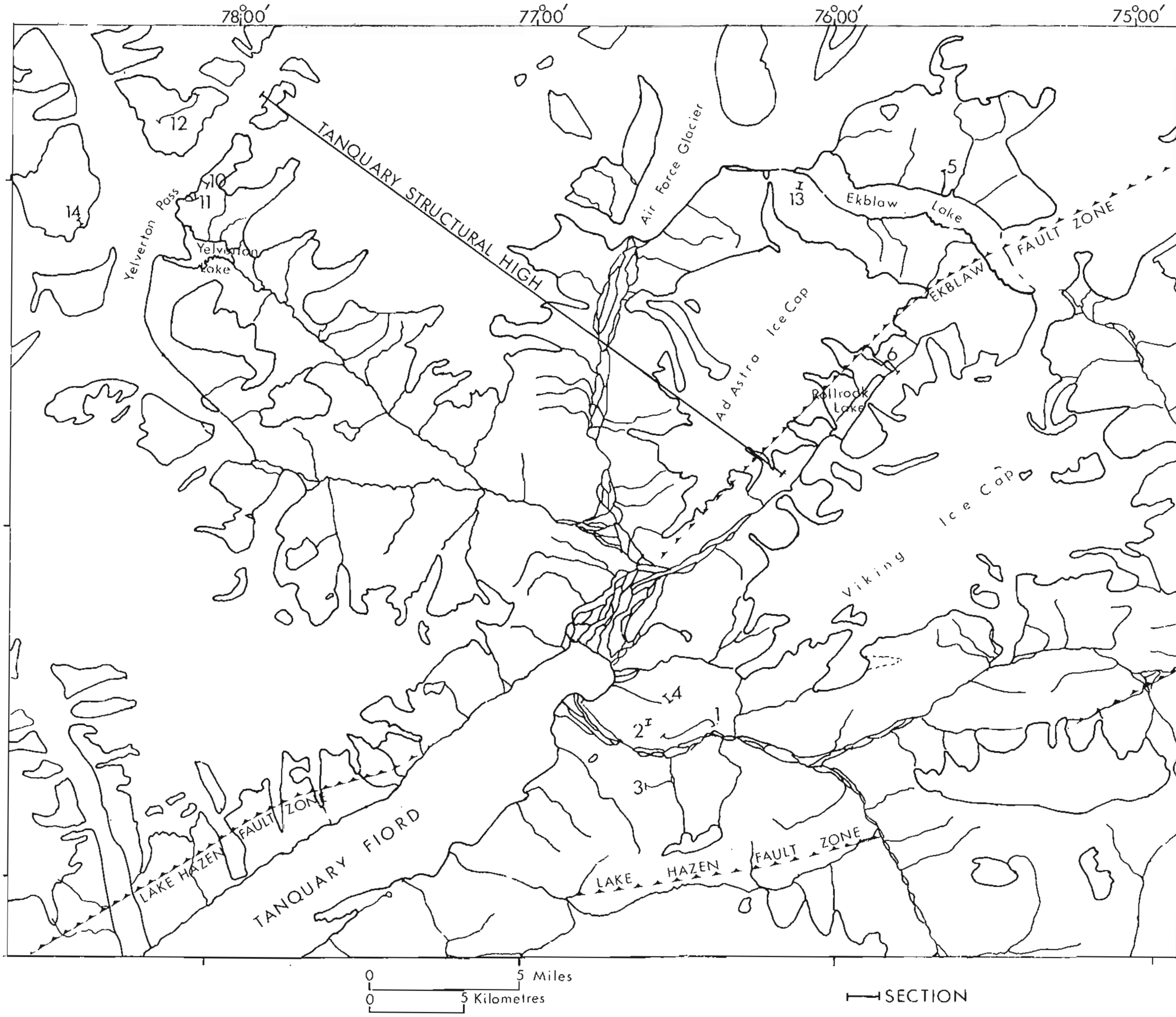

Figure 95.1. Index map of Tanquary Fiord, northern Ellesmere Island.

During the 1975 field season, the writer measured detailed sections of the entire Mesozoic succession, subdivided the Jurassic-Lower Cretaceous succession and recognized several other formations not previously known from this area. The results of this will be integrated with a continuing subsurface and surface study of the eastern Sverdrup Basin (Project 750091). In this preliminary study, brief comments on individual formations will be followed by some conclusions about the history of the Tanquary Structural High. For background information on the various units mentioned, the reader is referred to Nassichuk and Christie (1969) and summaries by Tozer (1961, 1963, 1970).

Bjorne Formation

The Lower Triassic Bjorne Formation is limited mainly to the area south of the Ekblaw fault where it overlies upper Paleozoic strata of the Sabine Bay Formation and is overlain by the Schei Point Formation. The lower three quarters of the formation consist of reddish weathering, medium to very thick bedded bioclastic limestone. The formation is 186.5 m thick at section 1 and 166.5 m thick at section 3.

A small outcrop, perhaps 50 m long, was found north of the Ekblaw fault at section 14 where a 9-m-thick sandstone unit appears to fill a topographic depression in the unconformity developed on the Nansen Formation.

Blaa Mountain Formation

The Blaa Mountain Formation, not previously known from this region, is now recognized north of the Tanquary Structural High where it unconformably overlies upper Paleozoic strata and is conformably overlain by the Schei Point Formation. Occurrence of *Gryphaea* demonstrate that the strata represent only a lower Upper Triassic (Karnian) tongue of the formation which elsewhere includes the entire Middle Triassic series.

At section 12, the formation is 66.5 m thick. The lower 17 m consists of rhythmic deposits grading from clayey siltstone to argillaceous sandstone. The upper 49.5 m is composed of rhythmic deposits grading from argillaceous sandstone to clayey siltstone.

Schei Point Formation

Nassichuk and Christie (1969) recognized that the Schei Point Formation is absent on top of the Tanquary Structural High but present on the flanks. This unit represents the uppermost (Karnian) part of more complete Schei Point sections elsewhere and suggests onlap onto the Tanquary Structural High.

The Schei Point is 28.5 m thick at section 3 and is composed mainly of medium-grained, medium- to thick-bedded, calcareous sandstone. Paleocurrent indicators (solitary troughs) suggest sediment dispersal away from the Tanquary Structural High.

Heiberg Formation

The upper Upper Triassic-Lower Jurassic Heiberg Formation is the most widely distributed Mesozoic formation in the Tanquary-Yelverton Pass region. Four detailed sections at localities 1, 3, 10 and 12 confirm the summary descriptions of Nassichuk and Christie (1969). It should be mentioned, however, that, in the upper part of the formation (upper 30 m of sec. 13 and upper 32 m of sec. 10), there are some dusky red weathering sandstones that resemble the Lower Jurassic (Pliensbachian) Borden Island Formation of Fosheim Peninsula both in lithology and ichnofauna. Positive identification as the Borden Island Formation, however, must await paleontological evidence.

Jurassic-Lower Cretaceous Sequence

The previously undifferentiated Jurassic-Lower Cretaceous sequence is now subdivided into the following units: Lower Savik Formation, Jaeger Formation, Upper Savik Formation, an unnamed formation, and Deer Bay Formation. The unnamed formation represents argillaceous equivalents of the Awingak Formation. Measured thicknesses of the total sequence differ around the Tanquary Structural High as follows: section 4, 46.5 m; section 2, 180 m; section 6, 540 m; section 5, 697 m; section south of Air Force Glacier (Christie, pers. comm., 1975), 198 m; sections 10 and 11 (composite section), 300 m.

Isachsen Formation

At section 6, the Isachsen Formation is 238.5 m thick and consists mainly of medium bedded to massive, fine- to coarse-grained, calcareous sandstone. The main succession is considered to be fluvial in origin, however 9 m of burrowed sandstone at the base and 49.5 m of clayey siltstone at the top are both interpreted as marine. At section 11, 31.5 m of probably marine burrowed sandstone are assigned to the Isachsen Formation. These sandstones may represent the upper part of the Deer Bay Formation.

Christopher Formation

The Christopher Formation, not previously reported from this area, was recognized at section 6. There, it is 162 m thick and conformably overlies the Isachsen Formation. At section 11, 12 m of clayey siltstone overlie the Isachsen Formation. The siltstone is tentatively assigned to the Christopher Formation pending paleontological confirmation.

The formation consists of dark grey, jarosite-stained clayey siltstone with minor interbeds of dark reddish brown, ferruginous, clayey siltstone layers and nodules. Vertical and horizontal burrows are common in the ferruginous siltstone and rare in the dark grey siltstone.

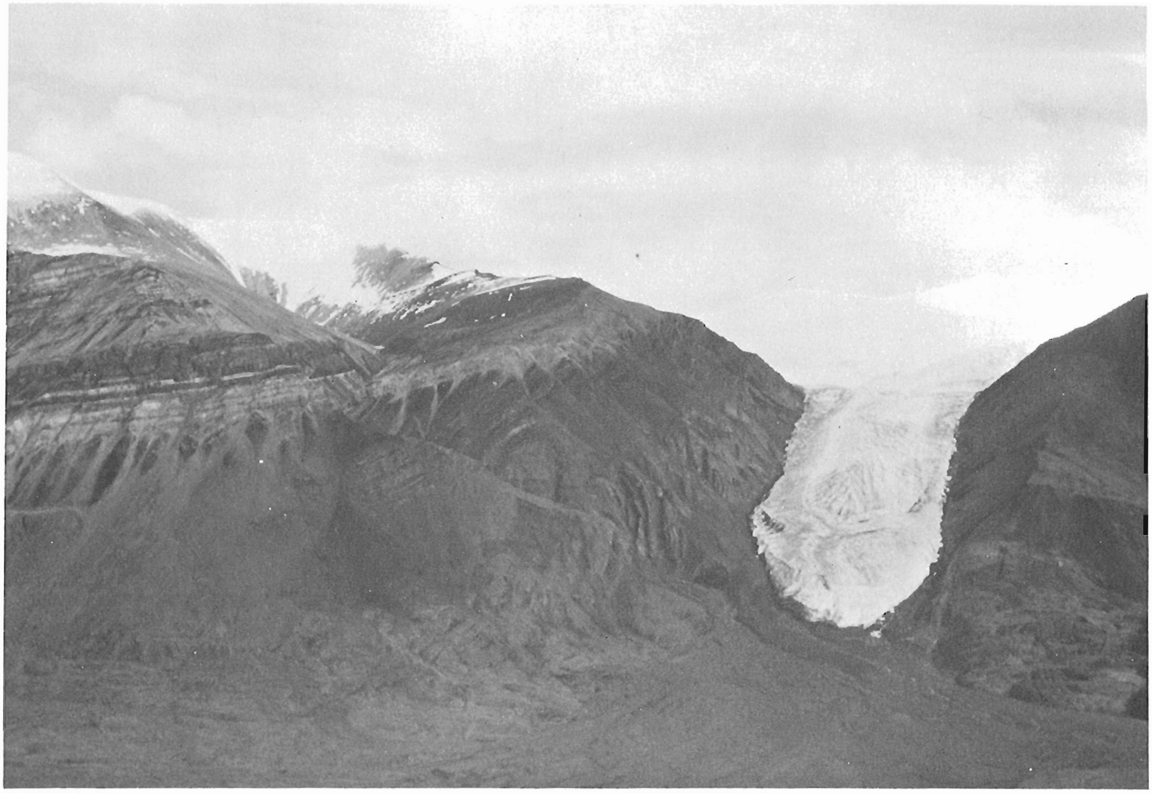

Figure 95.2. The Tanquary Structural High northwest of Tanquary Fiord, looking east. The angular unconformity separates the Grant Land Formation and the Lower Heiberg Formation.

Mesozoic history of the Tanquary Structural High

One of the main problems concerning the Tanquary Structural High is whether the Carboniferous to lower Upper Triassic succession is absent due to nondeposition or to erosion. Discovery of an outlier of the Bjorne Formation north of the Ekblaw fault suggests that the Bjorne Formation originally was more extensive but subsequently was eroded during Middle to early Late Triassic time. Significant subsidence of the Tanquary Structural High occurred during the Karnian when representatives of the Blaa Mountain and Schei Point formations was deposited. The subsidence, however, was more pronounced on the flanks of the high than over the high itself. This is apparent from the fact that Blaa Mountain and Schei Point formations both increase in thickness away from the high. Thus, on the northwest flank of the Tanquary Structural High, the Blaa Mountain Formation increases from 66.5 m at section 12 to 200 m at the head of Yelverton Inlet, and the Schei Point Formation from 101.5 to 225 m. On the southeastern flank of the high, the Schei Point Formation is 12 m thick at section 1 and 28.5 m thick at section 3. The Schei Point Formation is absent north of Rollrock Lake due to either nondeposition or erosion prior to Heiberg sedimentation.

References

Nassichuk, W.W. and Christie, R.L.
1969: Upper Paleozoic and Mesozoic stratigraphy in the Yelverton Pass region, Ellesmere Island, District of Franklin; Geol. Surv. Can., Paper 68-31.

Tozer, E.T.
1961: Triassic stratigraphy and faunas, Queen Elizabeth Islands, Arctic Archipelago; Geol. Surv. Can., Mem. 316.

1963: Mesozoic and Tertiary stratigraphy, western Ellesmere Island and Axel Heiberg Island, District of Franklin (preliminary account); Geol. Surv. Can., Paper 63-30.

1970: Geology of the Arctic Archipelago; Mesozoic; *in* Geology and economic minerals of Canada; Geol. Surv. Can., Econ. Geol. Rept. no. 1, p. 83-179.

Trettin, H.P.
1971: Geology of lower Paleozoic formations, Hazen Plateau and southern Grant Land Mountains, Ellesmere Island, Arctic Archipelago; Geol. Surv. Can., Bull. 203.

EUREKA SOUND AND BEAUFORT FORMATIONS, YELVERTON BAY, ELLESMERE ISLAND, DISTRICT OF FRANKLIN

Project 730051

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Clastic sediments west of Yelverton Bay, assigned to the Eureka Sound "Group" by Christie (1957) were re-examined by the writer during the 1975 field season. Approximately 700 m (2250 ft.) of poorly exposed, folded and faulted strata are present that are assigned here to the Upper Cretaceous to Lower Tertiary Eureka Sound Formation and the Upper Tertiary Beaufort Formation. Two locations were examined within the fault block at Yelverton Bay. The strata are preserved in an easterly dipping fault block bordered by lower Paleozoic formations on the east and south and by ice on the west. The southern boundary fault is a north-directed thrust (*see* Fig. 93.11); the eastern fault is concealed. Several reverse faults within the fault block offset both Eureka Sound and Beaufort formations. Faults cutting the Eureka Sound Formation are best exposed at locality 2 (Fig. 96.1). There, the basal strata of the formation are intensely crumpled and faulted and the deformation decreases away from the fault and stratigraphically upward.

Eureka Sound Formation

The Eureka Sound Formation consists of a minimum (base not exposed) of 500 m (1600 ft.) of interbedded clayey siltstone and sandstone. In the lower part of the succession, siltstone beds (8-12 cm thick) predominate over sandstone beds (2-4 cm thick); in the upper part, sandstone beds (10-22 cm thick) predominate over siltstone beds (2-4 cm thick). The siltstone is dark grey, micaceous, has an estimated clay content of about 30 to 40 per cent, and contains macerated plant fragments. Siltstone beds commonly include laminae of sandstone that form flaser structures. The sandstone beds are mostly very light grey, friable, and comprise subrounded, moderately well sorted and fine grains; these beds range up to medium grained in the upper part of the section. Minor amounts of climbing ripple-marks of low amplitudes were the only sedimentary structures observed in the sandstone. Fragments of silicified wood, up to 10 cm long, occur in the upper part of the formation.

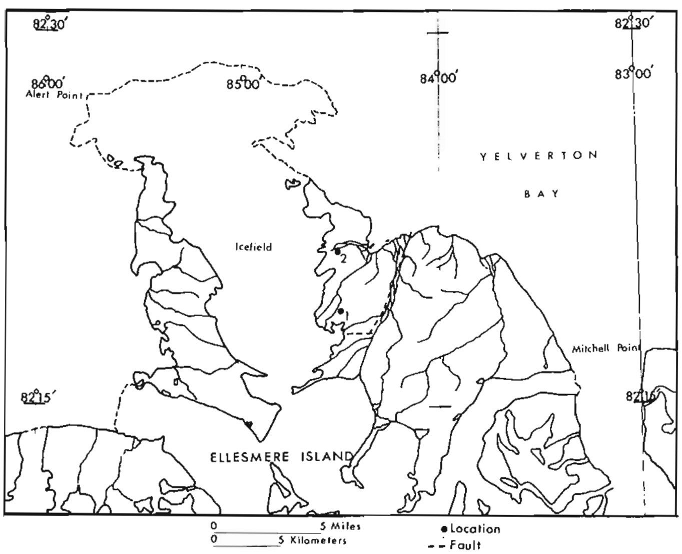

Figure 96.1. Index map of Yelverton Bay, northern Ellesmere Island showing localities for the Beaufort and the Eureka Sound formations.

The alternation of clayey siltstone and sandstone associated with flasers and low-amplitude climbing ripple-marks is characteristic of intertidal deposits (Reineck and Wunderlich, 1968). The upward transition to a sand-dominated facies suggests increasing proximity of a shoreline. The succession as a whole, therefore, characterizes a regressive sequence.

A fragment of *Baculites* (?) (paleontological identification not yet received) from locality 2 suggests an age not younger than Late Cretaceous. Several florules identified by W.S. Hopkins, Jr. (GSC locs. C-55017, C-55019-C55021) are of unspecified Maastrichtian to Eocene age, but one (GSC loc. C-55015), from locality 1, is definitely Early Tertiary.

Beaufort Formation

Unfossiliferous sandstone and conglomerate unconformably overlying the Eureka Sound Formation west of Yelverton Bay are assigned here to the Beaufort Formation. The Beaufort Formation at Yelverton Bay is about 200 m (640 ft.) thick and consists of rhythmic deposits composed of a basal conglomerate up to 1.5 m (4.8 ft.) thick (locally absent) and an overlying sandstone, 3 to 4 m (9.6-12.8 ft.) thick, that becomes finer grained upward. The basal conglomerates are channel deposits with erosional bases. The width of the channels at Yelverton Bay (loc. 1) is estimated to be 40 to 60 m (128-192 ft.). Phenoclasts there range from 10 to 80 cm and the mode is about 20 cm. They are composed of schist, gneiss, chert, and quartzite with minor amounts of marble, gabbro, and sandstone. The grain size of the matrix ranges from medium to very coarse grained with a coarse grained mode. Matrix of the conglomerate is composed mainly of quartz and rock fragments. The grain size of the sandstone beds above the conglomerate ranges from very coarse to medium grained and the grains are compositionally similar to those forming the matrix of the conglomerate.

At locality 2, the phenoclasts range from 0.2 to 7 cm, the mode being 2 cm. The matrix is a poorly sorted, very fine to very coarse grained, argillaceous sandstone consisting of quartz and rock fragments. The poorly exposed overlying sandstone beds are compositionally similar to the matrix of the conglomerate.

The phenoclasts are comparable to lower Paleozoic and older rocks exposed to the south. The presence of marble among them suggests proximity of the source area (Kuenen, 1956). Northeasterly flowing paleocurrents are inferred from the orientation of channels at locality 1 and the decrease in clast size from locality 2 to locality 1. The depositional environment is interpreted as the flood plain of a braided stream (cf. Picard and High, 1973).

The Beaufort Formation was defined (Tozer, 1956; Tozer and Thorsteinsson, 1964) to include Upper Tertiary or Pleistocene gravel and sand of the Arctic Coastal Plain lying unconformably on strata ranging in age from Devonian to Early Cretaceous. These deposits and their abundant floras were restudied by Hills (1969, and *in* Hills *et al.*, 1974 and Hills and Ogilvie, 1970) who assigned them a middle to late Miocene age (pers. comm., 1975). Cobble conglomerate at Gibs Fiord, east-central Axel Heiberg Island, previously (provisionally) included in the Eureka Sound Formation, has been reassigned to the Beaufort Formation on the basis of the age of pine cones (Balkwill and Bustin, 1975). The sandstone and conglomerate unconformably overlying the Eureka Sound Formation west of Yelverton Bay are assigned to the Beaufort Formation because of their similarity to the Gibs Fiord strata.

Conclusions

Balkwill *et al.* (1975) recognized three phases of the Eurekan Orogeny from their study at Flat Sound, Axel Heiberg Island; "a late Maastrichtian/late Paleocene phase of local uplift; a (?)middle Eocene/Oligocene phase of regional compression, and a Miocene/(?) Pliocene phase of local uplift". The deposits studied at Yelverton Bay record at least four phases of uplift in northern Ellesmere Island. However, these uplifts are not necessarily contemporaneous with those recorded from Axel Heiberg Island. (1) The clastic sediments of Eureka Sound Formation were derived from contemporaneous (latest Cretaceous to Early Tertiary) uplifts. (2) Tilting of Eureka Sound Formation resulted in a low angular unconformity and partial erosion prior to Beaufort Formation deposition. Any evidence of mid-Tertiary (Oligocene to Miocene?) sedimentation would have been removed during that period of erosion. (3) The Late Tertiary(?) uplift of the Grant Land Mountains that produced the conglomerate of the Beaufort Formation in the Boulder Hills northeast of Lake Hazen (Christie, 1974) and in the present area were deposited during the Late Tertiary (?) in response to uplift. (4) Faults cutting the Beaufort Formation demonstrate a still later phase of deformation.

The presence of Tertiary deposits at Yelverton Bay indicates that the thick Cenozoic clastic wedge,

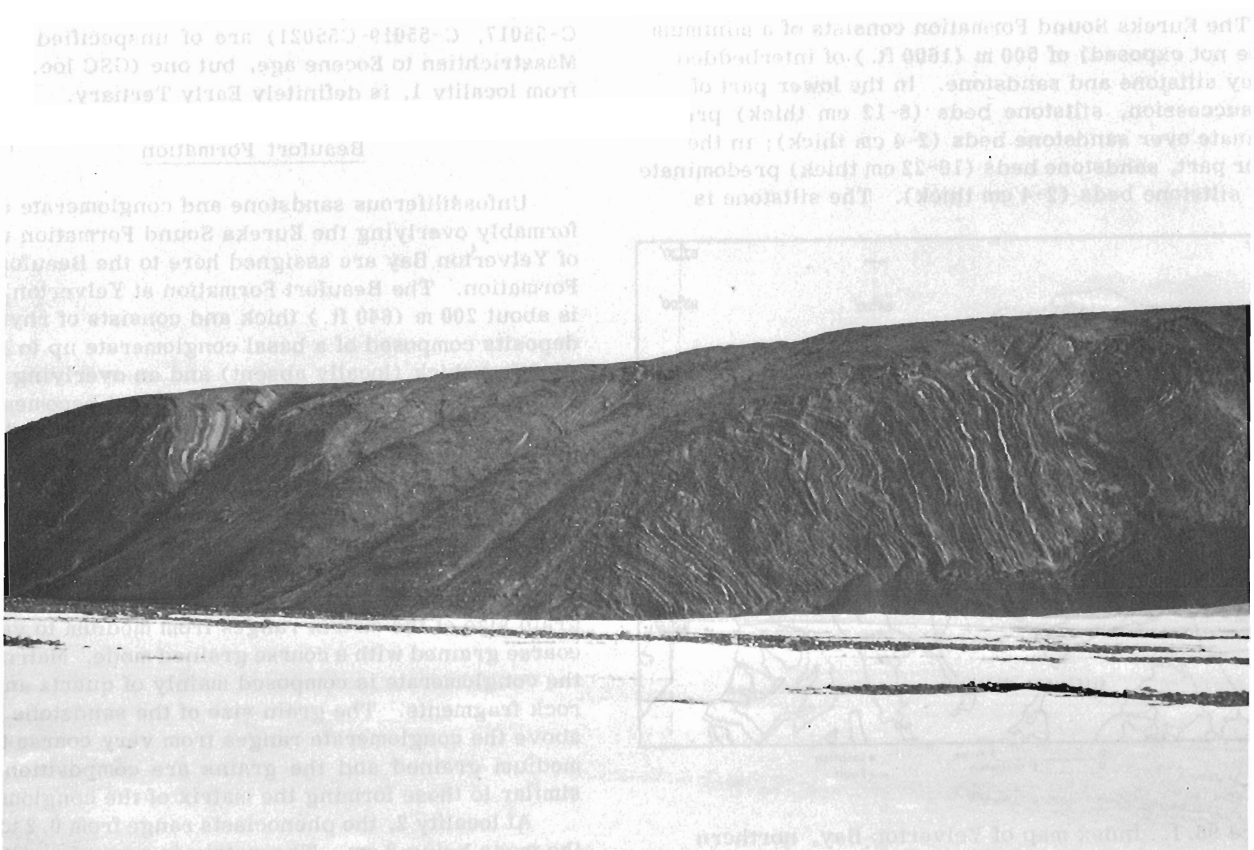

Figure 96. 2. Deformed basal strata of Eureka Sound Formation at locality 2 (Fig. 96. 1); view to the south.

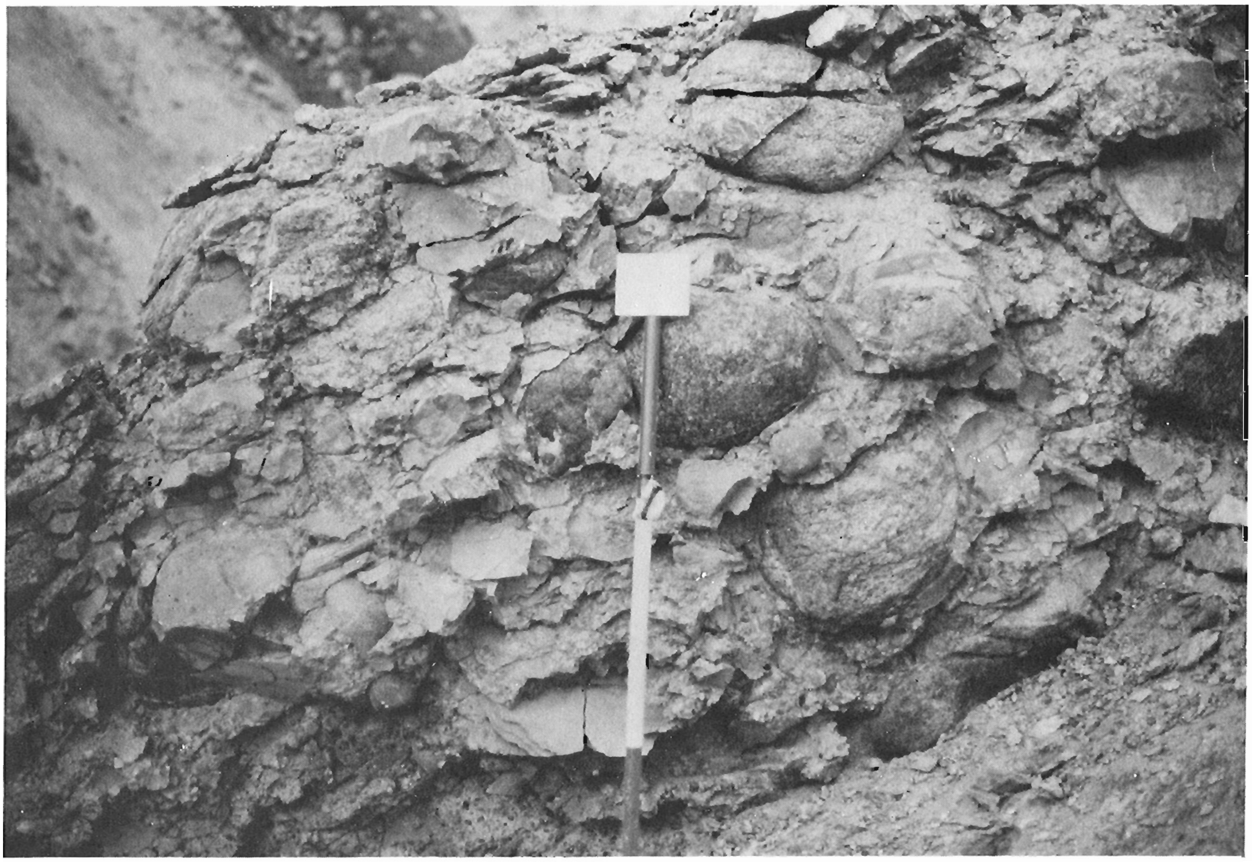

Figure 96.3. Basal conglomerate in the Beaufort Formation at locality 1. Light and dark sections on the Jacob's staff are 0.5 m.

inferred off the western Queen Elizabeth Islands (from geophysical studies by Sobczak and Weber, 1973) probably continues along the continental margin to northern Ellesmere Island.

References

Balkwill, H.R. and Bustin, R.M.
1975: Stratigraphic and structural studies, central Ellesmere Island and eastern Axel Heiberg Island, District of Franklin; *in* Report of Activities, Part A, Geol. Surv. Can., Paper 75-1A, p. 513-517.

Balkwill, H.R., Bustin, R.M. and Hopkins, W.S., Jr.
1975: Eureka Sound Formation at Flat Sound, Axel Heiberg Island, and chronology of the Eurekan Orogeny; *in* Report of Activities, Part B, Geol. Surv. Can., Paper 75-1B, p. 205-207.

Christie, R.L.
1957: Geological reconnaissance of the north coast of Ellesmere Island, District of Franklin, Northwest Territories; Geol. Surv. Can., Paper 56-9.

Christie, R.L. (cont.)
1974: Northeastern Ellesmere Island: Lake Hazen region and Judge Daly Promontory; *in* Report of of Activities, Part A, Geol. Surv. Can., Paper 74-1A, p. 297-299.

Hills, L.V.
1969: Beaufort Formation, northwestern Banks Island, District of Franklin; *in* Report of Activities, Part A, Geol. Surv. Can., Paper 69-1A, p. 204-207.

Hills, L.V., Klovan, J.E., and Sweet, A.R.
1974: *Juglans eocinera* n. sp., Beaufort Formation (Tertiary), southwestern Banks Island, Arctic Canada; Can. J. Bot., v. 52, p. 65-90.

Hills, L.V. and Ogilvie, R.T.
1970: *Picea banksii* n. sp., Beaufort Formation (Tertiary), southwestern Banks Island, Arctic Canada; Can. J. Bot., v. 48, p. 457-464.

Kuenen, P.H.
1956: Experimental abrasion of pebbles, 2. Rolling by current; J. Geol., v. 64, p. 336-368.

Picard, H.R. and High, L.R., Jr.
1973: Sedimentary structures of ephemeral streams; Developments in sedimentology, v. 17, Elsevier, Amsterdam.

Reineck, H.E. and Wunderlich, F.
1968: Classification and origin of flaser and lenticular bedding; Sedimentology, v. 11, p. 99-104.

Sobczak, L.W. and Weber, J.R.
1973: Crustal structure of Queen Elizabeth Islands and Polar Continental Margin, Canada; *in* Proc. Second Intern. Symp. on Arctic Geology; Am. Assoc. Pet. Geol., Mem. 19, p. 517-525.

Tozer, E.T.
1956: Geological reconnaissance, Prince Patrick, Eglinton and western Melville Islands, Arctic Archipelago; Geol. Surv. Can., Paper 55-5.

Tozer, E.T. and Thorsteinsson, R.
1964: Western Queen Elizabeth Islands, Arctic Archipelago; Geol. Surv. Can., Mem. 332.

97. STRUCTURAL AND STRATIGRAPHIC STUDIES IN THE NORTHERN CANADIAN CORDILLERA

Projects 610007 and 690005

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During the 1975 field season, the writer spent approximately three weeks studying structural and stratigraphic problems brought to light by final compilations of the Operation Porcupine area (Norris, 1963, p. 17-19). In all, 145 helicopter stops were made, approximately one half as many as that accomplished by the writer over a three and one-half month period during the first full year (1962) of the project. The increased efficiency and scientific return resulted from the growing knowledge and understanding of the geological framework of the region, and added significantly to the completion of maps and reports. Of the eighteen 1: 250 000 scale geological maps comprising the project area, sixteen are now in manuscript form and five are on open file (Norris, 1975a, b). There remain to be compiled the west half of the Porcupine River map-area (116J, K) and all of the Ogilvie River map-area (116F, G).

In attempting to summarize the highlights of the summer's work, the writer will present his findings more or less in stratigraphic order beginning with the oldest rocks, and ending with a discussion of two structural problems.

The writer re-examined the lower contact of the Cambrian volcanic and carbonate unit (Unit B, Geol. Surv. Can. Map 10-1963) in Romanzof Uplift at the headwaters of Mountain Creek. There, and elsewhere in Canada, he recognized a persistent, 150-m-(500 ft.) thick unit of dark olive-grey to dark grey, slaty argillite resting conformably beneath it. The argillite, as well as the volcanics and carbonates, commonly strike in the same direction as the underlying Neruokpuk Formation, but it may differ significantly in dip. Mapping revealed, moreover, that the argillite is in contact with a considerable thickness and variety of lithologies of Neruokpuk across the strike. Thus, it appears plausible that a major unconformity separates the two successions and that the Neruokpuk was deformed prior to deposition of the overlying rocks (Norris, 1974, p. 33). This argument is supported further by the structural position of the argillite-volcanic-carbonate assemblage. It does not rest on the youngest unit of the Neruokpuk, which presumably would have been the case if the Neruokpuk were more or less undeformed prior to deposition of the assemblage.

The grey argillite unit is herein excluded from the Precambrian Neruokpuk. It is probably Cambrian in age along with the volcanics and carbonates already reported (Dutro *et al.*, 1972, p. 183) to contain Early as well as Late Cambrian trilobites.

In the course of final compilation of the geology of Romanzof Uplift, it became apparent that an argillite and chert unit occurred commonly on the northeast flank of the structure and that the facies is not unlike part of the lower Paleozoic Road River Formation. At one helicopter stop, approximately 10 km (6 miles) southwest of Komakuk Beach in Demarcation Point map-area (lat. 69°32.5' North; long. 140°19' West), T. Poulton, who was travelling with the writer at the time, discovered a number of specimens of *Monograptus* cf. *M. transgrediens praecipuus* (Přibyl) in loose pieces of dark blue-grey, phyllitic argillite identical to that of the immediate bedrock. According to B. S. Norford who identified the collection, the graptolites indicate a latest Silurian age for the host rock. The precise location of the fossil locality (GSC loc. C-41895) may be positioned on NAPL vertical aerial photograph A13231-70 at the following Cartesian co-ordinates measured with respect to the centre of the picture where the north direction corresponds to the positive Y axis: X = +2.33 cm, Y = -5.09 cm.

The stratigraphic position of these uppermost Silurian rocks above the Neruokpuk and below unexposed beds which, from regional mapping, are interpreted to be part of the Jura-Cretaceous Kingak Formation along the Arctic Coastal Plain, would suggest strongly that these rocks are part of a seaward-thickening prism, unconformity bounded, coeval with the upper member of the Road River Formation. They support the hypothesis of progressive encroachment upon the carbonate banks in Ordovician and Silurian time (Norford, 1964, p. 67) and they can be expected in the subsurface of southern Beaufort Shelf.

Because of the similar appearance in hand specimen of dark-coloured argillite of the Neruokpuk Formation and those from which graptolites were found on the north flank of Romanzof Uplift, X-ray diffraction analyses were run on a number of fresh rock samples in an attempt to discover, in the absence of fossils, any basic compositional differences which might assist in the sorting out of these formations. According to A. Heinrich (Internal Reports, 1975), the XRD analyses are as follows in Table 97.1.

It is apparent from the table that, within the limits of the samples examined, the fossiliferous rocks (383NC1 and 1173NC1) lack the metamorphic minerals pyrophyllite and chlorite but contain significantly more quartz than the Precambrian Neruokpuk Formation. Sample 927NC1 was collected at the reported position from which a single specimen of *Orthograptus* sp. of Late Ordovician age was found (Dutro *et al.*, 1972). Compositionally, the sample does not appear to differ significantly from those of other areas of Neruokpuk. It is suggested that these Ordovician rocks along the Alaska-Yukon boundary may belong to a thin, structurally depressed slice of the landward feather edge of the prism of lower Paleozoic graptolitic rocks flanking Romanzof Uplift on the northeast.

With the completion of mapping on a scale of 1: 250 000 in the vicinity of Barn Mountains, yet another area underlain by grey, porphyritic, biotite granite was found approximately 4 km (2.5 miles) south of Mount Fitton and 1.6 km (1 mile) south of Boulder Creek. It may be located on NAPL vertical aerial photograph A14406-66 centred around a point at Cartesian co-ordinates X = +4.7 cm; Y = -4.7 cm measured with respect to the centre of the picture. The contact with overlying rocks was not evident and the recessive, covered interval between the granite and still higher conglomerates of the Aptian-Albian flyschoid sequence is presumed to be occupied by the Lower Cretaceous shales and siltstones laterally equivalent to those from which a Berriasian fauna was first collected by geologists of Triad Oil Company in 1959 at the headwaters of Fitton Creek (GSC loc. 44163). The occurrence is presumed to be a cupola of the Mount Fitton stock, connected downward with the main batholith, and dated at approximately 370 m.y. (Norris, 1964, p. 22).

To obtain additional control on thickness and facies of the southwestward continuation of the Jurassic and Lower Cretaceous clastic sequence into northern Kandik Thrust Belt, the writer revisited one of the few relatively unfaulted sections embracing this interval at the headwaters of Kandik River (Sec. 116F14) in northwestern Ogilvie River map-area (116F, G). There, in collaboration with T. Poulton, the writer measured and described approximately 1036 m (3400 ft.) of interbedded sandstone and shale resting disconformably on unnamed ?Middle Triassic limestone and overlain by a covered interval presumed to contain shale and siltstone coeval

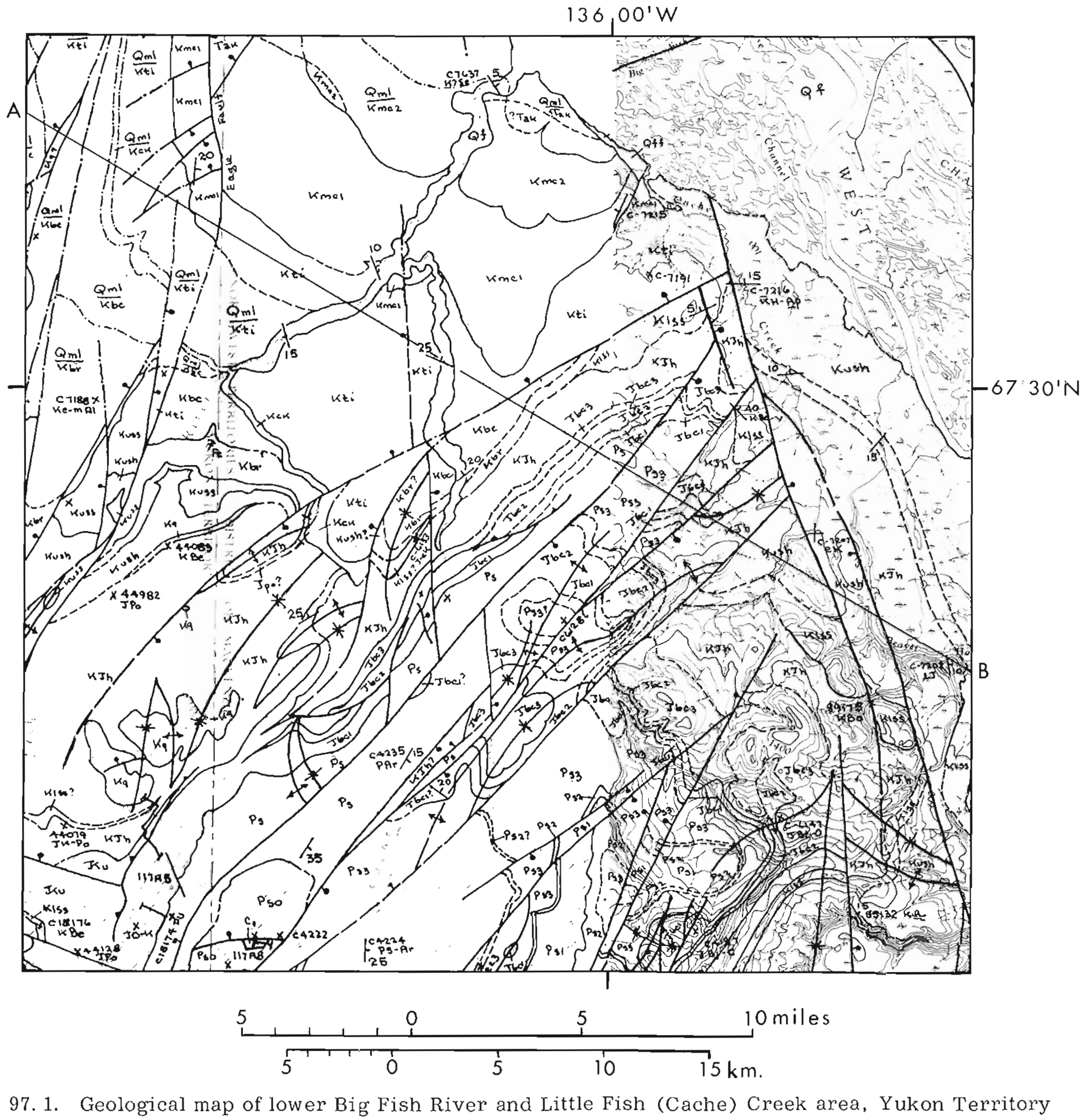

Figure 97.1. Geological map of lower Big Fish River and Little Fish (Cache) Creek area, Yukon Territory and northwestern District of Mackenzie, showing stratigraphy and structure on and adjacent to Cache Creek Uplift. Symbols used are essentially those defined in Open Files 279 and 302.

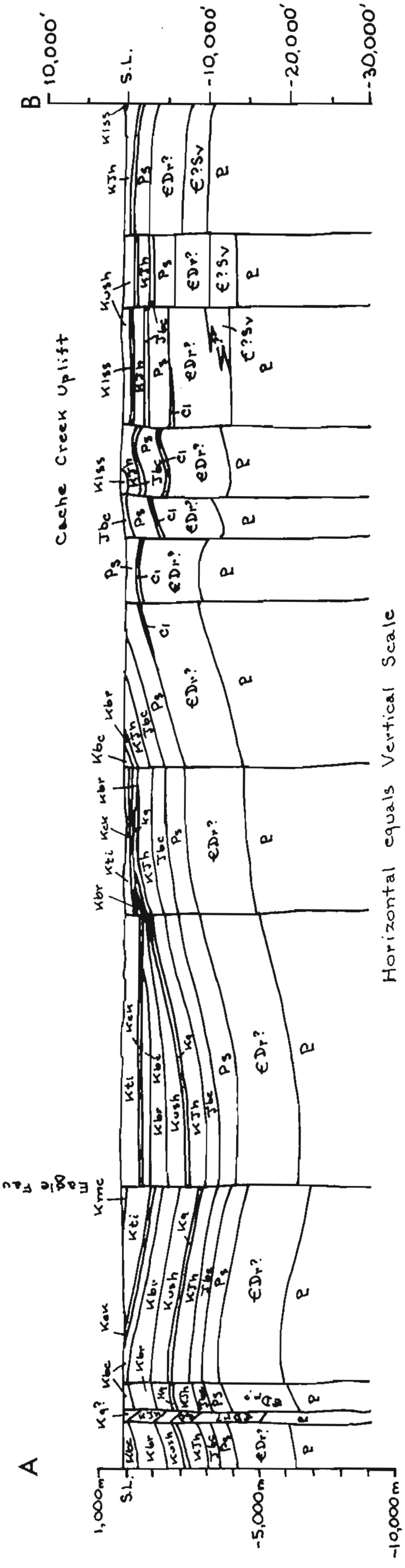

Figure 97.2. Structure section across the lower Big Fish River and little Fish (Cache) Creek area, Yukon Territory and northwestern District of Mackenzie, showing structure and stratigraphic relationships across Cache Creek Uplift. Line of section is shown on Figure 97.1. Symbols used are essentially those defined in Open Files 279 and 302.

with the "upper shale siltstone division" of the Rat Pass area.

The approximate position of the centre of the section is at longitude 140°27' West and latitude 65°43.0 North. Its precise position may be located on NAPL vertical aerial photograph A13231-147. The Cartesian co-ordinates of the base are: X = +1.00 cm, Y = -7.69 cm; and of the top, X = +3.70 cm, Y = -9.50 cm. It corresponds to the more southerly of the two sections described compositely by J. A. Jeletzky (1971, p. 218, 219) under the heading "Headwaters of Kandik River".

No additional faunal control was established and the reader is referred to the account of a few poorly preserved fossils reported by Jeletzky (ibid.) from the basal beds of the lowest carbonaceous, quartz, chert sandstones. They occur approximately 122 m (400 ft.) above the base of the writer's section. A sample of bright coal collected 389 m (1275 ft.) above the base was examined by A. Cameron and reported by him (pers. comm., 1975) to be anthracitic. The high rank of this coal suggests deep initial burial and/or thermal metamorphism of Cretaceous and older sediments in the Kandik Thrust Belt, conditions that are not favourable for the possible occurrence of natural hydrocarbons at stratigraphic levels below the Upper Cretaceous in the region.

Approximately one half of the measured Jura-Cretaceous interval is occupied by mature, epicontinental, carbonaceous, quartz, chert sandstone strongly resembling the "white and coaly quartzite division" of the Aklavik Range area. The remainder, including the covered intervals, is presumably occupied entirely by shale and siltstone. No feldspar grains were noted in thin sections of the sandstones and their quartz content is estimated to be consistently about 80 per cent. This mineralogical maturity is, therefore, consistent with that reported by Young (1974a, p. 198) for the coeval epicontinental phase of clastic sedimentation in northern Richardson Mountains.

The top of the section is in the hinge of a prominent, north-trending and north-plunging syncline containing strata equivalent to the "upper shale siltstone division" as well as feldspathic sandstones and conglomerates of the Aptian-Albian flyschoid sequence. A west-dipping reverse fault which cuts transversely across the hinge of the fold repeats part of the Jura-Cretaceous clastic sequence immediately north of the section so that the rapid southward thinning of this interval reported by Jeletzky (1971, p. 218, 219) can be attributed to the southward decrease in stratigraphic separation along the fault and not necessarily to paleogeographic considerations.

Table 97. 1

Sample No.	Formation	GSC Locality	Age	Mineral Composition					
				Illite	Pyrophyllite	Chlorite	Quartz	Pyrite	Feldspar
383NC1	unnamed	C-4236	E. Devonian	27%[1]	--	--	70%	--	3%
877NC1	Neruokpuk	--	Precambrian	36	16	--	42	2	4
927NC1	Neruokpuk	--	Precambrian	24	--	23	43	--	10
1173NC1	unnamed	C-41895	L. Silurian	26	--	--	59	--	15
1178NC1	Neruokpuk	--	Precambrian	15	--	27	40	--	18
1193NC1	Neruokpuk	--	Precambrian	19	--	31	45	--	5

[1] Percentages are volumetric

Figure 97. 3. Northwest flank of Cache Creek Uplift showing unconformities 1, between unnamed shales and limestones of the Permian (Ps) and Jurassic (Jbc) Systems; 2, between the Jurassic Bug Creek Formation (Jbc) and unnamed Upper Jurassic sandstones (Jpo); and 3, between the Upper Cretaceous Cuesta Creek Member (Kck) of the Tent Island Formation and deformed Jurassic and Cretaceous rocks. Two GSC fossil localities with ages are also shown. View is to the north-northwest. NAPL Oblique Photograph T-5 22R.

In 1973, the management of Inexco Oil Company (Canada) Ltd. kindly revealed to the writer the presence of an occurrence[1] of oolitic magnetite in the northwest quarter of the Porcupine River map-area (116J, K) at longitude 140°16' West, latitude 66°31.2' North. It may be located on NAPL vertical aerial photograph A13231-131 at the following Cartesian co-ordinates: X = +3.02 cm, Y = +3.30 cm. The occurrence has been staked recently. It is interesting economically because it has the potential of recurring at the same stratigraphic level elsewhere in the Kandik Thrust Belt.

The iron occurs as a bed estimated to be 30 m (100 ft.) thick and can be traced in outcrop for approximately 150 m (500 ft.) along the strike at a stratigraphic level about 150 m (500 ft.) above the base of the Jurassic and Lower Cretaceous Kingak Formation. It is dark grey to black, rusty brown weathering, massive, dense, oolitic magnetite that weathers into angular chunks and blocks. In thin section, the oolites are packed densely and are commonly flattened, with their long dimensions as much as 1 mm, but averaging about 0.5 mm. They are replaced locally by hematite and in many places are replaced by goethite.

As part of a continuing study of the structural and stratigraphic make-up of the so-called Aklavik Arch (Norris, 1974, p. 30-32), the writer re-examined some Jurassic and Lower Cretaceous outcrops between Big Fish River and Little Fish (Cache) Creek on the northwest flank of Cache Creek Uplift (*see* Figs. 97.1, 97.2, geological map and structure section). These helicopter stops in conjunction with observations made in 1972 have helped considerably to resolve some of the problems of the area and have contributed to the completion of geological maps of the Blow River (117A) and Aklavik (107B) areas from which Figure 97.1 was prepared. The minor discrepancies in fit where the two maps were spliced together is a compromise resulting from the joining of topography and geology across the splice.

Rocks exposed within, on the flanks and down the plunge of Cache Creek Uplift range in age from Late Carboniferous [Bashkirian-Moscovian (Bamber and Waterhouse, 1971, Fig. 10)] to Early Tertiary. Although partitioned by unconformities, they do not record the influence of the uplift on sedimentation patterns until Jurassic and Cretaceous times. Indeed, the uplift itself as a northeast-trending, broadly anticlinal feature cut by nearly vertical, longitudinal faults, is a mid-Tertiary phenomenon.

The lower Paleozoic graptolitic facies and the Permo-Carboniferous clastic and carbonate strata appear to thin systematically eastward independently of the uplift (*see* Fig. 97.2) whereas noticeable thickness changes and truncation can be discerned in Jurassic and younger rocks. The sandstone and siltstone of the Bug Creek Formation, for example, are observed to thicken westward into the uplift by a factor of four. The formation is approximately 150 m (500 ft.) thick in the general Aklavik Range area (Jeletzky, 1975, Fig. 8) whereas it is approximately 610 m (2000 ft.) on the west flank of the uplift (Mountjoy and Procter, 1970, Sec. 117A5, Units 3-8, inclusive). On the other hand, the interval from the top of the Bug Creek Formation to the base of the "upper shale siltstone division" remains roughly the same thickness between these two points. It is 610 m (2000 ft.) in the Aklavik Range area (Jeletzky, ibid.) and about 550 m (1800 ft.) on the west flank of the uplift (Mountjoy and Procter, 1970, Units 9-13, inclusive). Within this interval, the mature pale orange, quartz, chert sandstone of the unnamed Upper Jurassic Formation (Jpo) can be seen to truncate the underlying Bug Creek Formation southward onto the crest of the structurally higher part of the uplift (Fig. 97.3) denoting at least local differential erosion within the sequence.

Northeast-trending folding and faulting of the lower Upper Cretaceous and older rocks, followed by uplift and erosion preceded onlap of the Upper Cretaceous Cuesta Creek Member of the Tent Island Formation (Young, 1974b) and resulted in the angular unconformity and truncation shown in Figures 97.3 and 97.4.

In addition, an isopach map (Young, 1975, Fig. 8) of the Upper Cretaceous Moose Channel Formation indicates noticeable thickening of this regressive and transgressive formation over Cache Creek Uplift. In the direction of the plunge, moreover, the thickness maximum would appear to coincide with the Bouguer gravity minimum and depocentre postulated by the writer (Norris, 1974, p. 38 and Fig. 1) in Mackenzie Bay opposite the mouth of Big Fish River.

As an integral component of the Aklavik Arch Complex (Yorath and Norris, 1975, p. 596), Cache Creek Uplift was demonstratively active beginning in the Jurassic. It was a negative feature in the Early to Middle Jurassic and again in the Late Cretaceous; it was positive in the early Late Cretaceous and from the mid-Tertiary to the Holocene.

In the event that the style of deformation in Romanzof Uplift may be similar to that at equivalent structural levels in southern Beaufort Shelf, the writer re-examined a number of faults and folds in the general Firth River-Crow River area. Among those features was Crow Fault (Fig. 97.5) in the core of the uplift (long. 139°38' West, lat. 69°02' North). Because of the montonous nature of the stratigraphic succession within any of the bulk units now identified in the Neruokpuk Formation, it was difficult to recognize faults and also to map their changing stratigraphic separations along strike. Thus, in situations where rocks of contrasting facies become involved, as on Crow Fault in the upper reaches of Crow River, a major northeast-dipping structure was readily identifiable. The dip and direction of relative transport on Crow Fault is counter to that of the thrust belt on the southwest flank of the uplift and an element of symmetry is introduced to the deformation. Although considerably outnumbered in the uplift, similar northeast-dipping faults may occur in the offshore, and

[1] The writer thanks Amoco Canada Petroleum Company Ltd., W.R. Grace and Company Ltd., Husky Oil Operations Ltd. and Inexco Oil Company Ltd., for permission to publish the precise geographic position of this occurrence.

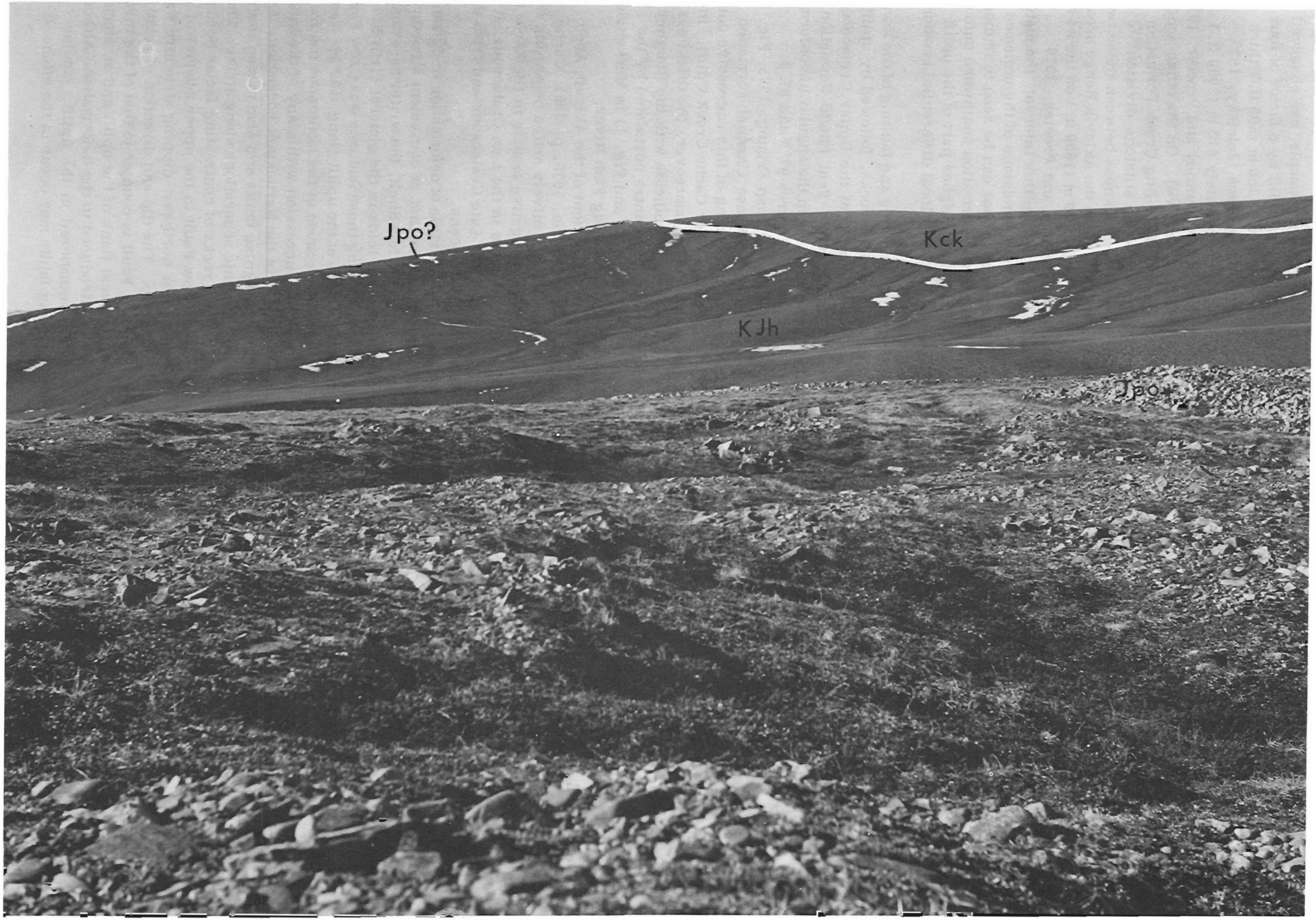

Figure 97. 4. Angular unconformity between the Upper Cretaceous Cuesta Creek Member (Kck) of the Tent Island Formation and folded, unnamed Upper Jurassic sandstones (Jpo) and Husky Formation (KJh) on the northwest flank of Cache Creek Uplift. View is to the north. Arrow in Figure 97. 3 identified position and direction of this photograph. GSC Photo 199126.

Figure 97.5. Crow Fault (CF), a major, northeast-dipping, high-angle, reverse fault in the core of Romanzof Uplift, placing Precambrian Neruokpuk Formation (Pn2) on Lower Carboniferous Kayak Formation (Cky) and Lisburne Group (C1) in the upper reaches of Crow River. View is to the northwest. GSC Photo 199125.

Figure 97.6. The Kaltag-Porcupine Fault (K-PF) on Porcupine River approximately nine miles (14 km) upstream from the village of Old Crow, Y.T. Unnamed dolomites and limestones (ЄDb) of early Paleozoic age are in juxtaposition across the fault with unnamed sandstones (Kdr) dated as Late Cretaceous (Santonian). View is almost due west. GSC Photo 199124.

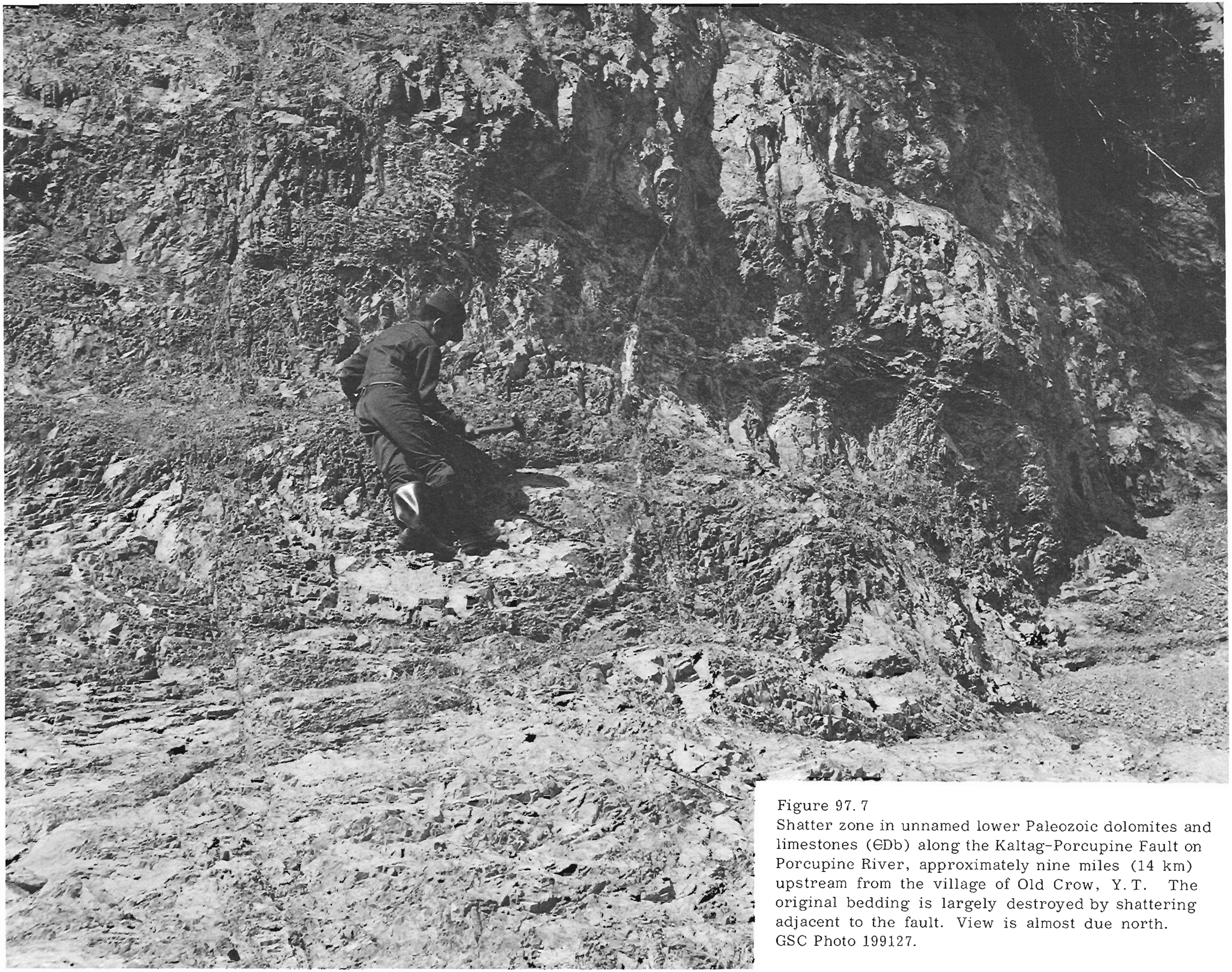

Figure 97.7
Shatter zone in unnamed lower Paleozoic dolomites and limestones (ЄDb) along the Kaltag-Porcupine Fault on Porcupine River, approximately nine miles (14 km) upstream from the village of Old Crow, Y.T. The original bedding is largely destroyed by shattering adjacent to the fault. View is almost due north. GSC Photo 199127.

their possible existence there may assist in the interpretation of deep seismic and other geophysical studies of that region.

The Kaltag-Porcupine Fault was studied further where it crosses the Porcupine River a few miles above the village of Old Crow, Yukon Territory, in order that its position and separation could be fixed at this readily accessible locality (long. 139°38' West, lat. 67°34.9' North). There, silty, light grey, finely crystalline dolomite and limestone (ƐDb) are in juxtaposition with limonitic, pale brown, quartz, chert, medium grained sandstone (Kdr) (Fig. 97.6). Both successions are inclined gently westward and the sandstone seems to dip beneath the carbonates. Although the dolomite and limestone are distinctly bedded away from the contact, they are totally shattered adjacent to it and bedding is obscure (Fig. 97.7). The fault contact is not exposed although its trace is confined to a covered interval a few hundred feet in width.

Neither the carbonates nor the sandstones appeared to be fossiliferous adjacent to the fault although fossils occur a mile or two upstream and downstream. Downstream and a few hundred yards north of Porcupine River (GSC loc. C-51521; long. 139°41' West, lat. 67°34.3' North), a sample of grey dolomite was found to contain conodonts of early Paleozoic age, tentatively between Middle Silurian and Early Devonian (T.T. Uyeno, pers. comm., 1975). Because of the westward dip of the carbonates, the horizon sampled probably lies stratigraphically higher than those exposed along the river and adjacent to the fault. A younger age there, within the lower Paleozoic carbonate platform, is to be expected and the tentative age assignment by Uyeno is consistent with the regional mapping. Upstream and along the north bank of the river (GSC loc. C-11870; long. 139°31.7' West, lat. 67°34.7' North), the writer collected a well-preserved pelecypod fauna dated by J.A. Jeletzky as Late Cretaceous (Santonian). The time-stratigraphic separation across the fault is, therefore, fixed between lower Paleozoic (Middle Silurian? to Lower Devonian?) and mid-Upper Cretaceous. The rock-stratigraphic separation, on the other hand, may not be large because of the magnitude of stratigraphic omissions on some of the unconformities within the succession on the northwest flank of the Aklavik Arch Complex (Norris, 1974, Table 1).

References

Bamber, E.W. and Waterhouse, J.B.
1971: Carboniferous and Permian stratigraphy and paleontology, northern Yukon Territory, Canada; Bull. Can. Pet. Geol., v. 19, no. 1, p. 29-250.

Dutro, J.T., Jr., Brosgé, W.P., and Reiser, H.N.
1972: Significance of recently discovered Cambrian fossils and reinterpretation of the Neruokpuk Formation, northeastern Alaska; Bull. Am. Assoc. Pet. Geol., v. 56, no. 4, p. 808-815.

Jeletzky, J.A.
1971: Stratigraphy, facies and paleogeography of Mesozoic rocks of northern and west central Yukon; *in* Report of Activities, Part A, Geol. Surv. Can., Paper 71-A, p. 203-221.

1975: Jurassic and Lower Cretaceous paleogeography and depositional tectonics of Porcupine Plateau, adjacent areas of northern Yukon and those of Mackenzie District; Geol. Surv. Can., Paper 74-16.

Mountjoy, E.W. and Procter, R.M.
1970: Eleven field descriptions of some Jurassic and Cretaceous rocks in Arctic Plateau and Arctic Coastal Plain; Geol. Surv. Can., Open File 16.

Norford, B.S.
1964: Reconnaissance of the Ordovician and Silurian rocks of northern Yukon Territory; Geol. Surv. Can., Paper 63-39.

Norris, D.K.
1963: Operation Porcupine *in* Summary of research: field, 1962; Geol. Surv. Can., Paper 63 1, p. 17-19.

1964: Yukon Territory *in* Age determinations and geological studies; Geol. Surv. Can., Paper 64-17, Pt. 1, p. 22.

1974: Structural geometry and geological history of the northern Canadian Cordillera; Can. Soc. Explor. Geophys., Proc. 1973 National Convention, p. 18-45.

1975a: Geological maps of parts of Yukon and Northwest Territories; Geol. Surv. Can., Open File 279.

1975b: Geological maps of part of Northwest Territories; Geol. Surv. Can., Open File 302.

Yorath, C.J. and Norris, D.K.
1975: The tectonic development of the southern Beaufort Sea and its relationship to the origin of the Arctic Ocean Basin in Canada's continental margins and offshore petroleum exploration; Can. Soc. Pet. Geol., Mem. 4, p. 589-611.

Young, F.G.
1974a: Mesozoic epicontinental, flyschoid and molassoid depositional phases of Yukon's North Slope *in* Geology of the Canadian Arctic; Can. Soc. Pet. Geol. and Geol. Assoc. Can., p. 181-202.

1974b: Upper Cretaceous stratigraphy, Yukon Coastal Plain and northwestern Mackenzie Delta; Geol. Surv. Can., Open File 193.

98. LOWER PALEOZOIC STRATIGRAPHY: SOUTHERN MACKENZIE MOUNTAINS, DISTRICT OF MACKENZIE

Project 750017

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Introduction

Stratigraphic studies of lower Paleozoic units exposed in the southern Mackenzie Mountains (map sheets 95F and 95G; Fig. 98.1) were conducted for a period of six weeks in June and July, 1975. During this time, 14 sections that ranged in thickness from 240 m to 2161 m were measured (Fig. 98.1). Formations examined included the Delorme, Camsell, Sombre, Arnica, Manetoe, Headless, Nahanni and Funeral formations. Data accumulated during the course of this project will provide the basis for refinements and, possibly, revisions of the existing stratigraphic nomenclature of this lower Paleozoic sequence. The most recent published stratigraphic framework for the units discussed in this report is contained in Douglas and Norris (1960, 1961, in press a, b). A few preliminary observations, based solely on field work concerning stratigraphy and sedimentology, are described in this report.

Delorme and Camsell formations

Partial sections indicate that the Silurian and Lower Devonian Delorme Formation in the northern part of the map-area (Fig. 98.2) is composed of thin-bedded, dark brownish grey, fossiliferous and argillaceous lime wackestone and medium-bedded, clean, light grey, lime grainstones. Farther south, these wackestones and grainstones are dolomitized (Fig. 98.2).

Dolomite of the Lower Devonian Camsell Formation appears to overlie the Delorme Formation conformably. This dolomite, as has been noted by Douglas and Norris (1961), is colour banded. It consists of repetitive intervals of dark, thick-bedded, brownish grey biospromal dolomite; thick bedded, medium grey, subtidal dolomite; and silty, light greyish yellow, thin bedded dolomite. These repetitive intervals form subtidal-intertidal cycles that are generally less than 10 m thick and commonly less than a few metres thick.

The southward thinning of the Camsell Formation is accompanied by a decrease in the number and thickness of subtidal phases. Dark brownish grey biostromal dolomite, which forms discrete small bioherms in Sections 1 and 2, is not present in Sections 9 and 11. These observations indicate that, during deposition of the Camsell Formation, land lay toward the south along the line of section presented in Figure 98.2.

Sombre and Arnica formations

The Lower Devonian Sombre Formation appears to overlie the Camsell Formation with a well-defined but gradational contact. Light to medium grey, thick- to very thick bedded dolomite forms most of the Sombre but thin bedded and laminated intertidal intervals are

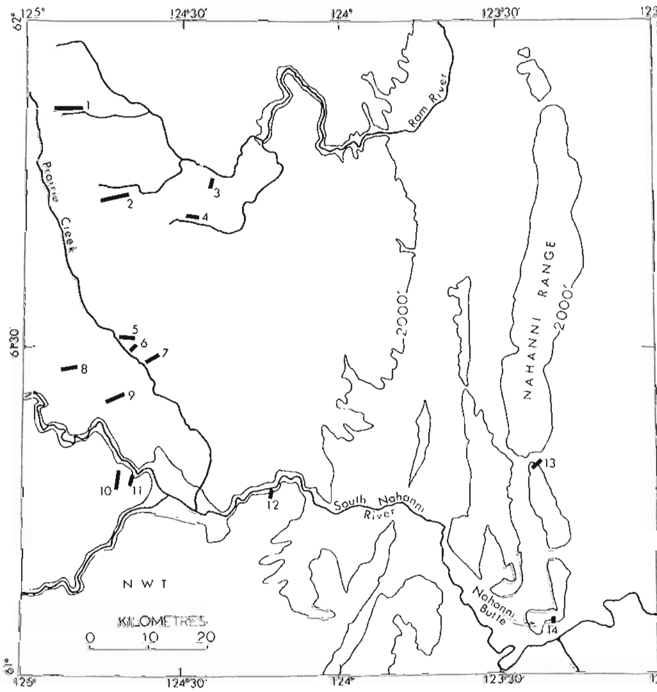

Figure 98.1. Map showing locations of stratigraphic sections numbered 1 to 14.

common near its contact with the Camsell Formation. Toward the upper part of the formation, some of the thick-bedded subtidal intervals contain good moldic porosity from leached biogenic material such as corals and amphiporids. Larger biogenic constituents are preserved only rarely.

Previous workers (Douglas and Norris, 1961; Gabrielse *et al.*, 1973) have suggested that the Arnica Formation unconformably overlies the Sombre. Field mapping during this project (D.G. Cook, pers. comm.) indicates that a facies contact between these units may be more likely in this area. Also, stratigraphic work shows that, in some places, Sombre-type lithologies are interbedded with Arnica-type lithologies, indicating an intertonguing facies relationship between the Sombre and Arnica formations (Fig. 98.2).

Medium- to very thick bedded, dark greyish brown, fetid dolomite forms most of the Arnica Formation. At the type section (Sec. 12, Fig. 2) in the First Canyon of South Nahanni River (Douglas and Norris, 1961), the Arnica Formation is prominently colour banded with beds of light grey, intertidal dolomite containing fenestrate fabric and shallow-water, dolomitized pisolitic grainstones scattered throughout a sequence of thick-bedded, dark brown dolomite. Farther west (e.g. Sec. 11), colour banding in the Arnica is less

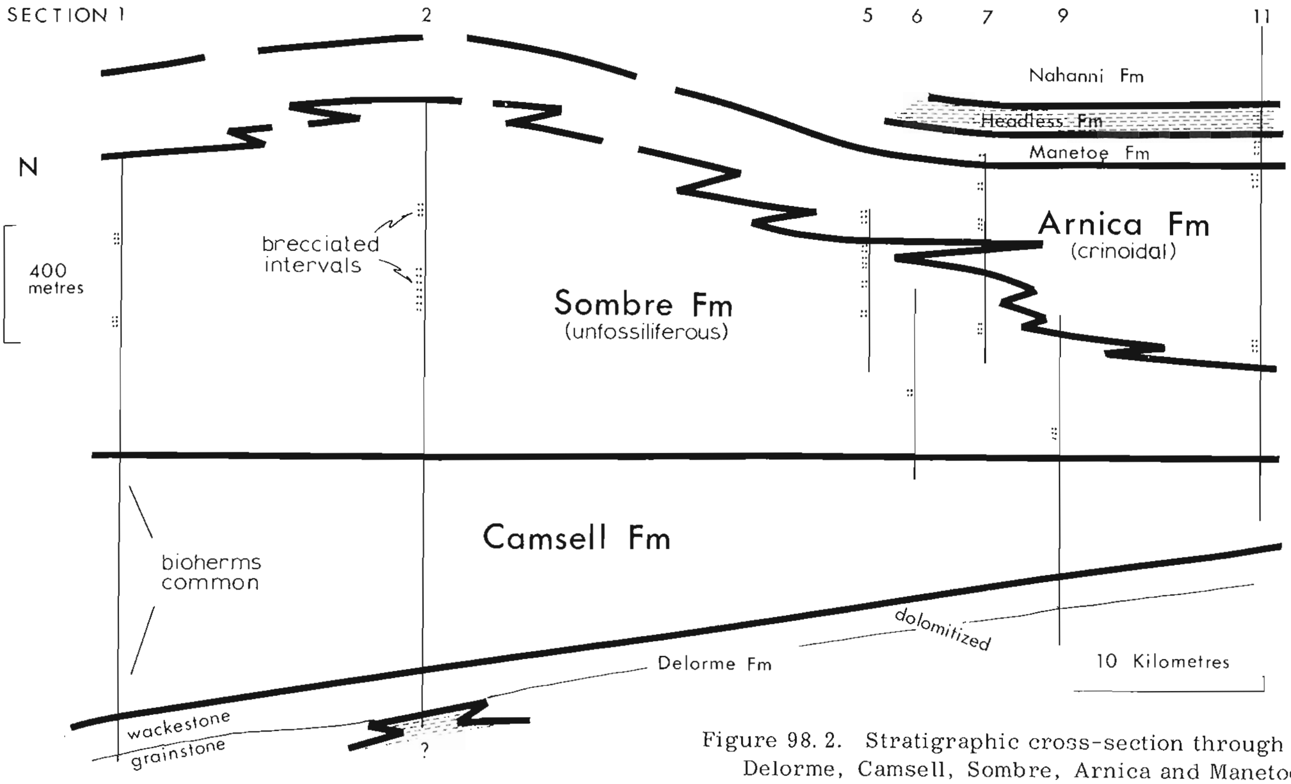

Figure 98. 2. Stratigraphic cross-section through the Delorme, Camsell, Sombre, Arnica and Manetoe formations.

obvious and the formation tends to be abundantly crinoidal (Fig. 98. 2). An interval, 174 m thick, of slope-deposited limestones in Section 8 (Fig. 98. 1) contains many fossiliferous carbonate debris flows that may be correlative with part of the Arnica-Sombre sequence farther east.

Both the Sombre and Arnica formations are extensively brecciated. The rubble breccias of the Sombre Formation tend to be stratiform and are locally in excess of 50 m thick as in Section 2 (Fig. 98. 2). The angular to subrounded, grey dolomite fragments of the Sombre breccias are very poorly sorted and are encased in dolomitized grey lime mud. Individual fragments greater than 3 m long were observed. Rubble and mosaic breccias near the top of the Sombre tend to contain some interfragment, very coarsely crystalline, white dolomite cement.

Crackle[1], mosaic[2] and rubble[3] breccias are common in the Arnica Formation at all stratigraphic levels. They are almost entirely cemented by white, coarsely crystalline dolomite. Arnica crackle and mosaic breccias tend to be stratiform, but rubble breccia bodies tend to be oriented vertically across the bedding. If the Arnica and Sombre breccias are cogenetic, the high proportion of lime mud matrix in Sombre breccias may be explained as the accumulation of silt-size carbonate debris in the lower galleries of an extensive Arnica-Sombre cavern system.

In the western part of the area (Fig. 98. 1), this late-stage white dolomite appears to be stratiform and is a well-defined, mappable horizon coincident with the Manetoe Formation (D. G. Cook, pers. comm.). Farther east, however, this late phase white dolomite is not stratiform. At Sections 13 and 14 at Grainger River and Nahanni Butte, breccia bodies of white dolomite extend more than 100 m above the Manetoe Formation into the medium bedded, fossiliferous, dark grey lime wackestones of the overlying Nahanni Formation. These vertical breccia bodies are surrounded

Manetoe, Headless and Nahanni formations

The lithologies, distribution and thicknesses of the Middle Devonian Manetoe, Headless, and Nahanni formations were found to correspond with the descriptions given by Douglas and Norris (1960, 1961). Dolomitization in the Manetoe Formation occurred in at least two distinct stages. In many places, the Manetoe Formation consists largely of very thick bedded to massive, dark grey, finely crystalline dolomite with a commonly well-preserved fauna of corals and stromatoporoids, differentiating it from the underlying relatively unfossiliferous Arnica Formation. In addition, a later event a solution-collapse occurred, forming a rubble breccia infilled with very coarsely crystalline, white dolomite.

[1] Breccia fragments almost in place.

[2] Breccia fragments slightly displaced but original fragment positions are readily discernible.

[3] Original position of fragments cannot be discerned.

with limestone beds that have recrystallized to medium crystalline, light brown dolomite. This halo of dolomite recrystallization extends tens of metres laterally from the breccia bodies into Nahanni limestone beds and creates an interesting problem in stratigraphic nomenclature for this part of the sequence, particularly if it proves to be of regional occurrence in the subsurface farther east.

Shale of the Headless Formation, which is more than 100 m thick in the western part of the area (Fig. 98.2), may have confined the circulation of subsurface fluids beneath the Nahanni Formation. In the eastern part of the area, the Headless shale is very thin or absent, as at Sections 13 and 14, and diagenetic fluids under enough hydrostatic head may have circulated upward from the Manetoe Formation into the overlying Nahanni limestone.

References

Douglas, R.J.W. and Norris, D.K.

1960: Virginia Falls and Sibbeston Lake map-areas, Northwest Territories; Geol. Surv. Can., Paper 60-19.

1961: Camsell Bend and Root River map-areas, District of Mackenzie, Northwest Territories; Geol. Surv. Can., Paper 61-13.

Geology, Virginia Falls, District of Mackenzie; Geol. Surv. Can., Map 1378A. (in press a)

Geology, Sibbeston Lake, District of Mackenzie; Geol. Surv. Can., Map 1377A. (in press b)

Gabrielse, H., Blusson, S.L., and Roddick, J.A.

1973: Geology of Flat River, Glacier Lake, and Wrigley Lake map-areas, District of Mackenzie and Yukon Territory; Geol. Surv. Can., Mem. 366.

99. SOME ASPECTS OF STRATIGRAPHY AND STRUCTURE IN THE HALFWAY RIVER MAP-AREA (94B), BRITISH COLUMBIA

Project 750019

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Introduction

During the summer of 1975, three months were spent mapping the northwestern quadrant of Halfway River (94B) map-area and the southwestern corner of Trutch (94G) map-area, at a scale of 1: 50 000 (Fig. 99. 1). Mapping in 1976 should complete the southwestern quadrant of the Halfway River sheet along with selected areas in Trutch and Ware (94F) map-areas. This project will provide a more thorough understanding of this segment of the northern Rocky Mountains including the regional geological setting of lead-zinc mineralization in the area.

This year's field work shows that lower and middle Paleozoic rocks can be subdivided further into mappable lithostratigraphic units; that regional Silurian and Devonian facies changes can be located in the area; and that the style of deformation is dominated by folds with few attendant thrust faults.

General Geology

The western half of the Halfway River sheet straddles the profound stratigraphic and structural changes that occur between the essentially flat-lying, undeformed platformal rocks beneath the plains region, and the thicker, strongly deformed miogeoclinal rocks that comprise the Rocky Mountains belt.

The area was mapped systematically by Irish (1970) during the period 1958-1962. Since that time, a considerable amount has been learned about regional stratigraphic and structural relationships in the northern Rocky Mountains; Irish's work now serves as an excellent base from which a more detailed geological evaluation of the western part of the Halfway River sheet can be made.

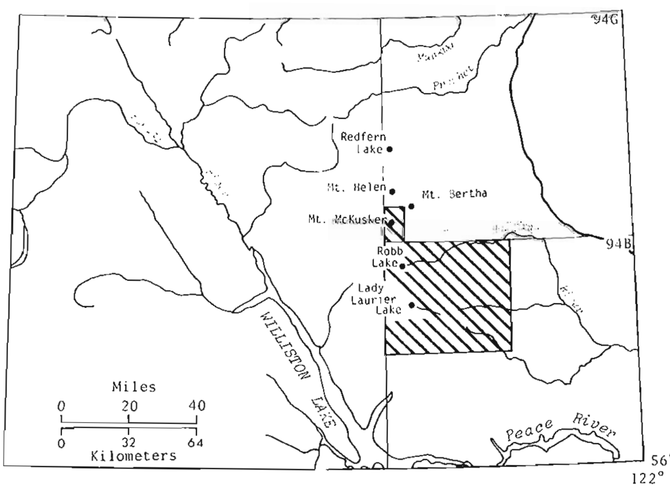


Figure 99. 1. Location of area mapped during 1975 field season.

Three main physiographic regions are spanned by the area mapped; from east to west they are the Plains, Foothills and Rocky Mountains. The Plains region consists of mainly flat-lying terrain underlain by gently folded rocks of Cretaceous age. The Foothills rise abruptly from the Plains as rolling mountains with a vertical relief of over 1000 m (3000 ft.) that are underlain by rocks of Mississippian through Cretaceous age. The western half of the map-area contains the eastern part of the Rocky Mountains belt; it is a rugged mountainous terrain having a vertical relief of over 1500 m (4500 ft.) that is underlain by rocks of middle and lower Paleozoic age. As the terrain becomes progressively more rugged from east to west, the age of the exposed rocks and intensity of deformation increase westward reflecting a structural style dominated by folding (Fig. 99. 2).

Facies changes are an important part of the geology in this region. Major Mississippian and Triassic facies transitions occur in the Foothills belt; major Ordovician, Silurian and Devonian facies changes occur in the Rocky Mountains belt (Fig. 99. 3). The facies changes are not only part of a complex stratigraphic evolution along a relatively narrow platform margin but also have helped control the nature of structural development in the area.

Upper Paleozoic and Mesozoic Stratigraphic Units

The stratigraphy of upper Paleozoic and Mesozoic rocks in the area has been studied and reported on previously: Bamber *et al.* (1968) on rocks of Mississippian through Permian ages; Gibson (1971) and Pelletier (1964) on rocks of Triassic age; and Stott (1967) on rocks of Jurassic and Cretaceous age. The stratigraphic units described by these authors were accepted and mapped, without further study, across the region.

Lower and Middle Paleozoic Stratigraphic Units

Ordovician, Silurian and Devonian rocks in the map-area were subdivided into seven rock units that are additional to Silurian and Devonian platform carbonate units known from the area. Rocks of Ordovician age comprise three mappable units including a thick platform carbonate that may correlate with the Skoki Formation in the southern Rocky Mountains. The Silurian Nonda Formation was mapped along with two other subdivisions that are deeper water equivalents

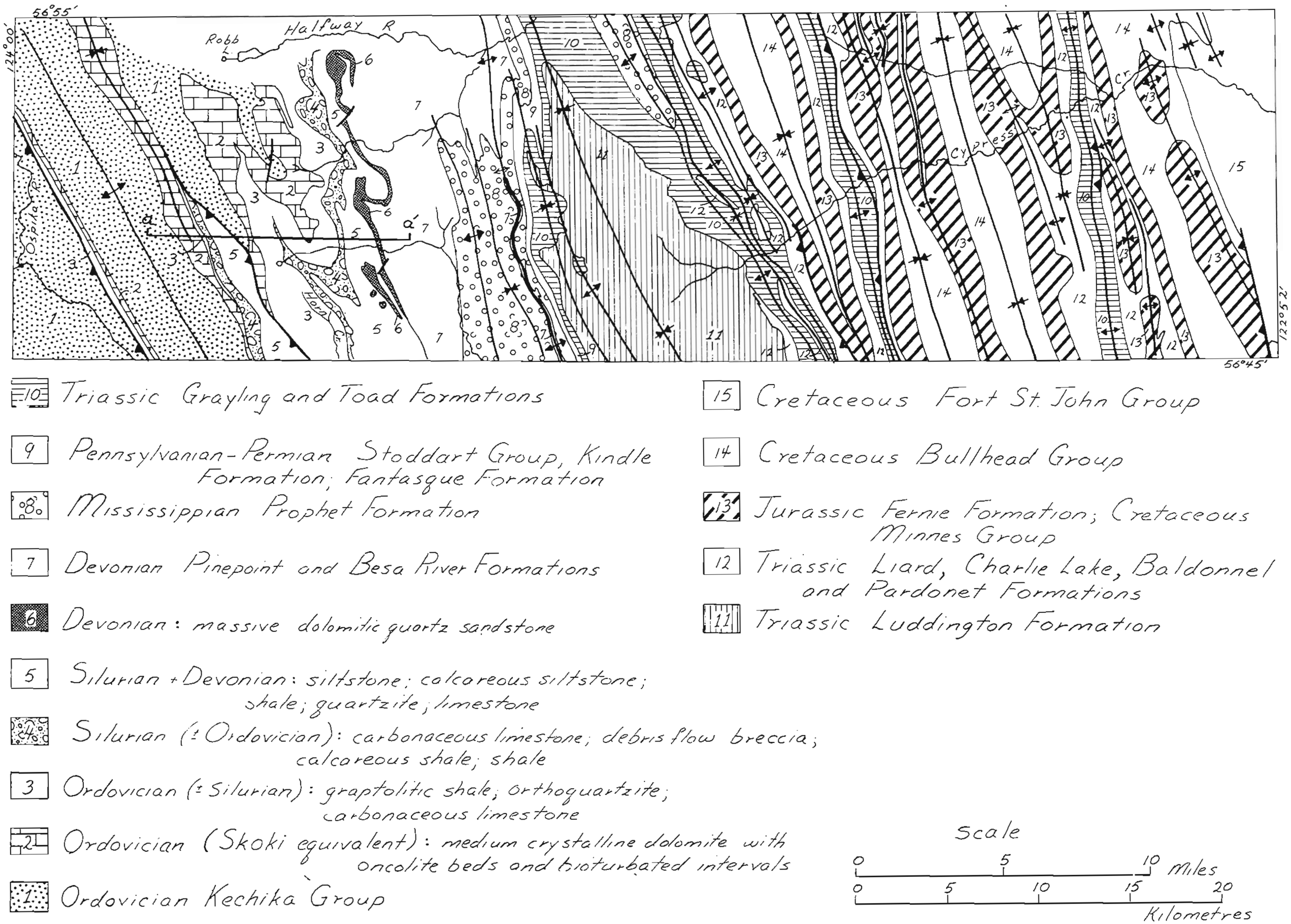


Figure 99. 2. Strip map across northern part of Halfway River map-area between Robb Lake and the headwaters of Horn Creek.

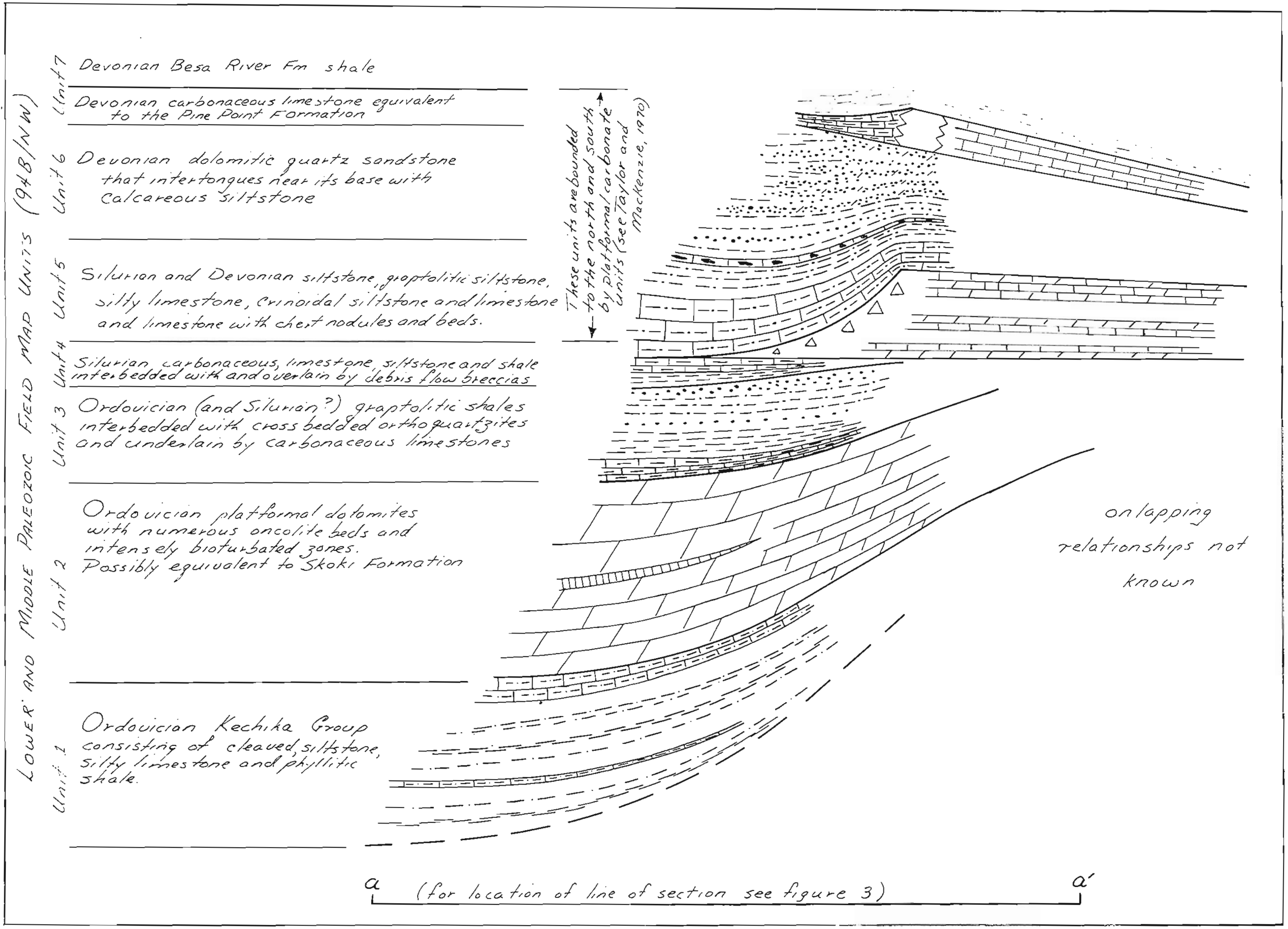


Figure 99. 3. Diagrammatic east-west cross-section approximately 10 km (6 mi.) south of Robb Lake showing stratigraphic relations between lower and middle Paleozoic map-units.

Figure 99.4

Facies change in Silurian Nonda Formation.

of the Nonda. Two, previously unmapped Devonian units were recognized in addition to the Muncho McConnell, Stone and Pine Point formations (Taylor and MacKenzie, 1970; Taylor *et al.*, 1975). Figure 99.2 is a generalized geological map showing the distribution of the previously unmapped rock units between Halfway River and the headwaters of Horn Creek; Figure 99.3 is a diagrammatic cross-section showing stratigraphic relations as they are interpreted to exist across this part of the region.

Ordovician (and Lower Silurian?) map-units

Rocks of Ordovician age can be subdivided into three mappable units (Fig. 99.2): an upper graptolitic shale-orthoquartzite unit; a massive platformal carbonate; and a lower well-cleaved silty limestone and siltstone unit belonging to the Kechika Group.

The graptolitic shale is interbedded with and overlain by tongues of orthoquartzite. This unit changes in character along strike: in some areas, the quartzite tongues can be mapped separately from the shale; in Trutch map-area, the shale is replaced by reddish weathering dolomite and silty dolomite.

The massive carbonate unit is up to 800 m (2500 ft.) thick, consists of thick bedded, medium crystalline dolomite characterized by two distinctive features: numerous oncolite beds, and zones of intense bioturbation. This map-unit may be correlative with the Skoki Formation. South of Lady Laurier Lake, a distinctive interval of calc-alkaline volcanic flows and interbedded reworked volcanic sedimentary rocks (sandstone and conglomerate) up to 200 m (600 ft.) thick occurs about midway in the unit. The volcanic unit extends at least as far south as the Mount Burden area (pers. comm., Brinex geologists).

The oldest Ordovician unit consists of thick, pervasively cleaved, calcareous siltstone, shale and silty limestone of undetermined thickness that belongs to the Kechika Group. Compositional changes are subtle in most areas and difficult to follow; this map-unit was not further subdivided.

Silurian map-units

The Silurian Nonda Formation (Norford, 1966) undergoes an impressive facies change (Fig. 99.4) from platformal and marginal bank facies to slope and distal slope deposits of debris flow breccias, and thin-bedded, chert-bearing carbonaceous limestone and calcareous shale. This facies change can be followed from north to south across the area mapped. Chaotic angular debris flow breccias overlie the thin-bedded carbonaceous limestone and were mapped separately from it. No sections have been measured through this part of the succession as yet.

Upper Silurian and Devonian map-units

Important facies changes occur along strike within platformal Devonian carbonate formations between Muskwa River in Trutch map-area to the north and Lapierre Creek in the Halfway River sheet. As Taylor and MacKenzie (1970, Fig. 9) indicated in their study of the Devonian, the fourfold subdivision of the Devonian carbonate succession into the Muncho-McConnell, Wokkpash, Stone and Dunedin formations becomes less distinct south of the Muskwa River. The Wokkpash sandstone is difficult to distinguish as a separate formation, the Dunedin Formation changes facies into the Pine Point Formation (Taylor *et al.*, 1975), and distinction between the Muncho-McConnell and the Stone formations is increasingly difficult to make southward. Northeast of Robb Lake, the succession is about 700 m (2000 ft.) thick; the Wokkpash sandstone cannot be identified; distinction between a Muncho-McConnell Formation and a Stone Formation could not be made by this author; and the Dunedin Formation has changed facies to a thin [70 m (≈200 ft.)] Pine Point Formation.

Between the Halfway River and Lady Laurier Lake, a more profound facies change occurs. Rocks younger than the Silurian Nonda Formation consist of siltstone and silty limestone of undetermined thickness [at least 350 m (1000 ft.)] capped by at least 350 m (1000 ft.) of slightly dolomitic quartz sandstone that thins to the south and west. Platform carbonates typical of the Devonian to the north and south do not occur. To what extent this previously unrecognized Devonian facies represents a more allochthonous block transported from the west and emplaced structurally on top of platform rocks has yet to be determined; however, large-scale lateral translation is not supported by regional map relationships. Instead, this profound change in lithologies is interpreted as a west-northwest-east-southeast-trending facies change that is oblique to the structural grain of the region.

South of Lady Laurier Lake area, platform Devonian rocks reappear. At Lapierre Creek, the section is over 1000 m (3000 ft.) thick and consists of a thick-bedded cliff forming Muncho-McConnell Formation of medium crystalline dolomite with sandy beds at the base; a distinctly bedded Stone Formation of sandy dolomite and dolomitic sandstone; and dark grey, fossiliferous, crystalline dolomite and limestone of the Pine Point Formation.

The nature of the transition between the platform carbonates that occur north and south of the map-area, and the siltstone-quartz sandstone facies between Halfway River and Lady Laurier Lake is not adequately known. At the northern end, on Halfway River, the platform dolomites plunge southward beneath an allochthonous block containing faulted tight folds of the siltstone-quartz sandstone facies; at the southern end, east of Lady Laurier Lake, a large anticline of Muncho-McConnell dolomite plunges northward and may be overlain stratigraphically by, and intertongue with, the siltstone quartz sandstone facies. Along the eastern slopes of the Rocky Mountains belt, the siltstone-quartz sandstone units are succeeded by carbonaceous limestone thought to be equivalent to the Pine Point Formation, and silty shale of the Besa River Formation. There is no evidence of a major thrust fault on which the siltstone-quartz sandstone facies could have been transported from farther west; instead, these units are deformed into large overturned anticlines with steeply dipping lower limbs that show no evidence of truncation (Fig. 99.8). At present, it is felt that the new Devonian facies are not significantly out of place relative to their platformal counterparts and it is probable that the siltstone-quartz sandstone facies extend farther to the east in the subsurface.

Structure

Figure 99.3, a strip map across the map-area, demonstrates the essential difference between the structural style of this area and that documented for the southern Canadian Rockies (e.g. Price and Mountjoy, 1970; Bally *et al.*, 1966). The rocks of the Halfway River map-area have not been imbricated extensively by flat thrust faults; instead, large folds that become more irregular and appressed toward the west control the distribution of rock types. Thrust faults occur, in most cases, along the limbs of closed folds; they cross-cut bedding at high angles in some places (Fig. 99.5), show little evidence of large horizontal displacements [an exception to this is a thrust north of Robb Lake that has at least 5 km (≈3 miles) of horizontal translation], and die out quickly and abruptly along strike.

Adjacent to the plains region, folds tend to be open and "box-like" (Fig. 99.6) in cross-section; wave-lengths decrease and amplitudes increase westward as evidenced in the Mississippian Prophet Formation exposed along the Halfway River (Fig. 99.7). Along the eastern flank of the Main Ranges, large, nappe-like overturned anticlines, involve middle Paleozoic rocks (Fig. 99.8); this style of folding persists to the western margin of the map-area. There is little doubt that folding and not thrust faulting was dominant throughout deformation of Paleozoic and Mesozoic rocks in this part of the Rocky Mountains.

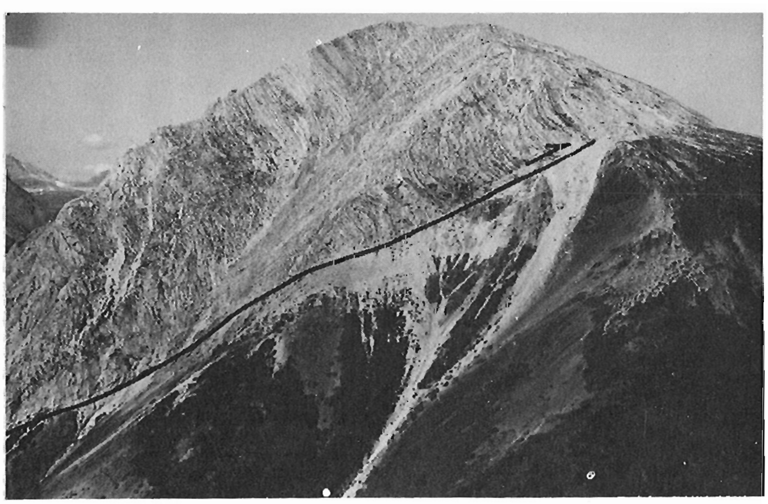


Figure 99.5

Middle Devonian Stone Formation in fault contact with underlying Upper Devonian Besa River shales along the eastern margin of the Rocky Mountains belt. Fault cuts at right angle to bedding in hanging wall.

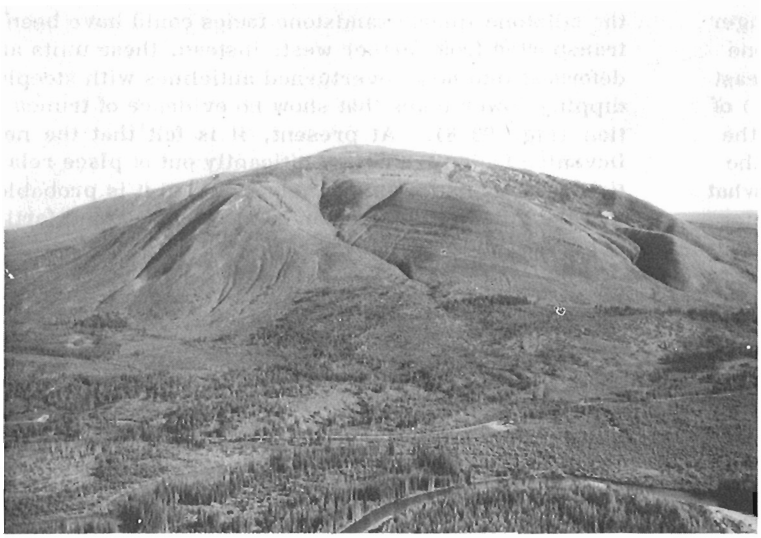
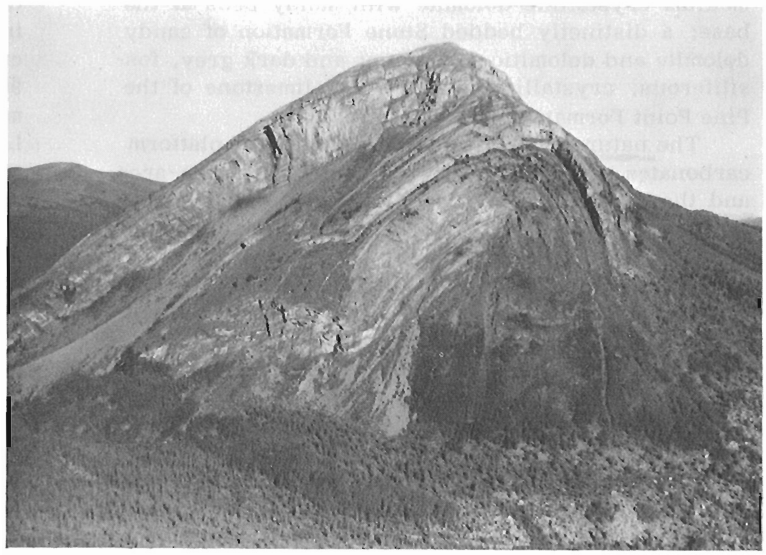


Figure 99. 6

Low amplitude 'box' anticline of Cretaceous Gething Formation along the eastern margin of the Foothills belt (viewed from the south).

Figure 99. 7

Large amplitude closed anticline of Mississippian Prophet Formation along the western margin of the Foothills belt (viewed from the south).



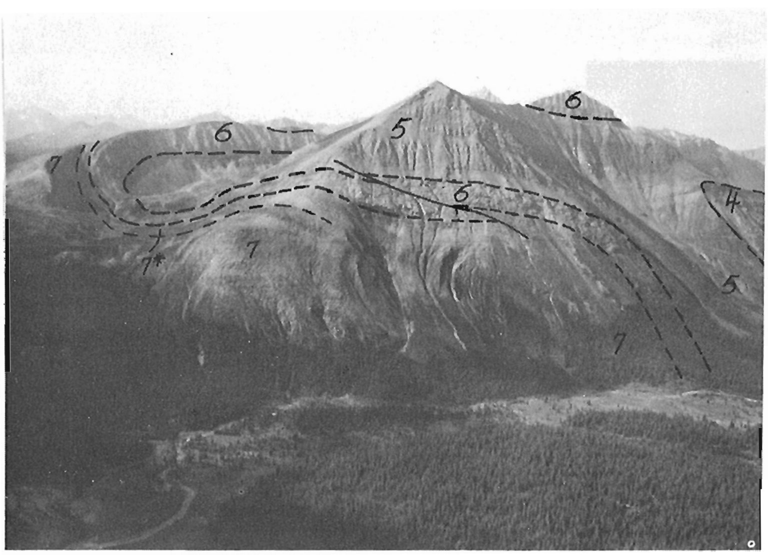


Figure 99. 8

Recumbent isoclinal anticline on the eastern margin of the Rocky Mountains belt (viewed from the north-west). Numbers refer to rock units described in Figure 99. 3 (7* is carbonaceous limestone equivalent to the Pine Point Formation).

Economic Geology

The lead-zinc occurrences at Robb Lake (*see* Taylor *et al.*, 1975) are in platform Devonian carbonate rocks adjacent to the Silurian-Devonian siltstone-quartz sandstone facies described above. What influence, if any, this facies change may have had on the localization of lead and zinc in adjacent carbonates has yet to be considered. If adjacent "basinal facies" rocks were the original hosts for lead and zinc, the siltstone-quartz sandstone facies may merit geochemical investigation.

References

Bally, A.W., Gordy, P.L., and Stewart, G.A.
1966: Structure, seismic data, and orogenic evolution of southern Canadian Rocky Mountains; Bull. Can. Pet. Geol., v. 14, p. 337-381.

Bamber, E.W., Taylor, G.C., and Procter, R.M.
1968: Carboniferous and Permian stratigraphy of northeastern British Columbia; Geol. Surv. Can., Paper 68-15.

Gibson, D.W.
1971: Triassic stratigraphy of the Sikanni Chief River-Pine Pass region, Rocky Mountain Foothills, northeastern British Columbia; Geol. Surv. Can., Paper 70-31.

Irish, E.J.W.
1970: Halfway River map-area, British Columbia; Geol. Surv. Can., Paper 69-11.

Norford, B.S., Gabrielse, H., and Taylor, G.C.
1966: Stratigraphy of Silurian carbonate rocks of the Rocky Mountains, northern British Columbia; Bull. Can. Pet. Geol., v. 14, no. 4, p. 504-519.

Pelletier, B.R.
1964: Triassic stratigraphy of the Rocky Mountain Foothills between Peace and Muskwa Rivers, northeastern British Columbia; Geol. Surv. Can., Paper 63-33.

Price, R.A. and Mountjoy, E.W.
1970: Geologic structure of the Canadian Rocky Mountains between Bow and Athabasca Rivers — a progress report; Geol. Assoc. Can., Spec. Paper 6, p. 7-39.

Stott, D.F.
1967: Fernie and Minnes strata north of Peace River, Foothills of northeastern British Columbia; Geol. Surv. Can., Paper 67-19, Part A.

Taylor, G.C. and MacKenzie, W.S.
1970: Devonian stratigraphy of northeastern British Columbia; Geol. Surv. Can., Bull. 186.

Taylor, G.C., Macqueen, R.W., and Thompson, R.I.
1975: Facies changes, breccias and mineralization in Devonian rocks of Rocky Mountains, northeastern British Columbia (94B, G, K, N); *in* Report of Activities, Part A, Geol. Surv. Can., Paper 75-1A, p. 577-585.

THERMISTOR CABLE INSTALLATION IN PERMAFROST MATERIALS WITH A WATER-JET DRILLING METHOD

Project 730006

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In an attempt to obtain subsurface temperature data in frozen surficial materials of the Tuktoyaktuk Peninsula-Mackenzie Delta Region, we are experimenting with a water-jet drilling technique. Water jetting in permafrost is well known in placer mining and the drilling technique has been used for installing piles in Alaska and northern Canada. Our application involves the installation of thermistor cables at shallow depths in a rapid, inexpensive manner.

With a relatively small water pump connected to a 1-inch I.D. steel pipe, we drilled a number of 3-m holes in frozen and unfrozen silts, sands and gravels during August 1975. No special preparation of the drilling end of the pipe was used. In unfrozen materials drilling rates exceeded 0.5 m/min. whereas in frozen materials drilling rates were in the order of 0.2-0.3 m/min. For these experiments we used sea-water at ≈ 6°C as a drilling fluid; the pump used was a WAJAX Mark 26 Centrifugal 2 stage unit capable of developing 150 psi.

Deeper drilling with this unit resulted in a hole to a depth of 20 m, limited only by lack of further drilling pipe. No appreciable change in the drilling rate was observed at depth.

To complete the preliminary experiments, a hole was drilled for a thermistor installation at the Polar Continental Shelf Project base at Tuktoyatuk. A depth of 20 m was achieved in 55 minutes through frozen sand and silt. The drill-rods were left in place and two thermistors were installed at 8 m and 15 m within 5 minutes of drillhole completion. Temperature readings commenced 30 minutes after cable installation and temperatures were monitored for 1750 minutes. Figure 100.1 shows the return of equilibrium of the drillhole; it is apparent that a "zero curtain" persisted for the first 1000 minutes representing refreezing of the surrounding walls of the drillhole. The apparent refreezing temperature of 0.08°C rather than 0°C falls within the calibration limit of ± 0.1°C and so the

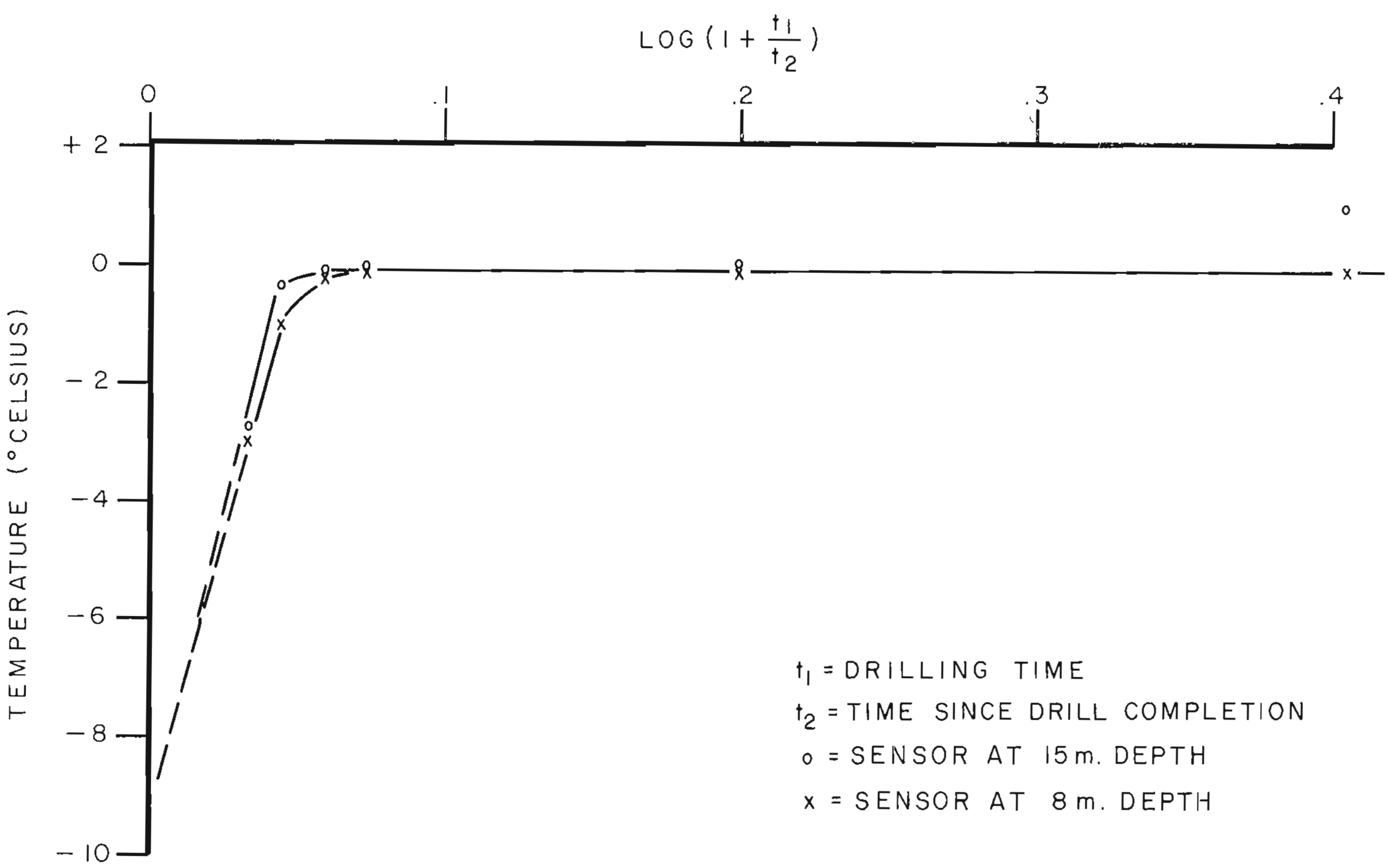

Figure 100.1.

[1]Earth Physics Branch.

difference is probably not significant. Once refreezing was complete the temperatures started to fall rapidly at a rate of approximately 0.005°C min^{-1}. If the logarithmic return to equilibrium is a valid extrapolation, the final equilibrium temperature is -8.8°C to -9.5°C; these results are consistent with other measurements in the Tuktoyaktuk area. If the extrapolation is valid, a very close approximation to equilibrium temperature is achieved in less than 10 days.

Measurements were again attempted a week later and both thermistor cables had failed. Attempts to recover the pipe and cables were impossible in the time available. The cause of failure is unknown, but may have resulted from differential freezing of the leads, as observed elsewhere. Displacing water in the drill pipe with antifreeze or the use of a stronger cable should overcome this.

The advantages of this technique of drilling lie in the speed of hole completion, the low weight and cost of the equipment, the simplicity of use and the ease of transportation. The major disadvantages are the need for a plentiful supply of water, which is pumped into the hole and surrounding media, and the lack of recovered cores. However, we believe that the water-jet drilling method has great promise as an inexpensive technique to acquire large amounts of shallow subsurface temperature information.

Acknowledgments

We wish to thank Dr. J. Ross Mackay for his support and the use of his water pump. We appreciate the long term loan of 21 m of water pipe made to us by the Polar Continental Shelf Project.

Project 740102

A.S. Judge[1], H.A. MacAulay[2], and J.A. Hunter[2]

A study has been completed of 250 km of first arrival data from marine seismic records provided by Pan Canadian Petroleum Limited and its partners, Gobles Oil and Gas Ltd., Amoco Canada Petroleum Company Limited, and Teck Corporation. The area covered by the survey lies within the boundaries of the "Mackenzie Canyon" as outlined by Shearer (1972) (*see* Fig. 101.1). Water depths over the area are generally less than 4 m

A high velocity refractor with an average velocity of 3000 m/s was observed over most of the area. Near the northeast offshore ends of the survey lines the refractor velocity decreases to 1800 m/s. An average velocity of 1520 m/s was used for the combined water-unfrozen sediment layer to calculate depths to the high velocity refractor. A contoured map of depths to the refractor is shown in Figure 101.2; the average depth is about 200 m but decreases to about 100 m near

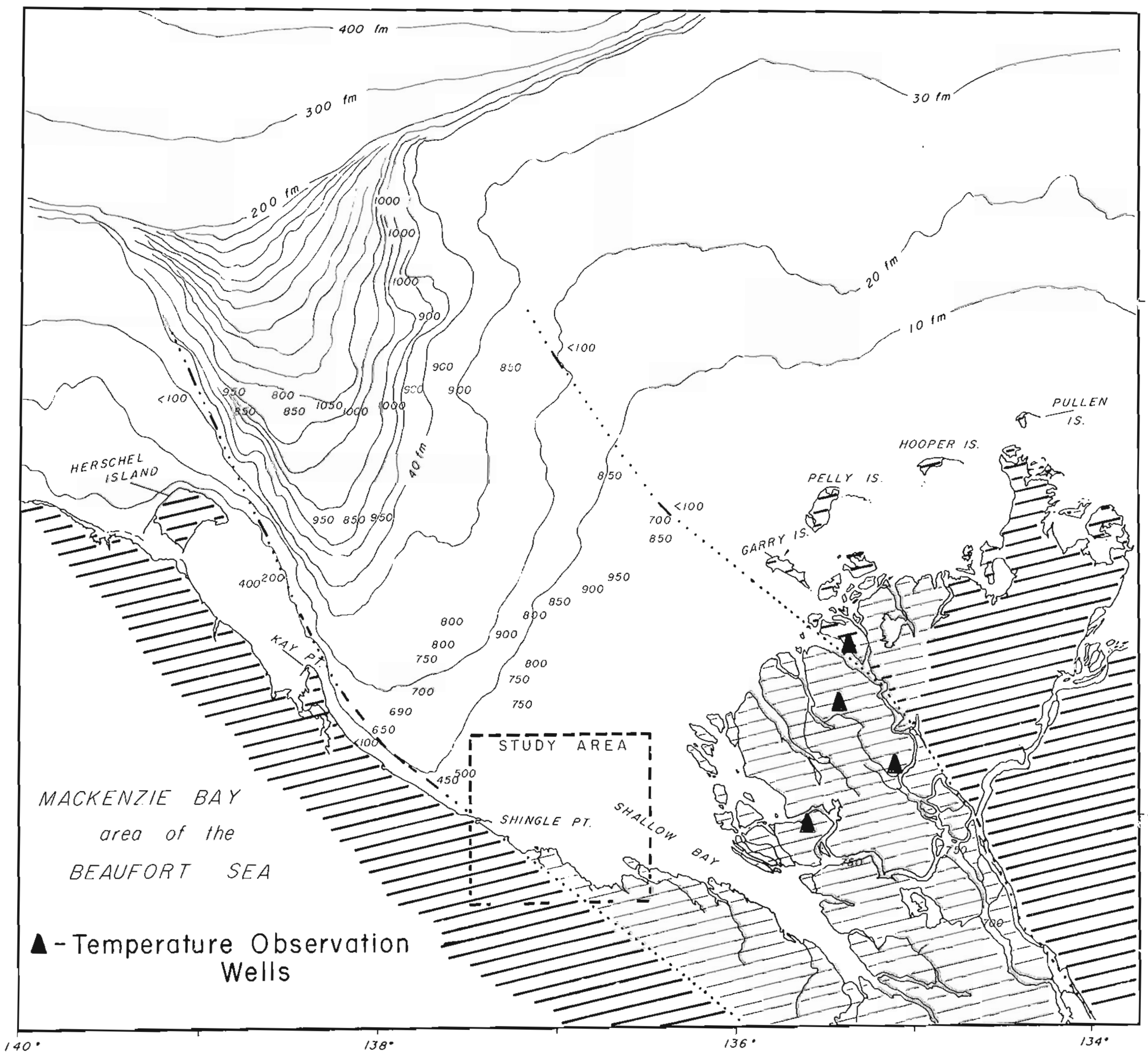

Figure 101.1. Location of the survey area in Mackenzie Bay. Offshore bathymetric contours show the extent and magnitude of the Mackenzie Canyon. Dotted Line denotes the interpreted edges of the buried ice scour channel, after Shearer 1972. Triangles show locations of deep temperature observation wells. Small figures are the depths in feet to the base of the scour channel and those outside of the channel represent thickness of sediment deposited in post scour times. Light hachured lines delineate distribution of modern Mackenzie Delta sedimentation: dark hachured, areas of older Quaternary sediment.

[1]Earth Physics Branch.

[2]Resource Geophysics and Geochemistry Division.

the shoreline. From a generalized velocity-depth function for Mackenzie Bay given by Hofer and Varga (1972), the high velocity refractor observed at these shallow depths is anomalous.

To seaward of the study area, Shearer (1972) has mapped the base of the buried Mackenzie scour channel from high resolution seismic reflection profiling. The depths obtained for the base of the channel are in the same range as the depths computed to the high velocity refractor in the study area.

Several records obtained near the shore indicate two high velocity events; the shallowest is derived from a refractor near the water-bottom interface followed by a deeper second event delayed in time by about 200 milliseconds. An example is shown in Figure 101. 3; the first event (A) attenuates rapidly and is interpreted to indicate a thin frozen layer; the second event (B) has a persistently high amplitude across the record, which merges into a wide angle reflection on traces close to the shot. The second event is that which is mapped throughout the survey area.

Included in Figure 101. 2 is an interpreted section beginning at the shoreline formed by recent deltaic sediments; the high velocity refractor apparently correlates with the bottom of the scour channel offshore but departs from it in the inshore region. Where the present shoreline coincides with the edge of the scour channel the top of the high velocity layer generally conforms to the base of the scour channel.

From laboratory studies by Nakano *et al.* (1971) a considerable velocity-temperature gradient may be found in coarse-grained non-saline water-saturated sediments at temperatures immediately below 0°C. Velocities of frozen sands are similar to those observed for the refractor. If the sediment type forming the bottom of the scour canyon is uniform and is below 0°C, then the seaward decrease of refractor velocity may result from increasing temperatures in frozen ice-bonded material.

Persuant to a permafrost interpretation in the vicinity of the shoreline the top of the frozen section would rise and the total thickness of it would probably increase due to the proximity of a cold northern shoreline of mean surface temperature -10°C compared with seabottom temperatures close to 0°C. However, the wedge-shape of the frozen ground persists to too great a distance offshore for an explanation based solely on the effect of the shoreline. A simple explanation of this elongation offshore could be found in a rapidly receding shoreline. At present the shoreline in the

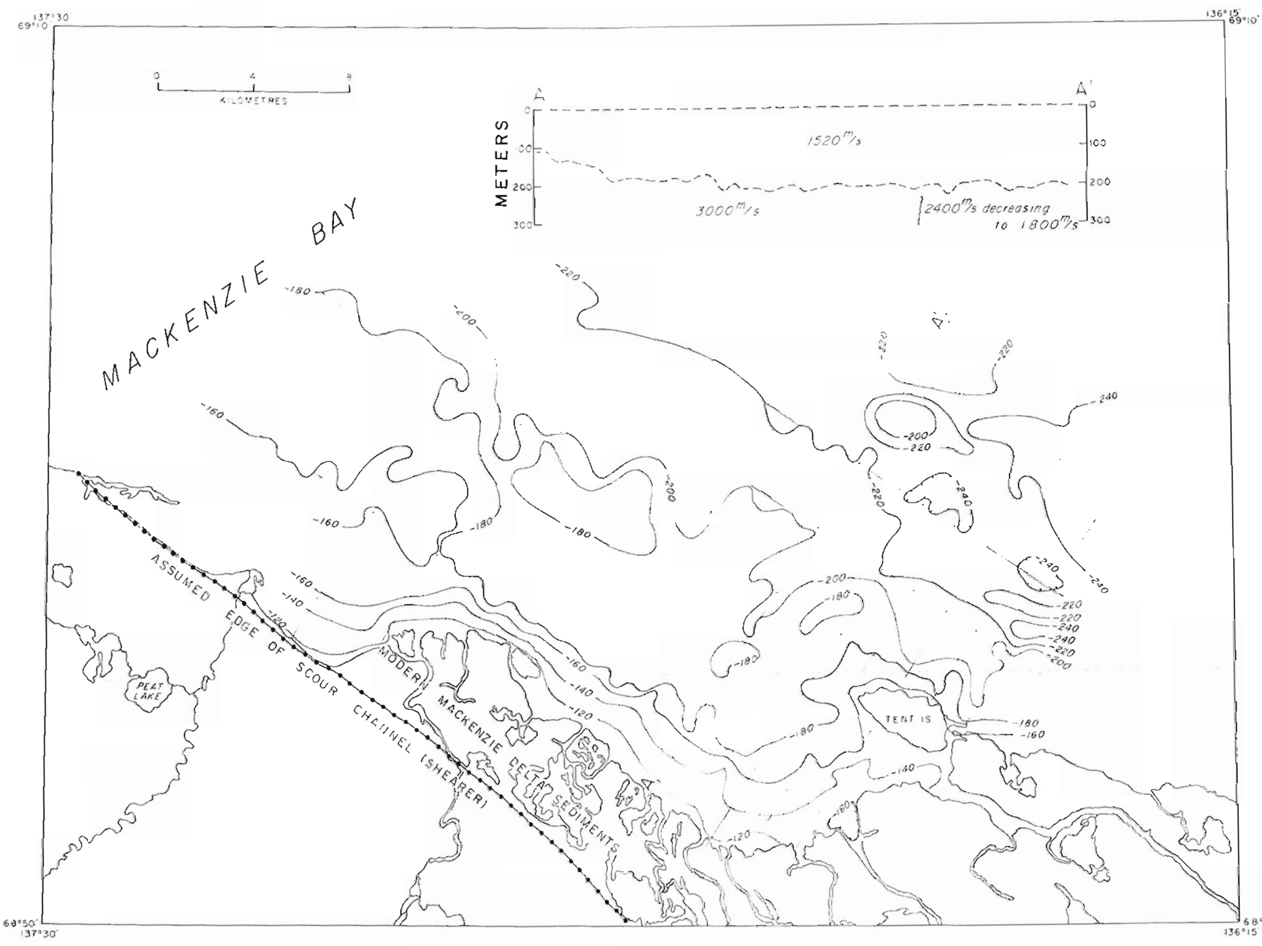

Figure 101. 2. Depths to high velocity refractor below sea level. Contours are in metres.

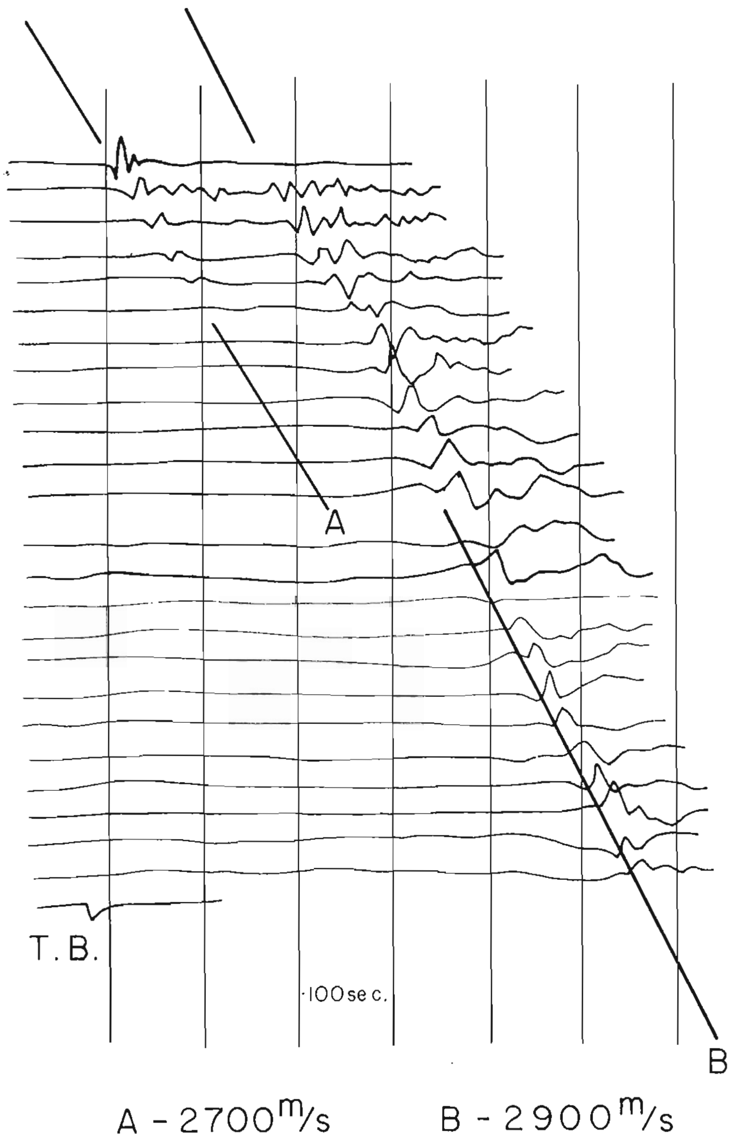

Figure 101.3. Marine refraction record showing two high velocity events indicating thin high-speed surface layer.

area is aggrading as is shallow permafrost evidence by a thin near-surface high velocity layer. Permafrost in the area is apparently very complex in nature. Aggradation is occurring at the present shoreline both from the surface and from an older shoreline producing a thin permafrost layer in the near-surface and a wedge-shaped section at depth extending out under the sea to a distance of several kilometres from the present coast. Hence, permafrost underlying the seabed remote from the shoreline would be relict in nature and degrading at present since present seabottom temperatures are too high to support deep permafrost in thermal equilibrium. Even if present seabottom temperatures in Shallow Bay were below 0°C, the terrestrial heat flux could only support a current permafrost thickness of 100 m. That same flux is sufficient to have also melted a maximum of 200 m of relict permafrost in the past 10 000 years.

Thus it is difficult to explain the high velocity refractor by the presence of the permafrost unless conditions were suitable in the past for thicknesses in excess of 300 m to have accumulated on the south side of Shallow Bay. As discussed by Judge (1974) the Mackenzie Canyon area of which it formed part underwent a very different history to that of the offshore north of Richards Island. For much of the Wisconsin it was covered by an ice sheet (Mackay *et al.*, 1972) and thus exposed to less severe surface temperatures than the unglaciated parts of the Beaufort Sea. Consequently permafrost at the end of the Wisconsin period was for example probably thinner than that on Richards Island today. At the time of recession of the ice sheet 14 000 to 16 000 yrs B.P., sea level was as much as 70 m below present. However, Shearer (1972) suggested that the base of the scour canyon lies at a depth of 200 m below present sea level. Therefore the rate of inundation of the area by the sea depends on the rate of accumulation of post-glacial sediments. Assuming an ice-base temperature during the Wisconsin of -2°C and an exposed land surface to have existed between glacial recession and the time of stabilization of sea level 5000 yrs B.P. as much as 200 m of permafrost could have grown. Decreasing the Wisconsin ice-base temperature increases this thickness by approximately 35 m per 1°C decrease. For the top of the frozen section to now be at a depth of 200 m, either very high rates of degradation or very rapid burial is necessary. Neither explanation seems likely.

In the central and northern parts of the Mackenzie Canyon area, seismic velocities indicate unfrozen sands at the base of the scour channel. These results are confirmed by subsurface temperature measurements made in the onshore portions of Shallow Bay (Taylor and Judge 1974). At each of these sites permafrost is relatively thin (60 to 150 m) and commences at the surface consistent with young sediments in which permafrost is aggrading. Positive temperature gradients of 22 to 45°C/km indicate no relict permafrost at depth. These observations place severe limits on the total amount of permafrost which could have accumulated in the Mackenzie Bay area.

It is perhaps worth noting that permafrost and hydrofrost (gas hydrates) may possess similar seismic velocities in coarse grained materials. Once again however the limited subsurface temperature measurements in the onshore portions of Shallow Bay would tend to rule out the presence of gas hydrates under equilibrium geothermal conditions unless they are of high specific gravity (> 0.6). Under highly non-equilibrium conditions thick, shallow gas reservoirs might remain at present in the hydrate form almost anywhere in Shallow and Mackenzie bays. A more reasonable history for the area such as that suggested in the permafrost interpretation could result in gas hydrate deposits at depths exceeding 100 m in areas that formed the margins of the Wisconsin ice lobe (the edges of the Canyon?).

Alternatively, the high refractor velocity may result from an older geological formation which has been structurally uplifted.

In conclusion, no definite interpretation of the seismic refractor can be made. It may be ice-bonded permafrost although from what is known of the surface history of the area it is difficult to explain its presence remote from the shoreline. It may be hydrofrost under suitable conditions which could pertain in certain parts of Mackenzie Bay or, yet a third alternative, it could represent an older uplifted formation. A final solution must await the acquisition of deep subsurface temperature profiles in the area.

This work reported here forms a part of a compilation of permafrost conditions in the Beaufort Sea and is being carried out with the close co-operation of all the exploration companies involved. Studies are nearing completion in adjacent areas and will be published shortly.

References

Hofer, H. and Varga, W.
1972: Seismogeologic experience in the Beaufort Sea; Geophysics, v. 37, no. 4, p. 605-619.

Judge, A.S.
1974: Occurrence of Offshore Permafrost in Northern Canada; *in* The Coast and Shelf of the Beaufort Sea, ed. by J.C. Reed and J.E. Sater; Arctic Institute of North America, p. 427-437.

Mackay, J.R., Rampton, V.N., and Fyles, J.G.
1972: Relict Pleistocene Permafrost, Western Arctic Canada; Science, v. 176, p. 1321-1323.

Nakano, Y., Smith, M., Martin, R., Stevens, H., and Knuth, K.
1971: Determination of the acoustic properties of frozen soils; report prepared by Cold Regions Research and Engineering Laboratory (CRREL) for Advanced Research Projects Agency (ARPA order no. 1525), 72 p.

Shearer, J.M.
1972: Geological structure of the Mackenzie Canyon Area of the Beaufort Sea; *in* Report of Activities, Part A, Geol. Surv. Can., Paper 72-1A, p. 179-180.

Taylor, A.E. and Judge, A.S.
1974: Canadian Geothermal Data Collection – Northern Wells, 1955 to February 1974, Geothermal Series No. 1, Can. Dep. Energy, Mines, Resour., Earth Phys. Br.

SEISMIC VELOCITY MEASUREMENTS ON SAMPLES AND SULPHIDE OREBODIES

Project 700061

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Experiments toward the detection and definition of sulphide orebodies by the seismic method have been continued. The initial attempt (Overton, 1972) using a highly redundant matrix of travel time measurements across an area containing a well defined orebody met with only marginal detection response. The attempt was unsuccessful in demonstrating a marked difference in seismic velocities between the orebody and its host rocks. Since this velocity contrast is crucial to the effectiveness of seismic methods, experiments were continued to investigate the seismic velocity characteristics of a massive orebody outcropping near Chapais, Quebec.

The outcrop area, being less than 15 m in diameter, required the use of the highest resolution recording equipment available. A high voltage power line traversing the test site created a high level of 60 hz electrical noise. Four recording systems were tested: An RS-49 amplifier system with a 60 hz notch filter was effective in counteracting the electrical noise problem, but the oscillograph speed of 127 cm/s did not provide the required resolution. An FS-3 hammer seismograph did not provide sufficient resolution and gave erratic readings caused by electrical noise. A Kelvin Hughes echo sounder using magnetostriction sound source (16 kilohz) and detector coupled to the ground through plastic water bags provided adequate resolution but encountered problems with the sound coupling and electrical noise. An electronic timing system with a 10 microsecond resolution provided the best timing resolution but was rendered ineffective by the high electrical noise level. Plans are continuing to deal with these problems preliminary to another field test.

A test was also conducted with seismic sources. The use of dynamite in the Province of Quebec is hampered by rigid control on purchase, use and storage. Also the use of explosives, in general, has required the costly and time consuming drilling of shotholes for effectiveness in energy coupling. Both these problems were solved by using a cannon constructed of a 20. 32 cm diameter by 20. 32 cm long steel cylinder with a 2. 54 cm diameter by 7. 62 cm long hole drilled for the detonation chamber. Shotgun powder, which may be purchased and used without restriction, was used effectively as the explosive energy source. The seismic energy was coupled to the ground by placing the loaded cannon on a 20. 32 cm^2 by 2. 54 cm thick steel breech block lying on the ground. Detonation was by electrical means but could also be accomplished by a mechanical or electromechanical trigger. The device produced seismic energy comparable to tamped charges in drilled shotholes and weighs 36 kg.

Experiments have been undertaken in the laboratory to resolve the erratic response of some seismic timing devices manufactured for close range velocity measurements in the field. Excellent correlation in results were obtained between one microtimer having a resolution of one microsecond, and a storage oscilloscope which facilitates timing directly from the recorded waveform, on aluminum rods and cylinders. The microtimer will now be tested under field conditions at the earliest opportunity. The two systems also afford a means for laboratory determination of seismic velocities in rock samples.

Reference

Overton, A.
1972: A seismic experiment over a metallic sulphide deposit; *in* Report of Activities, Part A, Geol. Surv. Can., Paper 72-1A, p. 57.

PORTABLE BOREHOLE GAMMA-RAY SPECTROMETER TESTS

Project 740085

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As part of an ongoing project to investigate nuclear techniques for borehole logging (Killeen, in prep.) two portable borehole gamma-ray spectrometry systems were tested and evaluated. The systems are built around the Scintrex GAD-1 and the Exploranium DISA-400A portable gamma-ray spectrometers.

In addition to evaluating the characteristics of two commercially available borehole spectrometers, the test results provide valuable information to be used in determining parameters for calibration facilities to be constructed for borehole radiometric equipment.

Most of the test measurements were carried out in the Bancroft area, with some additional work in holes in the Elliot Lake, South March, and the Sudbury areas. The latter tests were to determine the potential of borehole gamma-ray spectrometry as a lithological mapping aid in a non-uraniferous environment, and to yield preliminary data for a study of base metal-radioelement relationships.

Each borehole gamma-ray spectrometer system consisted of the spectrometer, chart recorder, interface unit, winch, cable, cable head, and borehole probe

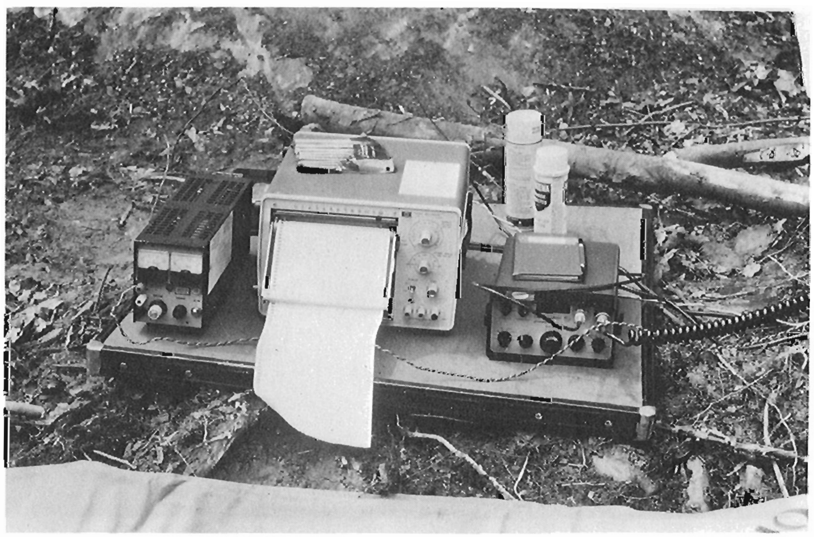

Figure 103.1a

The DISA-400A gamma-ray spectrometer and chart recorder.

Figure 103.1b

The Exploranium motorized winch and tripod operating from the tailgate of the truck.

Figure 103. 2a

The GAD-1 gamma-ray spectrometer and chart recorder.

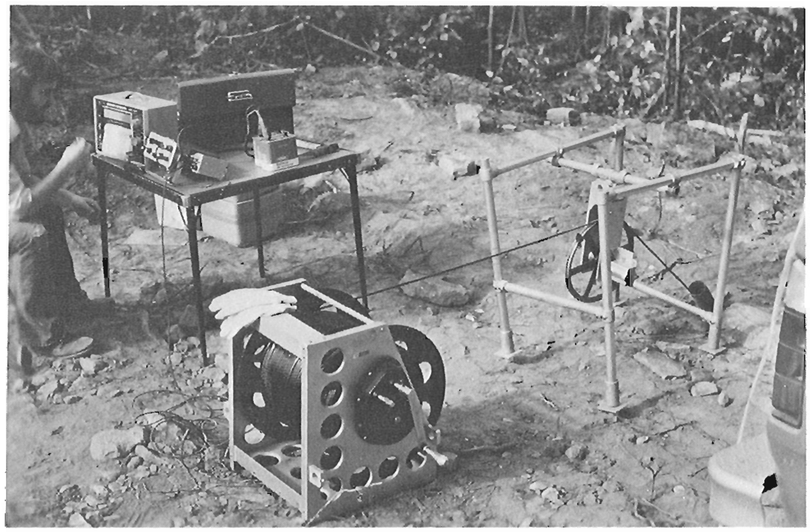

Figure 103. 2b

The Scintrex system showing the winch and quadrupod assembly.

containing the NaI(T1) detector. A tripod or quadrupod with a wheel and counter were also used to feed the cable from the winch into the hole, and measure the depth to the probe. The entire system was transported in a three quarter ton, 4-wheel-drive pickup truck (Fig. 103. 1b). Figures 103. 1a and 103. 1b illustrate the Exploranium instruments with the Exploranium winch and tripod. Figures 103. 2a and 103. 2b show the Scintrex instruments with the Scintrex quadropod assembly.

Because of the 500 m cable length which was being used, neither system was 'backpack' portable. The winches could be carried by two men with the aid of carrying rods which slipped into brackets on the side of the winch. For shallow holes, a short cable and small manual winch could make either system truly 'backpack' portable. The winch supplied by Exploranium was motorized, requiring a gasoline generator for power. This made it even less portable, but was considered invaluable for logging the deeper holes. A valuable feature which was not available with either system would be a method of synchronizing the movement of the chart paper with the lowering of the probe in the hole making the depth scale of the chart independent of variations in logging speed. Also a fiducial or depth marker on the chart recorder triggered by the counter wheel would be an advantage.

Perhaps one of the greatest drawbacks of both systems was their inability to record the output from more than one channel at a time.

In the case of both borehole systems tested, in order to record four windows, the total count could be recorded going down the hole, potassium coming up, uranium going down, and finally thorium coming up. This meant that the entire length of the borehole had to be traversed at logging speed four times. Suitable logging speeds were 3 m per minute or less, resulting in a time of about 50 minutes to log a 150 m hole with only one spectrometer window. Manual control of the

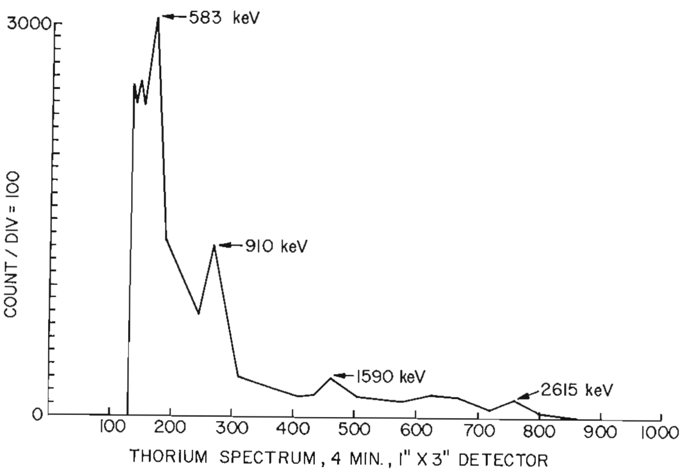

Figure 103. 3. 1024 channel gamma ray spectrum of thorium recorded from a 1 inch by 3 inch NaI(T1) detector.

winch resulted in slight variations in the speed and varying lengths of records of total count, potassium, uranium, and thorium for a given hole. The chart-winch synchronization mentioned above could alleviate the problem, but a four channel chart recorder with a spectrometer capable of four simultaneous outputs would be a better solution, saving considerable time. This is certainly available, but the portability of the system is decreased, and cost increased. In this context, a four channel gamma-ray spectrometer and recorder originally designed for use in a helicopter was tested, utilizing the signals from the Exploranium probe. The advantage of decrease in logging time, and improvement of correlation between records from the four spectrometer windows was confirmed.

Considerable testing of both borehole spectrometric systems was carried out in the lab before taking the equipment to the field. Gamma-ray energy spectra were recorded from the three borehole probes used in the test. Figure 103. 3 shows a 1024 channel spectrum of thorium from a 1 inch x 3 inch NaI(T1) detector housed in a 1. 50 inch O. D. probe. It can be seen that the thorium peak (T1-208) at 2. 62 Mev is quite small due to the low efficiency of the small crystal for detecting high energy gamma-rays.

In general it can be concluded that both borehole gamma-ray spectrometry systems worked, but are not ideal in their present configuration. A more complete comparison of gamma-ray spectra from the three different probes; field recorded 1024 channel gamma-ray spectra; and potassium, uranium, thorium and total count logs recorded by the different probes in the same hole are in preparation.

Reference

Killeen, P. G.
Nuclear techniques for borehole logging in mineral exploration; *in* Borehole Geophysics Applied to Metallic Mineral Prospecting -- A Review; ed. by A. V. Dyck, Geol. Surv. Can., Paper 75-31. (in prep.)

104. GEOLOGICAL INVESTIGATIONS OF SOMERSET ISLAND AND BOOTHIA PENINSULA (OPERATION BOOTHIA; 57 C, F, G; 58 A to F) DISTRICT OF FRANKLIN

Project 740030

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Institute of Sedimentary and Petroleum Geology, Calgary

Introduction

Operation Boothia is a project of the Geological Survey of Canada initiated for the purpose of studying the various aspects of the bedrock geology of Boothia Peninsula and Somerset Island, N. W. T. During the first field season, from June 17 to August 26, 1975, the northern half of Somerset Island was completed.

The base camp was situated at the mouth of the Hunting River on Aston Bay. A twin Otter aircraft based in Resolute was chartered to position and reposition the party, bring in supplies, and establish certain fly camps. Transport within the field area was by G4A helicopter and Beaver airplane. Jean Stewart (cook), Sandy Denton (geological assistant), Bill Christensen (camp manager), Carl Faulkner (helicopter pilot), Marty Lukst (helicopter engineer), and Carl Zberg (Beaver pilot) provided support for the scientific staff.

Spring was particularly early so that snow cover was largely gone by July 1. The month of July was characterized by almost continuous stormy weather, and work was hampered greatly by fog, rain, high winds, and occasional snow. The month of August was quite clear with good working weather; however, daily snowstorms that began on August 24 forced an early stop to geological work.

There are no settlements on Somerset Island and residents consist of only one Eskimo family who live periodically on the island for extended periods. Several scientific, exploration and other parties were based on Somerset Island in 1975, including other Geological Survey parties that have been reported in this volume, a mineral exploration party of Diapros Canada Ltd., as well as a federal fisheries study group, and an Eskimo fishing camp at Stanwell Fletcher Lake.

Previous geological investigations

Although the earliest geological observations of Somerset Island were made well over one hundred years ago by early explorers, systematic studies did not start until 1955. In that year, the Geological Survey of Canada studied Somerset Island as part of its early reconnaissance, Operation Franklin. A volume resulting from that project (Fortier *et al.*, 1963) contains several papers in which are established the main features of the stratigraphy and new formation names; the papers are accompanied by a geological map. In that project, the area near Aston Bay was studied by Blackadar (1963) who named the Proterozoic Aston and Hunting formations and described the younger Paleozoic formations. Formations of the north coast were described by Thorsteinsson and Tozer (1963) and McMillan (1963). An early paper on Cornwallis Island by Thorsteinsson (1958) provided a foundation for much of the stratigraphy of Somerset Island.

A later reconnaissance project of the Geological Survey in 1962 produced a more detailed geological map and stratigraphic report of Somerset Island (Blackadar and Christie, 1963). A study of the crystalline rocks of the Shield also resulted from that project (Blackadar, 1967). A paper by Kerr and Christie (1965) summarizes the knowledge to that time of the tectonic evolution of the Boothia Uplift, the main structural feature trending through Somerset Island.

The University of Ottawa has conducted field work on Somerset Island each year for a number of years. D. Dinely initiated the project and led it for several years. At present it is under the leadership of O. Dixon. A large number of papers have resulted from the efforts of the University of Ottawa. Only the most recent papers, providing pertinent background to the present bedrock study, are mentioned below. The Precambrian Shield rocks of Somerset Island were studied by Brown *et al.* (1969). The Proterozoic Hunting Formation was studied by J. Dixon (1974). Dixon also (1973) named and described the Lang River Formation, of western Somerset Island which is the basal unit of the Paleozoic (pre-Allen Bay) succession. Miall (1970) did important work on the Peel Sound Formation of nearby Prince of Wales Island that bears directly on the stratigraphy of Somerset Island. Jones and Dixon (1975) described the sedimentary column near Port Leopold, on northeastern Somerset Island, and proposed a new name, the Leopold Formation.

Objectives and method of current project

The project area of Operation Boothia includes all of Somerset Island and Boothia Peninsula. The objectives of the project are: (a) to produce up-to-date bedrock geological maps on 1: 125 000 or 1: 250 000 scale, as is appropriate; (b) to produce a comprehensive geological report that covers all aspects of the bedrock geology of the project area in order to provide essential information for construction of a proposed natural gas pipeline, and for the discovery and evaluation of potential resources.

The writer is responsible for co-ordinating the scientific aspects of the project as well as for its field administration. The collaborating scientists are responsible for planning and executing different aspects of the geology as follows:

Structure and petrology of the Precambrian Shield
C. de Vries, University of Calgary

Stratigraphy of Proterozoic and younger sediments
G.E. Reinson

Sedimentological study of the Aston Formation
D. Stewart, University of Calgary

Structural geology of post-Shield rocks and relationship to the Shield
J. Wm. Kerr

Geology of kimberlite pipes
R. Mitchell, Lakehead University

In addition to the above mentioned fields, support and collaboration were provided to related projects as follows:

Silurian stratigraphy – This is the subject of study by O. Dixon and B. Jones of the University of Ottawa. They were provided with transportation and assistance.

Paleomagnetic studies – P. LaPointe of the Earth Physics Branch, Ottawa, has begun a study of paleomagnetism of the Proterozoic Aston Formation and the Devonian Peel Sound Formation. He was provided with transportation assistance and advice.

Earthquake risk – A portable seismometer was set up at base camp and monitored for the summer. This was done in collaboration with A. Stevens of the Earth Physics Branch, Ottawa. The purpose of the study is to determine whether substantial seismic activity occurs in Barrow Strait and other parts of the Arctic.

The geological results obtained in 1975 are summarized in three separate contributions to this volume by Reinson, Kerr and Stewart (rep. 106); Kerr and de Vries (rep. 105); and Mitchell (rep. 107).

References

Blackadar, R.G.
1963: Somerset Island between Aston Bay and Creswell Bay; *in* Operation Franklin; Geol. Surv. Can., Mem. 320, p. 143-150.

1967: Precambrian geology of Boothia Peninsula, Somerset Island, and Prince of Wales Island, District of Franklin; Geol. Surv. Can., Bull. 151.

Blackadar, R.G. and Christie, R.L.
1963: Geological reconnaissance, Boothia Peninsula, and Somerset, King William, and Prince of Wales Islands, District of Franklin; Geol. Surv. Can., Paper 63-19.

Brown, R.L., Dalziel, I.W.D., and Rust, B.
1969: The structure, metamorphism and development of the Boothia Arch, Arctic Canada; Can. J. Earth Sci., v. 6, no. 4, p. 525-543.

Dixon, J.
1973: Stratigraphy and Sedimentary history of early Paleozoic rocks from Prince of Wales and Somerset Islands, N.W.T. *in* Aitken, J.D. and Glass, D.J., eds., Proc. Symp. Geol. Arctic; GAC-CSPG, Saskatoon, 1973, p. 127-142.

1974: Revised stratigraphy of the Hunting Formation (Proterozoic), Somerset Island, Northwest Territories; Can. J. Earth Sci., v. 11, no. 5, p. 635-642.

Fortier, Y.O., Blackadar, R.G., Glenister, B.F., Greiner, H.R., McLaren, D.J., McMillan, N.J., Norris, A.W., Roots, E.F., Souther, J.G., Thorsteinsson, R., and Tozer, E.T.
1963: Geology of the north-central part of the Arctic Archipelago, Northwest Territories (Operation Franklin); Geol. Surv. Can., Mem. 320.

Jones, B. and Dixon, O.A.
1975: The Leopold Formation: An Upper Silurian intertidal/supratidal carbonate succession on northeastern Somerset Island, Arctic Canada; Can. J. Earth Sci., v. 12, no. 3, p. 395-411.

Kerr, J.Wm. and Christie, R.L.
1965: Tectonic history of Boothia Uplift and Cornwallis Fold Belt, Arctic Canada; Bull. Am. Assoc. Pet. Geol., v. 49, p. 905-926.

McMillan, N.J.
1963: North coast of Somerset Island *in* Operation Franklin; Geol. Surv. Can., Mem. 320, p. 129-139.

Miall, A.D.
1970: Continental marine transition in the Devonian of Prince of Wales Island, Northwest Territories; Can. J. Earth Sci., v. 7, no. 1, p. 125-144.

Thorsteinsson, R.
1958: Cornwallis and Little Cornwallis Islands, District of Franklin, Northwest Territories; Geol. Surv. Can., Mem. 294. (1959).

Thorsteinsson, R. and Tozer, E.T.
1963: Geology of Northern Prince of Wales Island and northwestern Somerset Island *in* Operation Franklin; Geol. Surv. Can., Mem. 320, p. 117-129.

STRUCTURAL GEOLOGY OF SOMERSET ISLAND, DISTRICT OF FRANKLIN

Project 740030

J. Wm. Kerr[1] and C. D. S. de Vries[2]

Introduction

The geology of Somerset Island is dominated by the Boothia Uplift and Cornwallis Fold Belt, a pair of related north-trending structural features. The Boothia Uplift is part of the Churchill Province of the Canadian Shield, and the Cornwallis Fold Belt is the folded sedimentary cover overlying and adjacent to it. Positive and negative movements of the uplift produced the structures of the overlying fold belt (Kerr and Christie, 1965). A prime objective of the present study is to relate the timing and nature of deformations in the Shield and the overlying cover rocks. The work on the crystalline rocks was done by de Vries, and on the sedimentary rocks by Kerr.

Canadian Shield

The Precambrian basement rocks forming the core of the Boothia Uplift on Somerset Island comprise a thick sequence of high-grade gneisses and migmatites (Blackadar, 1967). A significant part of these metamorphic rocks is thought to be of sedimentary and volcanic origin. Abundant quartzites, rusty graphitic gneisses, banded gneisses and diopside marbles commonly are interlayered with poorly foliated hornblende- and biotite-bearing quartz-feldspar gneisses of less certain derivation. Metapegmatites and foliated migmatites occur intimately interlayered with paragneisses. Large, more or less homogeneous gneissic masses with schlieren-like structures are rare, suggesting that wholesale granitization of the metamorphic rocks was of limited extent.

Thin, but remarkably continuous amphibolites and pyroxene amphibolites with possibly some ultrabasic portions occur as conformable layers or as strings of rafted fragments in felsic gneisses. One locality was found where a pod-shaped body of pyroxene amphibolite, at least 100 m (328 ft.) in diameter, is discordant with the gneissic layering and suggests the intrusion of diabasic material in the metasediments prior to penetrative deformation and metamorphism.

The occurrence near M'Clure Bay on Peel Sound of a largely unfoliated mass of homogeneous granite, which displays clearly discordant relationships with the bordering gneisses and migmatites, marks a distinctly younger geological event. Slightly discordant and, in some places, unfoliated pegmatites could be associated with this late intrusion. The age of the emplacement of the discordant granite is unknown at present and must await radiometric dating. Structural relationships, however, suggest that it may be similar in age to the late-stage discordant granites of Hudsonian age described by Schau (1975) in the Hayes River region.

Structure of the Shield

The structure of the Precambrian basement rocks of Somerset Island is dominated by approximately north-south trending, tight folds with upright to slightly overturned axial planes. Individual folds can be traced best in the southernmost part of Somerset Island, particularly in the area between Fitz Roy Inlet and Bellot Strait, owing to the presence of good lithological markers and, in general, steeper axial plunges. In the northern part of Somerset Island, only a few major fold structures can be mapped, but numerous, isolated, tight hinges of folds can be recognized on aerial photographs. Macroscopic fold axes, where they can be traced for any distance on the map, generally show considerable change in amount and direction of plunge, although shallow plunges predominate. Several excellent exposures of doubly plunging, canoe-shaped synforms occur along the coast in the Bellot Strait area.

Mesoscopic folds display a disharmonic style that may be characteristic also of the macroscopic structures. The fold style can change from open concentric to tight and nearly similar over a distance of 50 cm (20 in.) across the layering.

The most prominent planar fabric element, apart from compositional layering (S_1), is flattened or platy quartz. It defines a foliation that can be recognized in most quartz-rich rocks throughout the area. Brown *et al.* (1969) used the relationship between the orientation of the platy quartz foliation (S_2) and the axial planes of folds in the compositional layering to distinguish between an older, isoclinal set of folds (F_1) and a younger set of tight folds (F_2), in which the quartz foliation runs parallel to the axial planes. The majority of the folds mapped in the field belong to the F_2 generation of structures. F_1 folds are rare and seem to occur as steeply plunging intrafolial folds in the transposed foliation that now forms the compositional layering (Brown *et al.*, 1969). In the Fitz Roy-Bellot Strait area one major antiform was mapped that has a nearly vertical plunge and may be the result of interference between F_1 and F_2 folding.

Metamorphism of the Shield

Both Blackadar (1967) and Brown *et al.* (1969) have remarked on the lack of systematic regional variation in the mineral assemblages of Somerset Island and Boothia Peninsula, thus precluding the drawing of isograds. Mineral assemblages characteristic of the hornblende granulite facies are present throughout the

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map-area. It appears that stable conditions of upper amphibolite to granulite facies metamorphism have been attained over a large area in the northern sector of the Churchill Province during middle to late Aphebian time.

The lack of isograds that can be mapped on the basis of criteria observed in the field would seem to suggest that the basement and the isogradic surfaces within it have not been arched upward during the various pulses of uplift that brought the basement to the surface intermittently from late Hadrynian to Devonian time. The basement probably behaved as rigid blocks that were faulted during uplift in that interval.

Cornwallis Fold Belt

In the latitude of the project area, the Cornwallis Fold Belt occupies two narrow belts of outcrop that are separated by the Boothia Uplift (Kerr and Christie, 1965). The eastern belt is on Somerset Island and trends from south to north for the full length of the island. Two Proterozoic formations, the Aston and the Hunting formations (Blackadar, 1963), are considered here for the first time to be part of the Cornwallis Fold Belt.

The Cornwallis Fold Belt has an overall northerly trend that is parallel to the overall trend of the Boothia Uplift, and is traced from Bellot Strait in the south to west of Pressure Point in the northwest. Dips in the fold belt immediately east of uplift are steeply eastward to vertical but, traced eastward, they rapidly diminish to become nearly horizontal.

The Cornwallis Fold Belt on Somerset Island is not straight, but rather is made up of short, straight, north-trending segments connected by shallow, north-dipping segments. The jogs in the fold belt probably reflect the northern ends of underlying individual uplifted basement blocks that may be faulted.

The Cornwallis Fold Belt plunges generally to the north, roughly parallel to the shape of the surface of the underlying Precambrian basement. Northeast of Cape Granite, a long stratigraphic section of the Hunting Formation resting unconformably on the Shield dips northeast at approximately 15 degrees and occupies an outcrop breadth of 9.6 km (6 miles). This dip indicates a strong northward component to the slope of the basement surface at the north end of the exposed Boothia Uplift.

Certain important discoveries concerning tectonic events that affected the Cornwallis Fold Belt were made; the localities are given by UTM co-ordinates and the features are described below. It had earlier been thought that the prominent dyke swarm of northwestern Somerset Island was a single swarm that post-dated the Hunting Formation. It is now known that there were at least two phases in this intrusive swarm. The youngest does in fact cut the Hunting Formation. The other, however, preceded that formation and includes dykes, sills, and at least one plug. It is documented by an angular unconformity at UTM 15X; 446,050mE, 8,165,250mN where the Hunting Formation lies on a truncated plug that itself intrudes the Aston Formation.

It is now known also that the Aston and Hunting formations are not a simple homocline, but were deformed severely into northerly-trending folds. The dating of these folds is still in question. A western outlier of the Lang River Formation that is included in these folds lies on the Hunting Formation with what appears to be only slight angularity at UTM 15X; 436,800mE, 8,171,350mN. In itself, this suggests that these folds developed during one of the Paleozoic or younger deformations of the Cornwallis Fold Belt. On the other hand, the dykes that intrude Hunting Formation clearly postdate the folding of those formations. This is particularly evident in the case of a long northeast-trending dyke at UTM 15X; 438,500mE, 8,170,000mN. This latter evidence suggests that the folding of the Aston and Hunting formations is of pre-Lang River, and presumably of Proterozoic age. Further work should resolve this problem.

A revised summary of early history of the deformed belt is as follows. Northerly-trending deformation has affected western Somerset Island intermittently since Aphebian time. F_1 and F_2 were by flow and these created the strong northerly-trending foliation of the Shield. Post-F_2 granitic intrusion of Hudsonian(?) age marked the end of the deep-seated history of the basement. Uplift and planation of the Shield were followed by the unconformable deposition of the Proterozoic Aston Formation. The deformations that are younger than the Aston Formation appear to have produced only brittle deformation in all rocks presently exposed at the surface. The sequence of events in the Cornwallis Fold Belt that is documented on Somerset Island is as follows:

1. Deposition of the Aston Formation (sandstone and conglomerate) on the crystalline basement complex.
2. Broad arching of the Boothia Uplift on western Somerset Island and intrusion of a diabase dyke swarm, sills and plugs into the crystalline basement complex and into the Aston Formation.
3. Erosion of the entire Aston Formation from extreme western Somerset Island, including the area near Cape Granite.
4. Westward encroachment of the Hunting Formation (dolomite with basal conglomerate) across the eroded Aston Formation and its contained dykes to lie unconformably on the Shield at Cape Granite.
5. Broad arching of the Boothia Uplift on western Somerset Island accompanied by intrusion of a second diabase dyke swarm, this one also cutting the Hunting Formation.
6. Deep erosion occurred, presumably following this latest deformation and intrusion. It was greatest in the south and west where the entire Hunting Formation was removed from the Shield.

7. The Lower Paleozoic Lang River Formation (Dixon, 1973) encroached westward to be deposited unconformably upon the Shield, the Aston Formation, and the Hunting Formation. A sequence of younger events that affected the deformed belt has been described by Kerr and Christie (1965).

References

Blackadar, R. G.
1963: Somerset Island between Aston Bay and Creswell Bay, *in* Operation Franklin; Geol. Surv. Can., Mem. 320, p. 143 150.

1967: Precambrian Geology of Boothia Peninsula, Somerset Island, and Prince of Wales Island, District of Franklin; Geol. Surv. Can., Bull. 151.

Brown, R. L., Dalziel, I. W. D., and Rust, B. R.
1969: The structure, metamorphism, and development of the Boothia Arch, Arctic Canada; Can. J. Earth Sci., v. 6, no. 4, p. 525-543.

Dixon, J.
1973: Revised stratigraphy of the Hunting Formation (Proterozoic), Somerset Island, Northwest Territories; Can. J. Earth Sci., v. 11, p. 635-642.

Kerr, J. W. and Christie, R. L.
1965: Tectonic history of Boothia Uplift and Cornwallis Fold Belt, Arctic Canada; Bull. Am. Assoc. Pet. Geol., v. 49, no. 7, p. 905-926.

Schau, M.
1975: Gneiss distinctions in the Hayes River Region: magnetic and geochemical parameters; *in* Report of Activities, Part B, Geol. Surv. Can., Paper 75-1B, p. 89-96.

STRATIGRAPHIC FIELD STUDIES, SOMERSET ISLAND, DISTRICT OF FRANKLIN (58 B to F)

Projects 750084 and 740030

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Introduction

During the 1975 field season, a ten week field program was conducted on Somerset Island. Reinson's work (Project 750084) included the gathering of data on the stratigraphy, paleontology and sedimentology of the Paleozoic and Proterozoic sedimentary succession of northern Somerset Island. This work was confined largely to the Proterozoic Hunting Formation, and the Silurian and Devonian Read Bay-Peel Sound sequence (Thorsteinsson and Tozer, 1963; Blackadar and Christie, 1963). Kerr was responsible for the planning and logistics of the overall field program (*see* Kerr, 1976), and worked closely with Reinson on stratigraphic and mapping problems. Stewart studied the Aston Formation; he is presently preparing a Master's thesis on this subject at the University of Calgary. In addition, Reinson collaborated with P. Lapointe, Earth Physics Branch, Energy, Mines and Resources, and B. Jones, University of Ottawa, during field investigations of the Peel Sound red beds (Lapointe) and the Read Bay and Leopold formations (Jones).

Proterozoic stratigraphy

Sedimentary strata, of Proterozoic age (Dixon, 1974), outcrop over a limited area south of Aston Bay (Fig. 106.1). These rocks were described first by Blackadar (1963), who recognized two formations, the Aston Formation, which is mainly quartzite, and the overlying Hunting Formation, which is mainly dolomite.

Aston Formation

The Aston Formation has an estimated maximum thickness of 1400 m; the base of the formation, where exposed, is marked by a breccia unit with large subangular gneiss boulders, resting unconformably on the Precambrian Shield. The lower part of the formation consists largely of well sorted orthoquartzite. The dominant lithology of the immediately overlying strata is silica-cemented red sandstone and minor conglomeratic sandstone. These exhibit abundant low-angle, large-scale trough and planar crossbeds in association with small scale cross-laminae, ripple-marks and parallel laminae. Three major units of purple and red, sandy siltstone occur near the base and in the lower middle and middle upper parts of the section. The latter two appear laterally discontinuous. In the lower and upper siltstone units, localized colonies of club-shaped columnar stromatolites occur. The middle siltstone unit displays abundant desiccation cracks and ripple-marks.

The extreme textural maturity of the majority of the sandstone units in the succession is indicative of intense reworking and is suggestive of beach and near-shore environments. The low angle trough and planar cross-sets could have developed in migrating offshore bars.

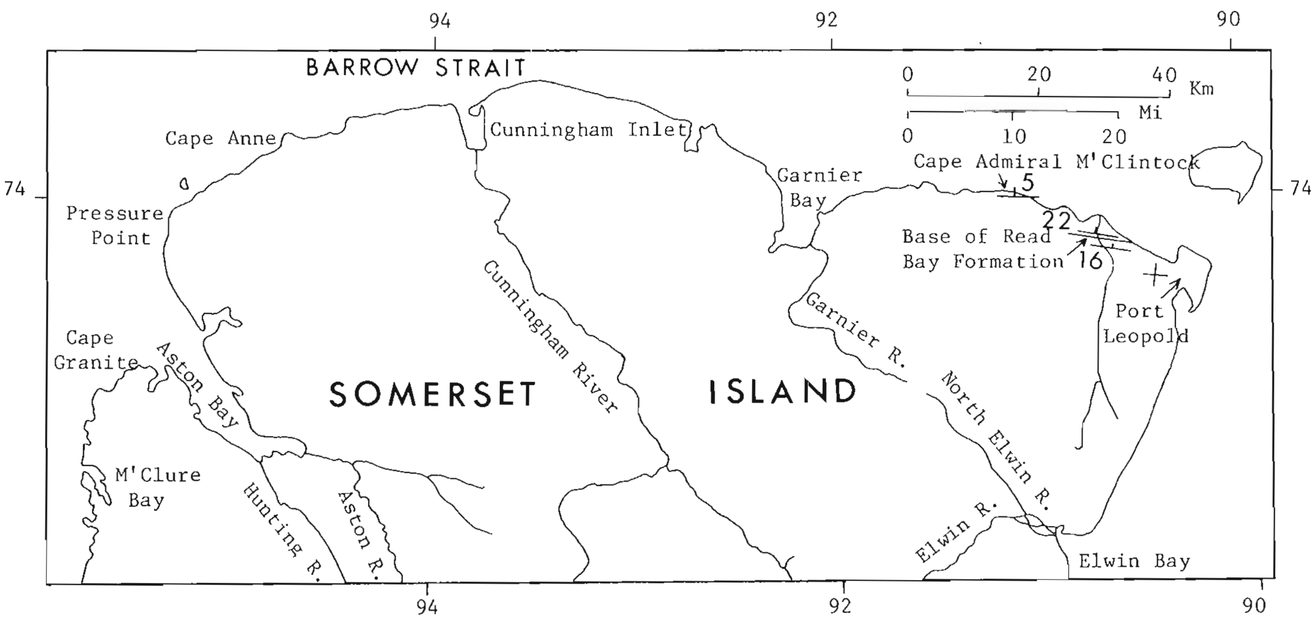

Figure 106. 1. Index map of northern Somerset Island showing localities mentioned in the text, and the location of the base of the Read Bay Formation northwest of Port Leopold.

Table 106. 1

Silurian-Devonian stratigraphy of northwestern Somerset Island

	Formation	Lithology	Thickness in metres
Upper Silurian and/or Devonian	Peel Sound	Red sandstone, siltstone, conglomerate; thin to thick bedded	320+
Upper Silurian and/or Devonian	Unnamed	Red- and green-weathering, olive-grey, calcareous and dolomitic siltstone, silty limestone; thin bedded	150+
Upper Silurian	Read Bay	Grey to dark grey, lumpy limestone, in part dolomitic, irregularly bedded, fossiliferous	180-240
Lower to Upper Silurian	Cape Storm	Grey to grey-brown dolomite, limestone, and platy calcareous sandstone; thin to medium bedded	175+
Ordovician-Silurian	Allen Bay	Tan to grey dolomite, medium to thick bedded	

The intervals of siltstone suggest intertidal flats. Thin, truncated, fining-upward cycles at the top of the section indicate deposition in fluvial environments.

Hunting Formation

The Hunting Formation overlies the Aston Formation unconformably and is about 1000 m thick. The lower part of the formation is characterized by thin- to medium-bedded, parallel laminated to homogeneous, buff-grey dolomite. The middle part consists of thick-bedded, parallel to wavy laminated, grey dolomite units alternating with columnar and domal stromatolite units. The laminated and dendritic stromatolitic structures are accentuated in places by secondary chert, which forms 50 per cent of the mineralogy in some units. The upper part of the formation is characterized by medium- to thick-bedded alternating units of buff dolomite with laminated intraclastic fabrics and teepee structures, and pink-grey dolomite with laminated-fenestral fabrics. Quartz sand content increases upsection in the upper part of the formation, and red sand stone and siltstone lithologies occur periodically near the top of the formation.

Silurian-Devonian stratigraphy

Northwestern Somerset Island

During detailed stratigraphic studies around the Cape Anne syncline (northwestern Somerset Island), it became apparent that the Silurian-Devonian succession could be divided into more rock-stratigraphic units than had previously been designated. Thorsteinsson and Tozer (1963) and Blackadar and Christie (1963) did the reconnaissance stratigraphy of this region and recognized, in ascending order, the Allen Bay Formation (Ordovician to Silurian), Read Bay Formation (Silurian), and the Peel Sound Formation (Siluro-Devonian). This general sequence was accepted by the writers but refined to include two additional units as shown in Table 106. 1.

The Cape Storm Formation has been defined recently by Kerr (1975) as a limestone and dolomite unit of Silurian age that overlies the Allen Bay Formation and underlies either the Read Bay or Douro Formation on Ellesmere Island, northwestern Devon Island, Cornwallis Island and Griffith Island. It is considered by the writers to underlie an extensive area on northern Somerset Island. J. Savelle, University of Ottawa, recognized it on southern Somerset Island and in the regions of Cape Garry and Creswell Bay (pers. comm., 1975).

The unnamed unit between the Read Bay and Peel Sound formations was recognized earlier as a mappable unit by R. Thorsteinsson (pers. comm., 1975) and by M. Gibling of the University of Ottawa (pers. comm., 1975), and field work in 1975 substantiated these views. This unit deserves formational status and will be named when the revision of the stratigraphy and regional mapping are completed.

North-central Somerset Island

In north-central Somerset Island between Cape Anne and Garnier Bay, extensive areas of almost flat-lying Silurian strata are disrupted by several vertical fault zones. The Peel Sound Formation is absent but the unnamed unit is present sporadically, overlying the Read Bay Formation. Most of the area to the south, particularly the Cunningham River valley, contains rocks that are lithologically similar to the Read Bay Formation.

Northeastern Somerset Island

A well-exposed horizontal Silurian section at Port Leopold, consisting of grey to brown dolomite and limestone, with minor amounts of sandy carbonate, anhydrite, and sandstone, was named the Leopold Formation by Jones and Dixon (1975) who considered the formation to be of Pridolian age, and correlated it with the Read Bay Formation farther west. Mapping in 1975 showed that the flat-lying strata at Port Leopold could be traced northwestward to a point where they flex downward, dip northward, and are overlain stratigraphically by the Read Bay Formation (Fig. 106.1). The contact between the two formations occurs midway between Port Leopold and Cape Admiral M'Clintock.

Collections made near the basal contact of the Read Bay Formation yielded conodont faunas identified and dated by T.T. Uyeno. At a height of 9.2 m (30 ft.) below the top of the older formation (GSC loc. C-45647), *Ozarkodina confluens* (Branson and Mehl) γ morphotype of Klapper and Murphy (1974), and a form approaching ε morphotype of Klapper and Murphy (1974) were reported. At a height of 3 m (10 ft.) above the base of the Read Bay Formation (GSC loc. C-45648), *Ozarkodina* n. sp. B of Klapper and Murphy (1974) was found, in addition to the above two morphotypes of *O. confluens*. Both collections were dated by Uyeno (pers. comm., 1975) as Late Silurian, late middle Ludlovian to early Pridolian (*Polygnathoides siluricus* Zone to level of *Pelekysgnathus index* fauna).

East coast of Somerset Island

The east coast of Somerset Island is characterized by cliffs up to 365 m high, consisting largely of very low angle, westward-dipping rocks of the Leopold Formation. Inland from the coast, the stratigraphic contact between the Read Bay Formation and the underlying Leopold Formation is fairly distinct. The contact relationship between the two formations is one of rapid vertical transition, with interbedding occurring only in a limited interval, perhaps 15 m. This contact occurs in northern Somerset Island about 25 to 35 km inland from the east coast in the North Elwin River region and was traced southward from there to where it is exposed in the coastal cliffs near Two Rivers Bay. The Leopold Formation, because of its lithology, ostracoderm content, mappability and regional relationship to the Read Bay Formation, is considered to be correlative, for the most part, with the Cape Storm Formation of northwestern Somerset Island.

References

Blackadar, R.G.
1963: Somerset Island between Aston Bay and Creswell Bay; *in* Geology of the north-central part of the Arctic Archipelago, Northwest Territories (Operation Franklin); Geol. Surv. Can., Mem. 320, p. 143-150.

Blackadar, R.G. and Christie, R.L.
1963: Geological reconnaissance, Boothia Peninsula, and Somerset, King William, and Prince of Wales Islands, District of Franklin; Geol. Surv. Can., Paper 63-19, 15 p.

Dixon, J.
1974: Revised stratigraphy of the Hunting Formation (Proterozoic), Somerset Island, Northwest Territories; Can. J. Earth Sci., v. 11, p. 635-642.

Jones, B. and Dixon, O.A.
1975: The Leopold Formation: an Upper Silurian intertidal/supratidal carbonate succession on northeastern Somerset Island, Arctic Canada; Can. J. Earth Sci., v. 12, p. 395-411.

Kerr, J. Wm.
1975: Cape Storm Formation — a new Silurian unit in the Canadian Arctic; Bull. Can. Petrol. Geol., v. 23, p. 67-83.

1976: Geological investigations of Somerset Island (57C, F, G; 58A, B, F) and Boothia Peninsula; *in* Report of Activities, Part A, Geol. Surv. Can., Paper 76-1A, rept. 104.

Klapper, G. and Murphy, M.A.
1974: Silurian-Lower Devonian conodont sequence in Roberts Mountains Formation of central Nevada; Univ. Calif. Publ. in Geol. Sci., v. 111.

Thorsteinsson, R. and Tozer, E.T.
1963: Geology of northern Prince of Wales Island and northwestern Somerset Island; *in* Geology of the north-central part of the Arctic Archipelago, Northwest Territories (Operation Franklin); Geol. Surv. Can., Mem. 320, p. 117-129.

107. KIMBERLITES OF SOMERSET ISLAND, DISTRICT OF FRANKLIN

Project 740031

Roger H. Mitchell[1]

Introduction

The locations of all known kimberlites on Somerset Island are shown in Figure 107.1. The kimberlites form a belt trending approximately northeast-southwest. Dykes associated with the Batty Diatremes and the Ham Diatreme also are aligned on this trend. Intrusion appears to have been controlled by a northeast-southwest element of the fracture system in the Paleozoic sediments rather than by north-south trending structures of the Boothia Uplift, the major tectonic features of this region. The Jos Dyke, however, is aligned along the north south Boothia trend. No radiogenic ages have been obtained yet on the diatremes and associated dykes. They are known to intrude rocks as young as the Read Bay Formation and, therefore, must be post-Late Silurian in age. The individual diatremes are listed below by name and brief notes are given on each.

Elwin Bay Diatreme

This is a single pipe characterized by at least two phases of kimberlite, a marginal massive porphyritic variety with few xenoliths and a central core with ubiquitous ultrabasic xenoliths. These xenoliths are small, very fresh, and are in all stages of fragmentation; in fact, it is difficult to distinguish between true phenocrysts and xenocrysts derived by fragmentation of nodules in these rocks. The ultrabasic xenoliths are either slightly deformed granular spinel harzburgites or porphyroclastic tabular garnet lherzolites. Large megacrysts (1 cm in diameter) of garnets, possibly not related to the lherzolites, are common. Rare nucleated autoliths also are found.

Batty Diatremes

Near Batty Bay a complex group of pipes and dykes (Fig. 107.2) occur with many varieties of kimberlite. Some of these are:

1. Tunraq — This is a small pipe (300 m in diameter) containing at least three phases of kimberlite. The dominant phase is characterized by a coarse grained porphyritic type. Phenocrysts include altered olivines, phlogopite and magnesian ilmenite and pyrope. This coarse grained phase grades into massive kimberlite breccia, the xenoliths being Paleozoic carbonate rocks. Both types are cut by a sheet of micaceous kimberlite that is a highly altered, laminated, grey-green, fine grained rock rich in spinel and perovskite.

Figure 107.1. Index map showing the locations of kimberlites in Somerset Island.

2. Batty Pipe and Inugpasagsuk — These are extremely altered and weathered kimberlites, characterized by the presence of spinels showing complex mantles of the type found at Peuyuk (Mitchell, 1975).

3. Nanorluk — This kimberlite is similar to the Batty Pipe but is characterized by the presence of abundant, slightly deformed, garnet lherzolite and spinel harzburgite xenoliths, and rare olivine pyroxenite. All of the ultrabasic xenoliths are considerably altered and do not appear to be undergoing fragmentation of the type seen at Elwin Bay.

4. Amayersuk — This is an enlarged dyke of altered to very fresh, massive porphyritic kimberlite containing very few country rock xenoliths. It is characterized by fresh, essentially undeformed, garnet lherzolites and altered, slightly deformed, granular, spinel harzburgite xenoliths.

5. Arlu — This is an enlarged dyke of very fresh kimberlite characterized by abundant carbonate ocelli and abundant apatite.

Ham Diatreme

Here an enlarged dyke of highly altered kimberlite in the west grades into a northeast-southwest trending dyke of massive kimberlite characterized by carbonate ocelli.

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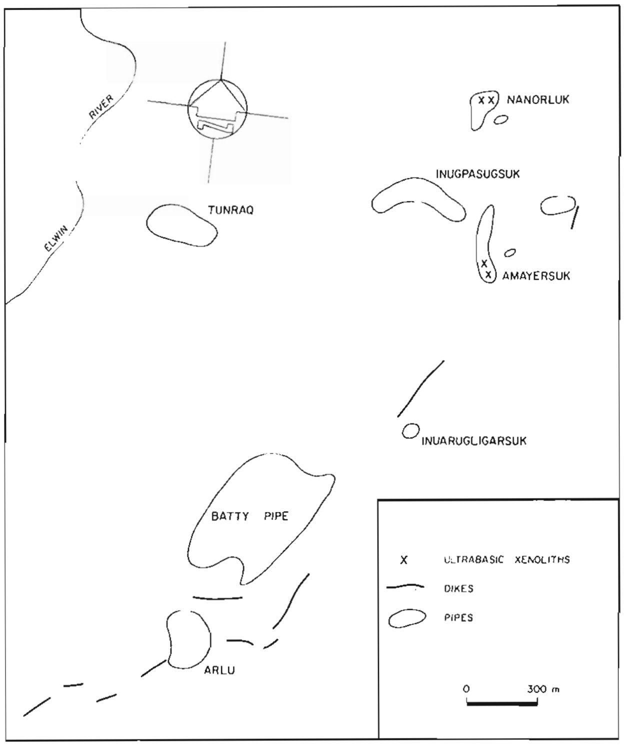


Figure 107.2. Map showing the complex group of kimberlite pipes and dykes near Batty Bay, Somerset Island.

Ouest and Nord Diatremes

These are highly altered porphyritic kimberlites.

Jos Diatreme

This is a north-south trending dyke, about 1 km long, composed of very fresh micaceous kimberlite. The rock consists of phenocrysts of olivine and phlogopite set in a coarse grained, dominantly carbonate groundmass. The rocks are very similar in texture to Peuyuk C (Mitchell, 1975) except that here mica is more abundant and the silicate ground mass is rarely present.

Peuyuk-Korvik-Selatiavak Diatremes

This system of enlarged dykes and diatremes has been described previously by Mitchell and Fritz (1963) and Mitchell (1975). The largest diatreme, Peuyuk, an enlarged fissure, is a composite intrusion. Three petrographically distinct phases are recognized on the basis of oxide mineralogy and the presence or absence of immiscible carbonate. These are: (A) spinel and perovskite with no zonations or reaction rims, no immiscible carbonate; (B) epitaxially zoned chrome spinels, magnetite and perovskite with reaction rims, no immiscible carbonate; and (C) oxides as in B, but with no magnetite and abundant immiscible carbonate. Selatiavak and Korvik each are composed of two small intrusions and are petrographically similar to Peuyuk C. All the kimberlites are characterized by a positive (500-2000γ) magnetic anomaly.

Creswell Diatreme

This pipe was not visited during the 1975 season.

In summary, the Somerset Island kimberlites exhibit the full spectrum of kimberlite types and ultrabasic inclusions except eclogite as are found in the classical examples of kimberlite magmatism in South Africa and Russia.

References

Mitchell, R.H.
1975: Geology, magnetic expression and structural control of the central Somerset Island kimberlites; Can. J. Earth Sci., v. 12, p. 757-764.

Mitchell, R.H. and Fritz, P.
1973: Kimberlite from Somerset Island, District of Franklin, Northwest Territories; Can. J. Earth Sci., v. 10, p. 384-393.

PRELIMINARY TEST RESULTS OF THE "BANCQES" THROUGH-ICE SUB-BOTTOM ACOUSTIC PROFILING SYSTEM AT TUKTOYAKTUK, NORTHWEST TERRITORIES

D.D. Caulfield[1], A. Liron[1], C.F.M. Lewis[2] and J.A. Hunter[3]

In April 1975, a joint program was undertaken by Banister Technical Services and the Geological Survey of Canada, Department of Energy, Mines and Resources, to field test the BANCQES through-ice sub-bottom acoustic profiling system in the shallow waters of the Mackenzie Delta near Tuktoyaktuk, N.W.T. Test data was acquired at four sites where cores were previously taken and analyzed by the Geological Survey of Canada. Multiple samples were acquired in three different frequency regions.

The identification of soil properties in both the marine and land environment is an important part of any construction or civil engineering project. Identification by remote sensors becomes of greater importance in an environment such as the Arctic for at least two major reasons:

1. The cost of physical coring is very high in such a harsh environment.
2. From an ecological point of view, the remote sensing may be preferable to other methods that cause disturbances.

An initial large scale application of the BANCQES System was done in 1974 for Polar Gas Ltd. across Byam Channel where shallow sediments were predicted for a pipeline crossing survey. However, no work has been done in an area of shallow water, where calibration cores were available at the same time to check the capabilities of the system for a penetration deeper than a few metres.

Field Procedures and Data Acquisition

The test locations are shown in Figure 108.1. A Bell 205 helicopter flew the instrument shack along with a survival tent and the back-up equipment to each previously surveyed location. At each site an "X" pattern of transducer stations, 10 feet apart was surveyed as shown in Figure 108.2. Data was acquired in three major frequency bands, 3.5 Khz, 7.0 Khz, and 12.0 Khz by the BANCQES field system. This system makes use of two unique devices that improve the system performance over conventional shallow seismic equipment, namely:

1. The BANCQES Field Amplified – for improvement of the signal-to-noise ratio.
2. The BANCQES Digital Signal Expander – to expand the signal five-fold along the time base making possible the field interpretation of graphic records taken in shallow water.

[1]Banister Technical Servies Ltd.

[2]Terrain Sciences Division.

[3]Resource Geophysics and Geochemistry Division.

Figure 108.3 is a schematic diagram of the field data acquisition system.

The objective in this program was to acquire data at each of the core sites at a frequency of 3.5 Khz for maximum penetration of the signal in the sub-bottom layers. The 7.0 Khz data was acquired to provide higher resolution of the upper layers. The 12.0 Khz data was needed to determine the water depth. The water was too shallow for the lower frequency signals to detect and separate on the records the true bottom. The depths encountered were on the average in the range

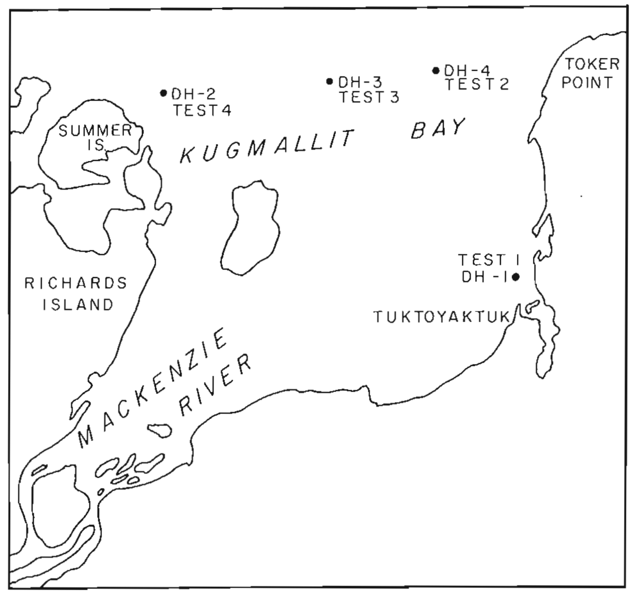


Figure 108.1.

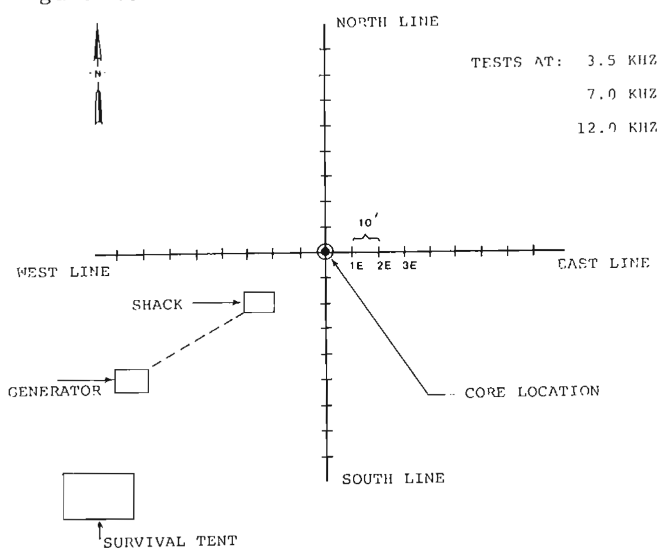


Figure 108.2. Detailed tests at each test site showing "X" patterning.

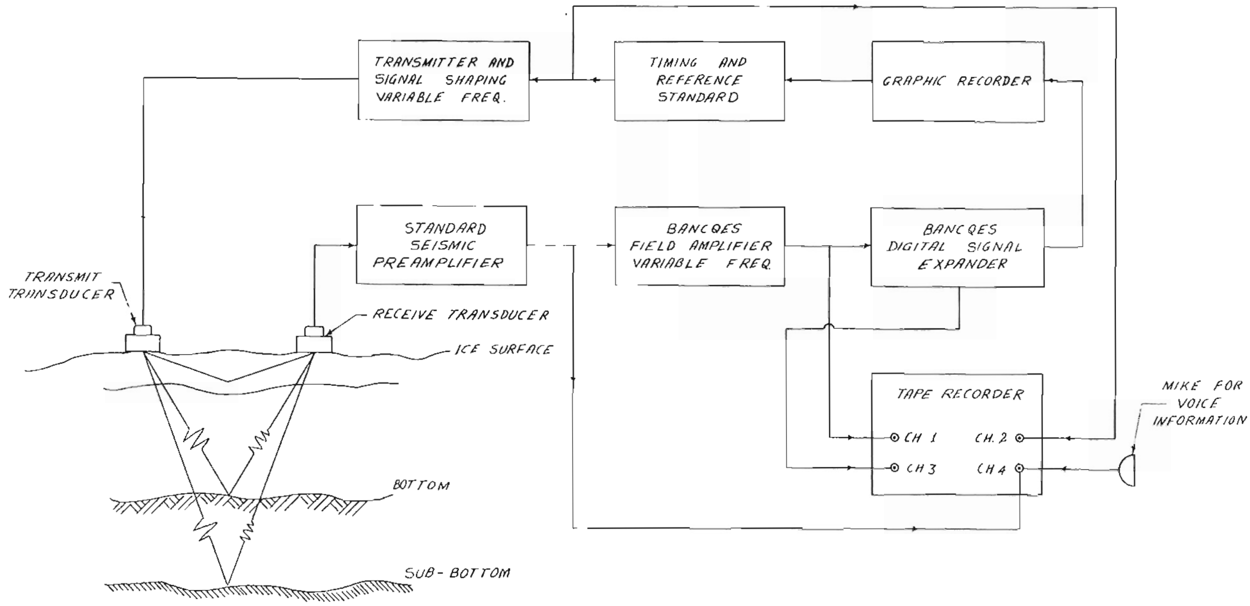
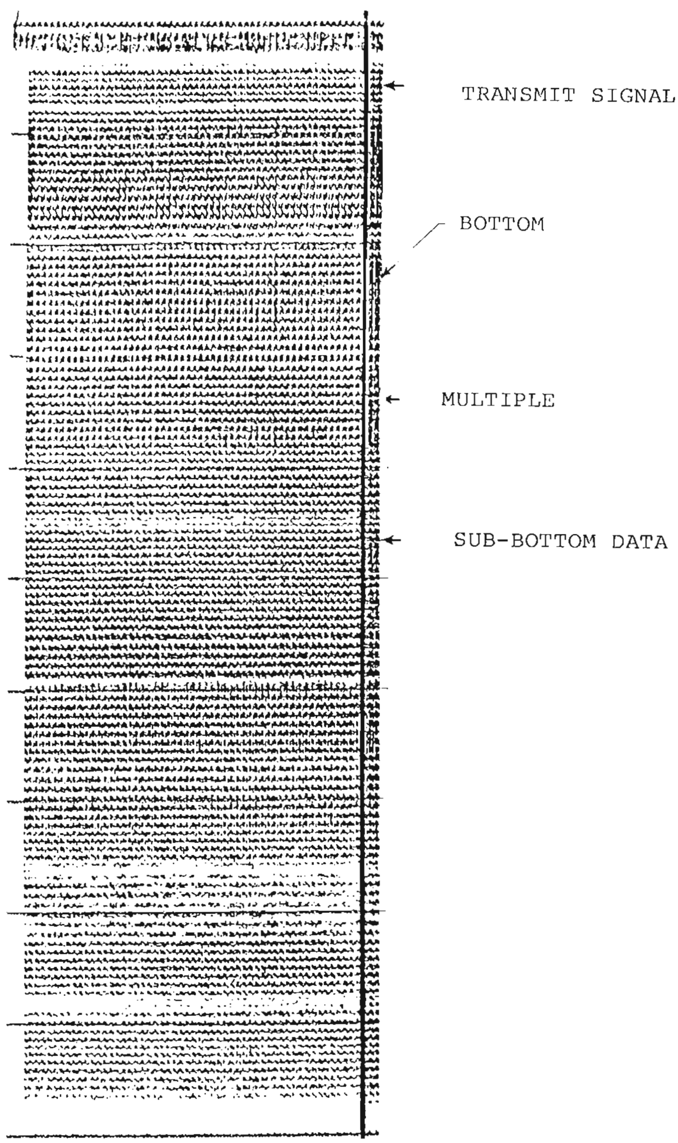


Figure 108. 3 (above). Schematic field data acquisition.

Figure 108. 4 (right). Typical 3. 5 Khz field record.



of 4 to 6 m under ice 2 m thick. All quantitative soundings were made through-the-ice.

Figure 108. 4 illustrates a typical field graphic record of the 3. 5 Khz signal. This record exhibits the existing reverberation. Figure 108. 5 illustrates the 12. 0 Khz field graphic record. Here, it is easier to determine the bottom depth. Note that as one goes higher in frequency, the reverberation diminishes, but then the penetration into the sub-bottom is substantially reduced. Figure 108. 6 illustrates the advantages of utilizing the BANCQES Digital Expander in the field, especially in very shallow water.

In all previous experiments with the BANCQES system the entire system, including the generator, was assembled in one housing. Noise and cabling problems were sorted out before the data acquisition started. At Tuktoyaktuk the complete field system was tested in the Field Survey shack while in the shop/warehouse both the shack and warehouse had metal floors which provided a good electrical ground.

While working on the ice, the generator was attached to the house only through long cables, and because the house was not standing directly on the snow it lacked a good ground. Consequently, the power supply from the transmitter contained transients which were coupled into the preamplifier and the tape recorder causing the first milliseconds of data to be distorted. Normally, this would be of no concern, except where the water depth is shallow. Part of the data at Location no. 1 was destroyed because of this transient. It was only at Location no. 2 that it was realized that the distortion observed was electrically induced and not a geophysical phenomena. The first part of the records remained practically unchanged even when the transducers were moved to an adjacent station with a different water depth. Field calibration detected this distortion and a partial remedy was found. The grounding problem can reach severe proportions under certain conditions in the Arctic. For example, on another occasion it was so dry, that when the wind blew across the transmission cables charges were built up causing distortion in the records. Again, when following the regular calibration procedure this effect was noticed. A modification made in the transmission cable impedance isolated the problem. Extreme care has to be taken in selecting a sufficiently powerful generator and in ensuring proper grounding of all connections, otherwise a serious problem may arise when operating in very shallow water.

Calibration is probably the most critical factor in the process of obtaining analytical information on sub-bottom properties. This step ensures that no changes in the characteristics of the system have occurred during the field operations. Calibration plots for the critical components are generated and stored with the acquired field data for future reference. Any changes from the normal operating conditions are inserted into the computer to be accounted for when the data are processed.

The use of analog magnetic tapes for the bulk storage of raw data is one of BANCQES unique features. The selected data at a given location can be digitized at rates from one quarter real time to real time, depending on the accuracy required and the maximum signal length.

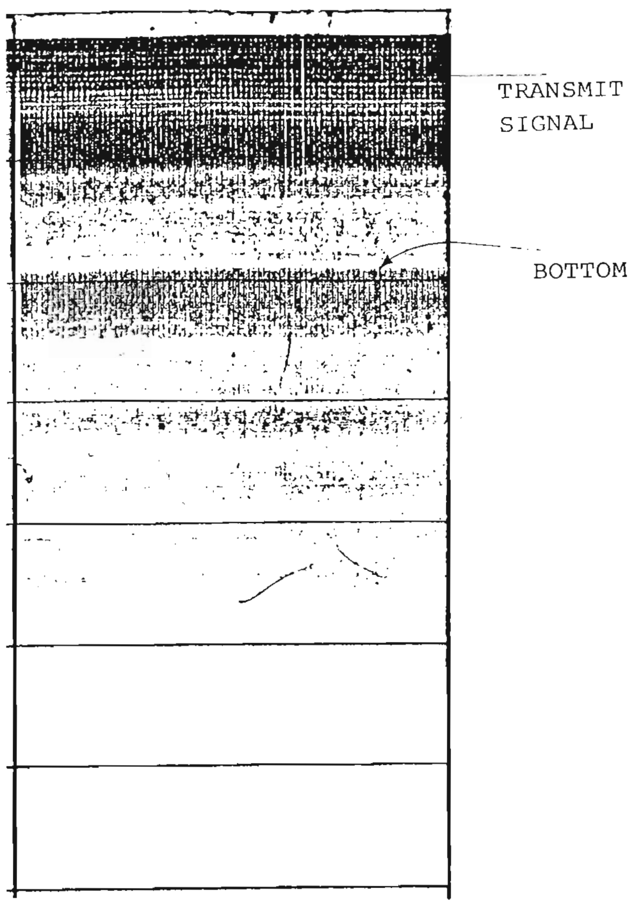


Figure 108. 5. Typical 12. 0 Khz field record.

Once the signal is digitized and while still in the high speed computer, the BANCQES processing is applied to the signal. This enables it to quantitatively relate calculated values of acoustic impedance to bottom properties. The results of this processing is passed through a digital editing console to a mass storage disc. All the information required to regenerate the sub-bottom structure can be retrieved at any time from this disc. In addition, information related to the calibration procedures is also stored. The complete process, from the data input to the computer storage for prediction takes 10 minutes per station. When requested, the data that has been entered into the disc memory, can be quickly retrieved to be printed or plotted in the desired format. The data can also be examined for calibration factors, and stability of results.

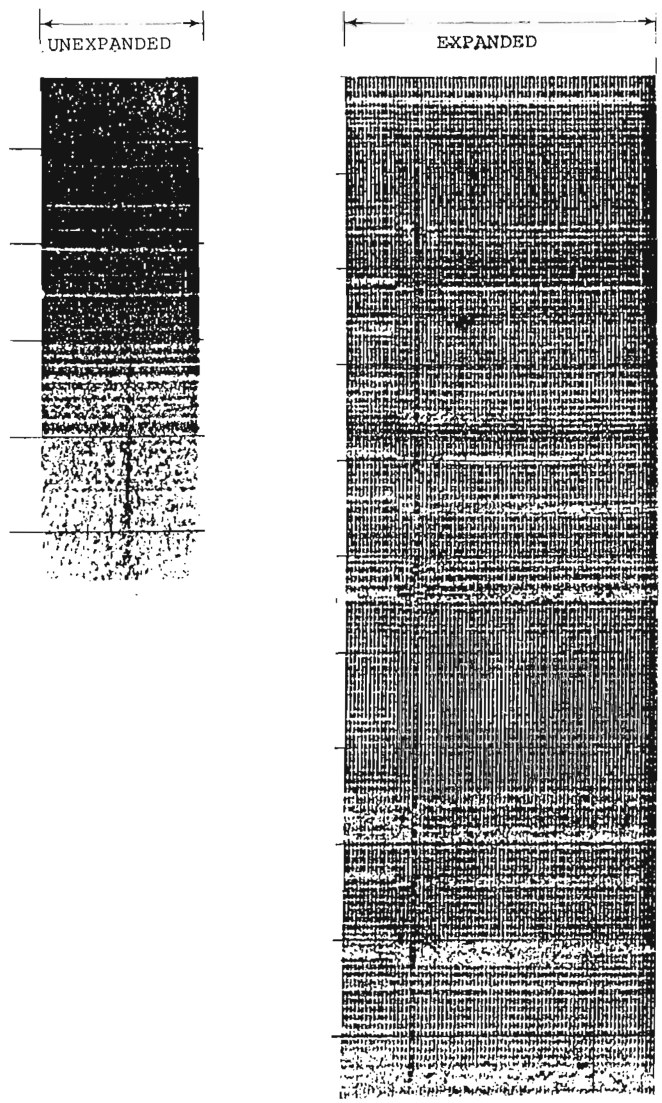


Figure 108. 6. Comparison of standard shallow field record and the field record generated by BANCQES digital expander.

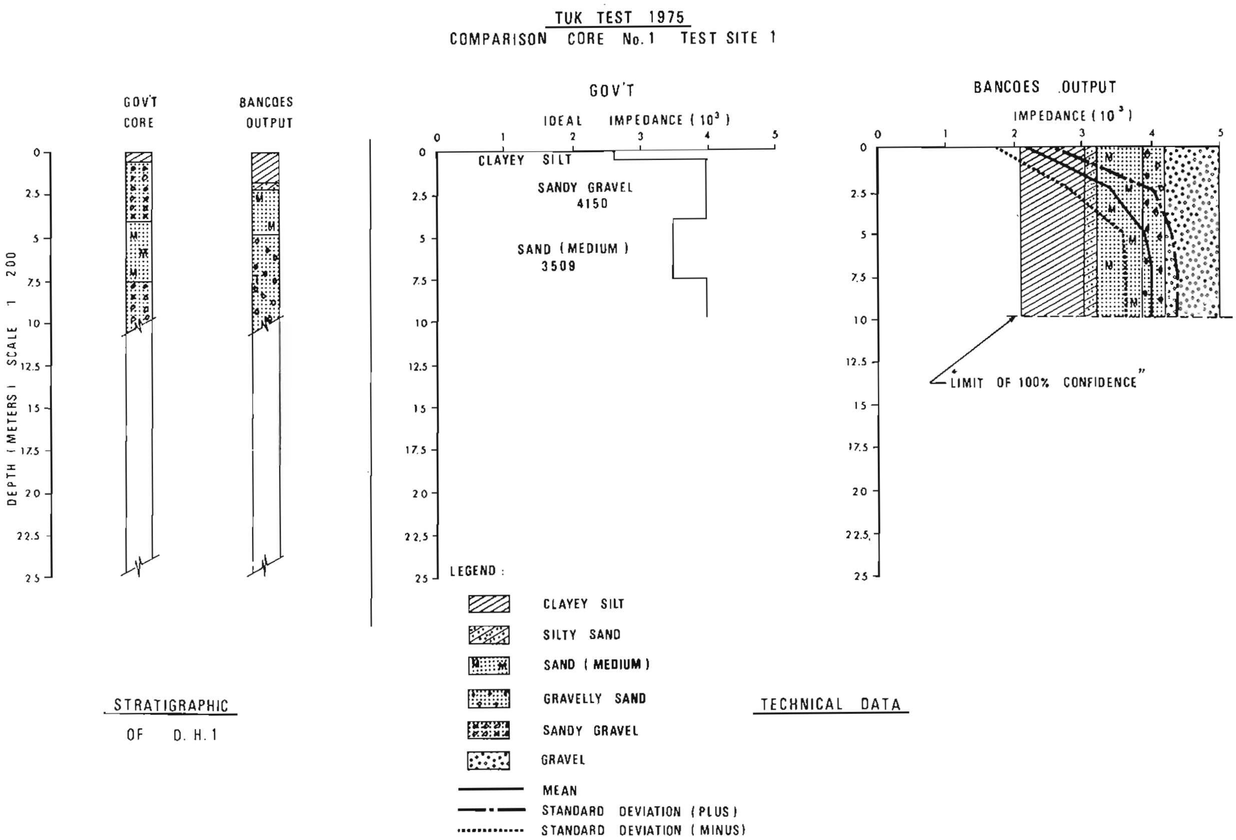


Figure 108. 7.

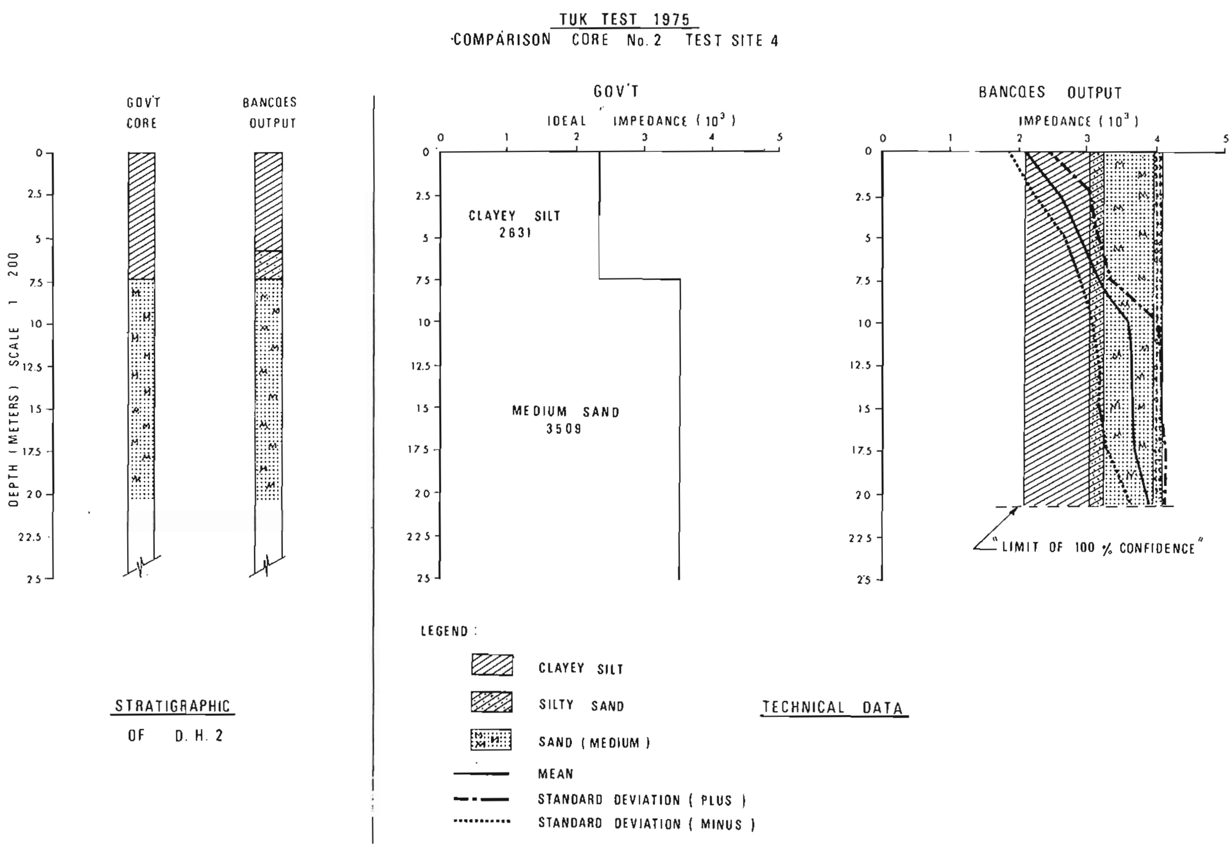


Figure 108. 8.

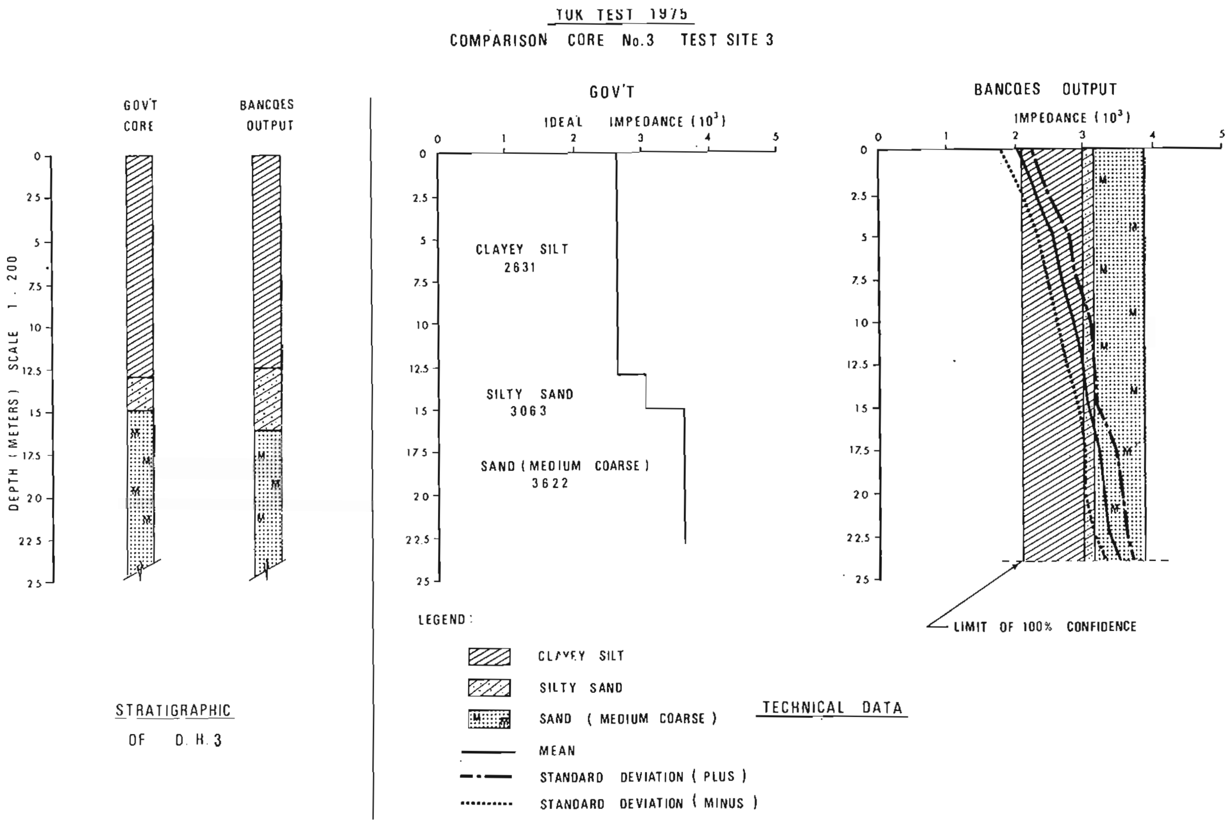


Figure 108. 9.

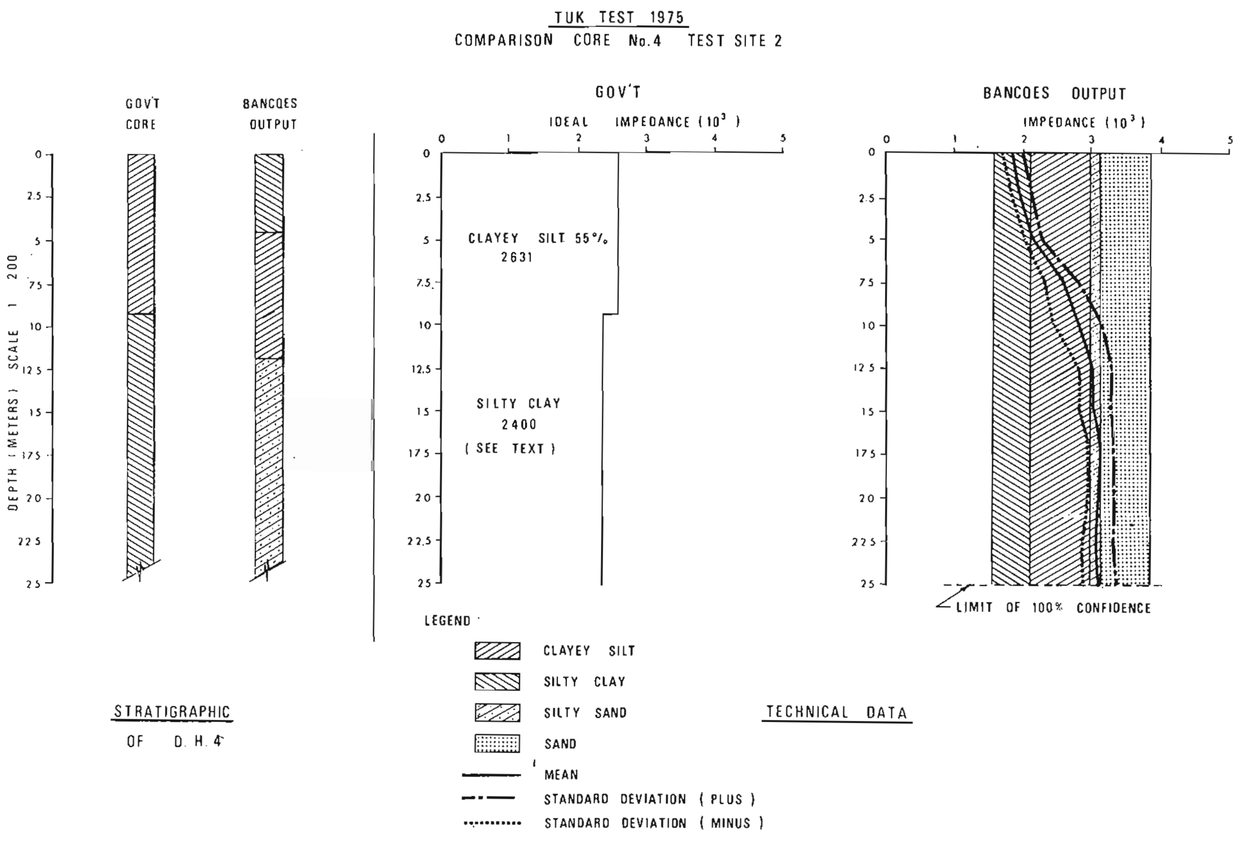


Figure 108. 10.

Discussion of Results

Acknowledging the problem of multiple reflections and reverberation in shallow waters near Tuktoyaktuk the question was to what accuracy could BANCQES predict the sub-bottom structure without deconvolution and dereverberation programs being applied? The decision to test the BANCQES predictions was made on the basis that the top layer was of clay-silt nature reducing the reverberation effects.

Preliminary data reduction was done for a number of observations at each core location. These observations are as follows:

1. To prove that the data observed were independent of the coupling to the ice and its thickness, data were recorded from a number of different transducers spaced 10 feet apart.
2. At most of these locations two different types of signals were transmitted. These were obtained by transmitting:
   a) A burst of 3.5 Khz 0.25 millisecond long;
   b) A burst of 3.5 Khz 0.50 millisecond long.

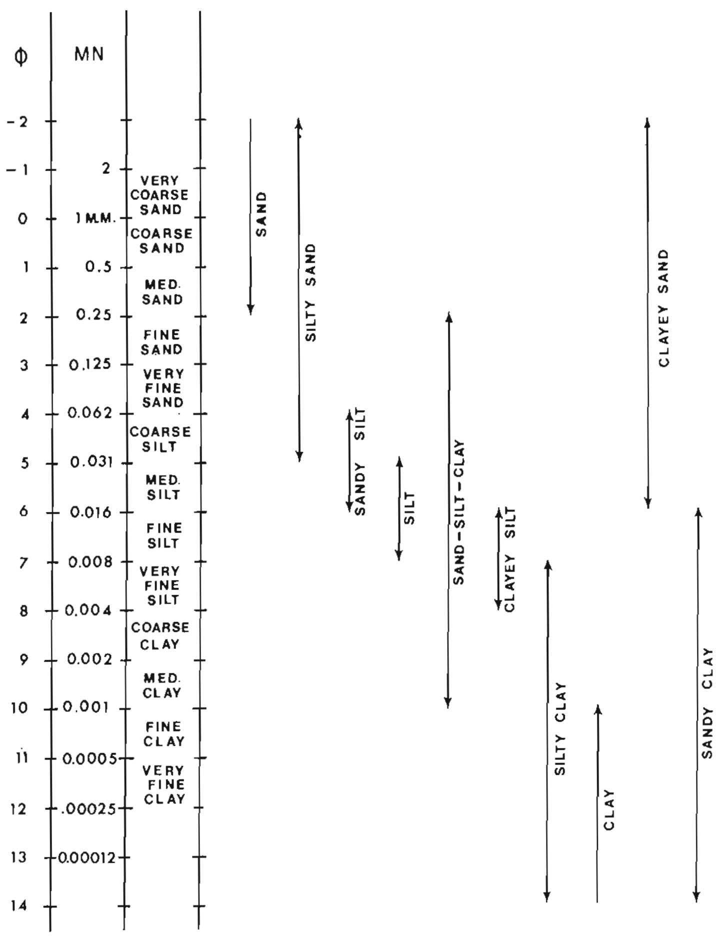


Figure 108.11. General range of median grain sizes for sediments classified according to a sand-silt-clay system.

For the predictions to be considered accurate, the same results should be obtained regardless of the physical location, the coupling procedures, and the transmitted signal.

The analysis of the cores by the Geological Survey provided a classification of the sediments present in the sub-bottom layers. BANCQES predictions classified the layers according to their acoustic impedance and then calculated a list of interrelated engineering parameters. Since the layers are defined by BANCQES within certain limits of the acoustic impedance value, it was important to establish the acoustic impedance of the sediments identified by the core analysis. This provided a common ground for comparison between the BANCQES predictions and the actual core data.

In the literature, information is available in the form of tables (Hamilton *et al.*, 1956) which supply the characteristic value of acoustic impedance for saturated sediments. Table 108.1 shows acoustic impedance versus soil classification for the sediments present in the core analysis. Direct measurements of porosity from selected samples of core made possible the cross checking of the impedance value attributed to some of these sediments in Table 108.1. Based on these values impedance was plotted versus depth and compared with the BANCQES system predictions for the impedance as a function of depth in the sub-bottom.

Figures 108.7 to 108.10 show the results of the comparison of the BANCQES predicted soil properties versus the core measured properties. The right side

Table 108.1

ACOUSTIC IMPEDANCE VERSUS SOIL CLASSIFICATION

DESCRIPTION	ACOUSTIC IMPEDANCE $\left(\frac{\times 10^2}{cm^2}\ \frac{g}{s}\right)$
ICE	8000
WATER	1528
SILTY CLAY	2157
CLAYEY SILT	2631
SILTY SAND	3063
VERY FINE SAND	3264
FINE SAND	3443
MEDIUM SAND	3509
COARSE SAND	3735
GRAVELLY SAND	3900
SANDY GRAVEL	4150
GRAVEL	4200-5000
TILL	5000-8000

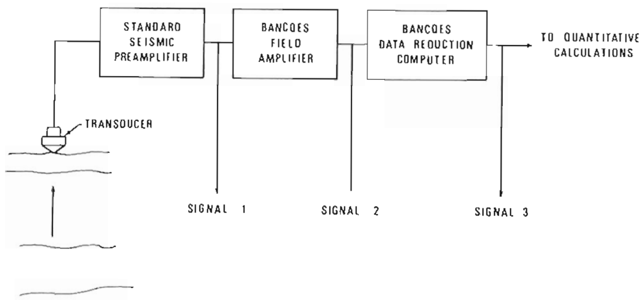


Figure 108.12(a). Block diagram for the signal-to-noise examination.

of each of these figures shows the literature values of impedance for the core plus the observed BANCQES impedance as a function of depth. The values of the acoustic impedance from each independent observation as a function of depth at each core location was used to derive the statistical mean, standard deviation and mean standard deviation. Considering all of the possible variations in acoustic coupling, ice thickness, and the lack of dereverberation analysis this standard deviation is quite small. Four important facotors are to be considered in reviewing this data:

1. A depth sampling of 2.5 m interval was utilized in this first analysis. Hence, interfaces between the sample intervals are in effect averaged.
2. Since there was a distortion factor caused by the equipment for the initial part of the signal return, the predicted impedances of the first 5 m would appear less than they should have been.
3. Predictions at Core no. 1 were produced from data noisier than the data from the three other core sites.
4. The observed porosity in the Geological Survey cores was used to adjust the impedance values of the soil type to the nearest possible core impedance. A particular problem arises with core no. 2. Here the core states that the porosity in the silty clay region is in the range of 50-60 per cent. Previous workers define this porosity for soils nearer to sandy silt than to silty clays as is shown on the core log. More work must be done in the finer classification of sand-clay-silt regions.

It is important to note that ambiguity still exists in the way soils are identified. Consideration must be given to the ambiguity. For example, engineering judgment must be made for the sediment types that overlap in their classification (Breslau, 1964). Figure 108.11 shows the typical cases of overlapping of the sediment classification. This area is still open for continued research and definitions should be established for specific applications.

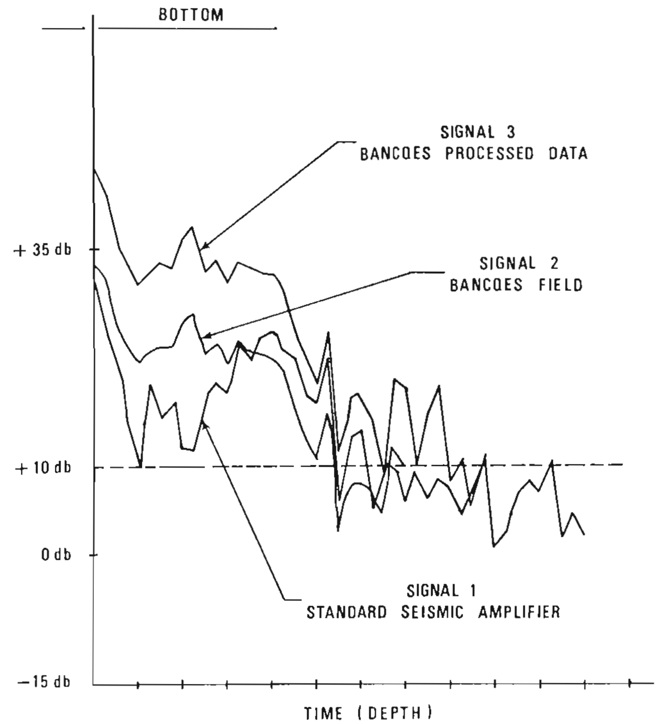


Figure 108.12(b). Illustration of improvement of signal/noise through the BANCQES system.

A Discussion of Signal-to-Noise

On each of Figure 108.7 to 108.10 appears the statement "limit of 100 per cent confidence". This is derived from the signal to-noise ratio analysis of the signal at this observation. This is the limit at which the noise level reduces the confidence for quantitative results.

The investigation of the ways in which the signal-to-noise ratio may be degraded in passing along a transmission path, and the techniques which may be used to improve the signal-to-noise ratio in detection at the receiver, have received a great deal of attention since the early days of radar. The fact that the disturbances to which a signal is subject are generally of a random nature is the basis for statistical analysis in modern signal theory. This science is well described in Helstrom (1960), and shows that the signal-to-noise ratio provides an excellent measure for the information content of the transmitted signal when complemented by its statistical analysis. The BANCQES system makes use of this knowledge in optimizing its quantitative predictions.

The theory (Sams, 1973) demonstrates that below the S/N ratio of 10 db, the probability of false information content in the signal increases rapidly. The uncertainty of the predictions, when operating under this limit has been confirmed.

To keep a high level of confidence in the BANCQES prediction, the "limit of 100 per cent confidence" was set as the level of 10 db for the S/N ratio. No quantitative predictions are made below this limit since it would then require an interpretation associated with a large risk factor.

The data from Test Site 2 Core 4, Station 2F has been chosen as an example of the actual data characteristics. Figure 108.12(a) is a symbolic block diagram illustrating the stages at which the signal-to-noise ratio is examined in the system. This diagram shows the signal flow from the receivers-transducers (hydrophone) through the seismic preamplified, to the BANCQES field amplifier, and finally through the signal processing done at the data reduction facilities. The three signals present in Figure 108.12(b) are all processed on exactly the same band-width and undergo the same sampling procedures. The values of the signal-to-noise ratio are plotted against the depth of penetration (time) into the sub-bottom. It is important to note that the 10 bd reference is the point where one must be concerned with the theoretical limit of the analytical predictions. Beyond this limit a higher level of processing is required.

Summary

Signal processing through the BANCQES system was performed for measurements taken at four sites in the Mackenzie Delta near Tuktoyaktuk. Measurements were made through 2 m of sea-ice and 2 to 4 m of water. Interpreted acoustic impedances were related to grain size and porosity of sediments to depths of 25 m below sea bottom. The method shows promise for determining engineering and sediment properties of sub sea-bottom materials in ice-covered areas without the necessity of a detailed coring program.

References

Breslau, L. R.
1964: Sound reflection from the sea floor and its geological significance; M.I.T.

Hamilton, E.L., Shumway, G.S., Menard, H.W., and Shipek, J.
1956: Acoustic and other physical properties of shallow water sediments: J. Acoustic Soc. Am., v. 28, no. 1.

Helstrom, Carl W.
1960: Statistical theory of signal detection; Pergamon Press.

H.W. Sams and Co., ITT, Fifth Edition
1973: Reference data for radio engineers, p. 38-12, 38-13).

TUNDRA FIRES IN THE DISTRICT OF KEEWATIN

Project 730013

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Terrain Sciences Division

Information on tundra fire occurrence and ecological impact in the Western Arctic recently has been accumulating (Cody, 1964, 1965; Heginbottom, 1972; Mackay, 1970; Wein, 1974, 1975; Wein and Bliss, 1973), but only a few observations have been made in the Eastern Arctic (Cochrane and Rowe, 1969). Since there is a great difference in terrain between Western and Eastern Arctic, caution should be used in extrapolating detailed information from one area to the other.

Few fires had been recorded for the Eastern Arctic until 1973 when 14 tundra fires were located (Shilts and Boydell, 1974; Shilts, 1975). These papers discussed the general frequency and extent of tundra fires, vegetation recovery, slope stability, and drainage pattern changes.

The objectives of the present work were 1) to quantify some of the past observations for future compilation and publication and 2) to make preliminary qualitative comparisons with tundra fires studied in the Western Arctic.

Study area and methods

Three tundra areas that burned in 1973 and one tundra area that burned in 1970 were visited in July 1975 (Table 109.1). The ideas presented in this report should be considered in the context of the short examination time. Observations were made on the extent of the burn, the types of plant communities burned, and the depth of peat oxidized. Quantitative data were taken on burned and unburned communities to determine which species recovered and how much plant biomass had accumulated to date.

For each burned versus unburned comparison, 20 randomly located quadrats (50 x 100 cm) were photographed with colour positive film for plant cover analysis in the laboratory. Biomass was gathered from five burned and five unburned plots to determine the recovery rate. Particular attention was paid to the amount of surface organic soil oxidized and the relation of this to species recovery.

Results and discussion

Fire dynamics

Almost all vegetation types burned, even though in some vegetation types available fuel was very low (Shilts, 1975). The fire commonly was stopped by drainage areas, streams, or lakes, but in a few cases the fire was extinguished by rain as it burned through a vegetation type. Under favourable meteorological conditions, the vegetation quickly carried the fire, even in wet meadows and lichen-dominated beach ridges or eskers. In addition, the surface peat layers of the more mesic vegetation types (e. g., cotton grass tussocks and dwarf birch heath) continued to burn under less than favourable conditions. Fires that burned for more than one day in the same place were peat fires.

Mudboils deserve special attention because of their prominence in the landscape and because they have a discontinuous fuel. The central part of the mudboil, which is mineral soil with only scattered plants, does not burn. Generally the inter-mudboil area burns completely, but in moist areas only the vegetation on the ridges of the mudboil burned. The ridge vegetation and organic matter include or rest directly on rocks that have been pushed or moved to the edges of the mudboils through frost action and diapirism of mud near the boil centre (Shilts, 1974). When the mudboil ridge vegetation is burned the rocks are exposed, forming stone nets (Fig. 109.1). The exposed rocks

Table 109.1

Characteristics of the four areas of burned tundra examined during July 1975

Fire Name	Location	Dates	Size	Cause	Other Comments
Carr Lake	62°01'N, 95°30'W	August 2-19, 1973	5 mi^2	lightning	
Kazan River	64°00'N, 95°30'W	July 30, 1973	4 mi^2	hunter	burned most of summer according to local residents
Ketyet River	64°24'N, 95°05'W	July 29, 1973	3-4 mi^2		observed burning on ERTS imagery
Pitz Lake	63°54'N, 96°42'W	1970	5x50 m		

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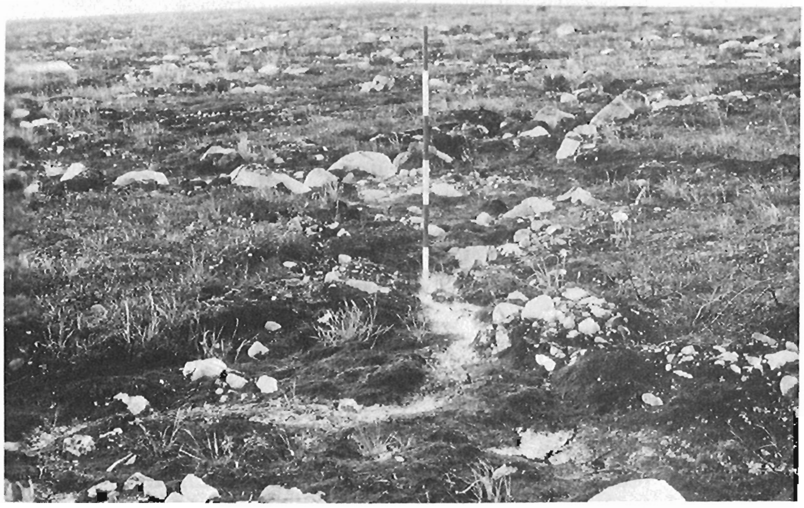

Figure 109. 1

After a fire in mudboil terrain, the mudboil centres remain unburned. Surface organic materials are oxidized and rocks in the newly created drainageway are exposed.

Figure 109. 2

Unburned (A) and burned (B) wet cotton-grass dominated area that shows strong recovery after two growing seasons.

Figure 109. 3

Unburned (A) and burned (B) lichen-heath area. The surface organic mat that contained all of the living plants has been oxidized. The rocky burned surface contains little fine soil and shows almost no recolonization after two growing seasons.

may be used as a key to finding other burned areas in the future, but mudboils in extremely stony till or shallow till on bedrock can form similar features without the aid of fire.

Vegetation Recovery

Except for lichens, most vegetation showed recovery in recently burned tundra areas (Figs. 109. 2, 109. 3 and 109. 4). Although quantitative data are not yet available, some generalizations are readily apparent.

Recovery in wet areas: Wet Carex-Eriophorum meadows show the greatest recovery; sedges are as abundant on burned areas as on the unburned areas. Only litter accumulation is lacking on the burned areas.

When cotton-grass (Eriophorum vaginatum) tussock vegetation with associated shrubs such as *Betula* are burned, the associated shrubs recover slowly so that cotton grass appears to be even more prevalent in the burned areas.

In these moist areas of the landscape, wet peat seems to have protected the underground plant parts, facilitating immediate regeneration.

Recovery in mesic areas: Areas with mesic vegetation, such as a high moss and shrub cover, have a different species composition after burning. An almost complete moss cover of *Ceratodon purpureus* is found on these areas. The grass *Calamagrostis canadensis* is the next most abundant species. The liverwort *Marchantia polymorpha* commonly was found in association with the moss. Shrubs such as *Betula glandulosa, Vaccinium uliginosum, Vaccinium vitis ideae*, and *Ledum palustre decumbens* are recovering in inverse proportion to the intensity of the burn. Where the peat layer, containing roots and stems, was completely oxidized, regrowth was absent or very sparse.

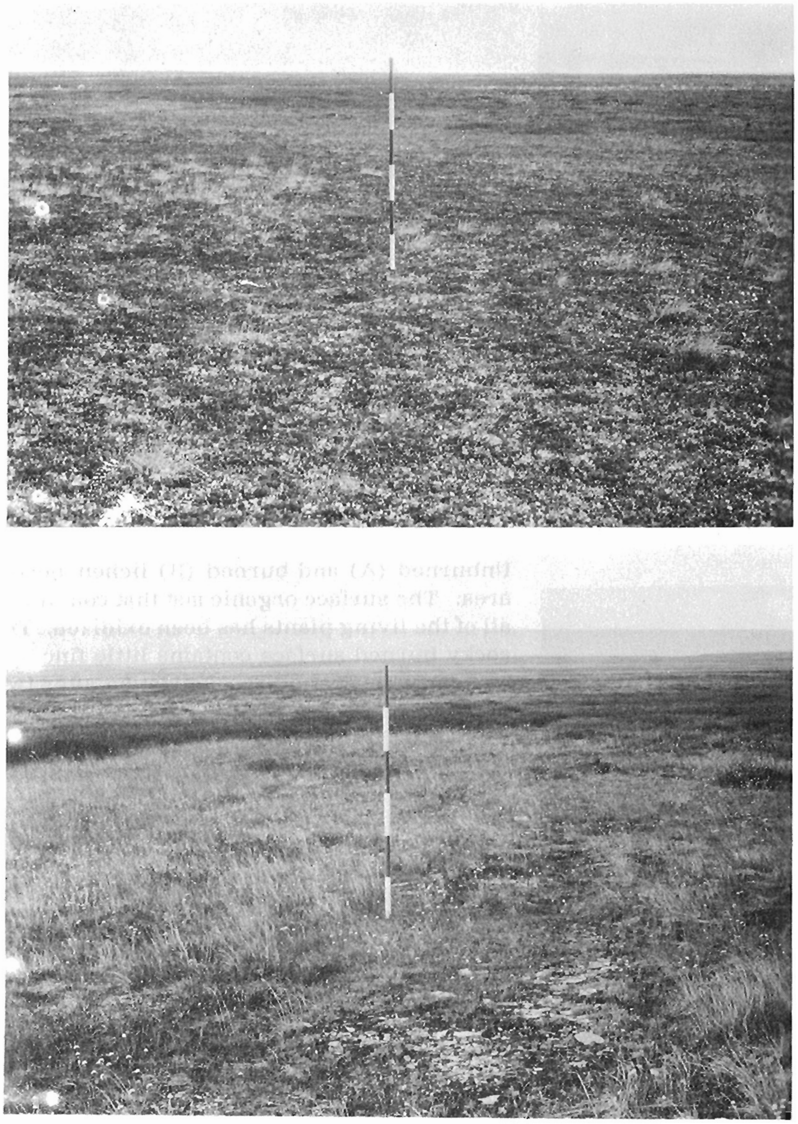

Figure 109. 4

Unburned (A) and burned (B) lichen-heath beach ridge. After five growing seasons a great variety of grass species covers the gravel soils.

Recovery in xeric areas: Lichens that dominated the vegetation on eskers and beach ridges have been replaced by scattered plants of the grass *Hierochloe alpina*. From observations of the area burned in 1970 it was obvious that *Hierochloe alpina* and other grasses have become even more important on older burns.

The most striking difference between the species recolonizing Keewatin tundra fire areas and those in the Western Canadian Arctic is that fireweed (*Epilobium angustifolium*), *Senecio congestus*, *Arctagrostis latifolia* and *Equisetum* spp. appear to be much more prevalent in the Western Arctic.

Effect of fire on soil thermal conditions

Much of the landscape in this area has a layer of organic material on the surface which burns during a fire. Changes in active layer depth following fire are extremely difficult to measure because of the high rock content of the soil. In mudboil areas, where mudboil centres are bare and rocks are exposed around the mudboil, loss of some organic cover probably has little effect on final active layer depth. Where mudboils have few stones and prominent vegetation ridges, the inter-mudboil areas will have an increased active layer thickness following the burning of the more uniformly deep organic matter layer.

Because of the interconnected nature of the vegetation ridges surrounding mudboils, an integrated drainage system is established in the stone-filled, moat-like depressions that are left after fire has removed the vegetation mass. This results in more rapid runoff with concomitant erosion of ash and fine mineral soil from the troughs, accentuating their depth below the mudboil "islands" and probably depressing the permafrost table by several centimetres.

Other vegetation types lying in low, flat areas of alluvial, marine, or lacustrine sedimentation have uniform organic matter that supports predominantly sedges with *Betula glandulosa* occurring in dryer types. In these areas water is normally close to the surface and the fire has removed only a small amount of litter and organic matter; the increase of active layer depth following fire is probably small.

There is no evidence of massive ground ice in the area and only small amounts of segregated ice are found at the depth of maximum active layer. As a result, the change in thermal conditions after fire does not cause obvious subsidence as in the Western Arctic, at least in the areas studied.

Other potential effects of these tundra fires

A number of questions are raised that could point the way to useful research investigations. Some hypotheses are supported by few data or observations at the present time.

It has been pointed out that there could be a reactivation or increased action of mudboil diapirism following fire. This could lead to increased sediment loss from the area because the organic matter and vegetation mat is no longer available to filter out the soil particles (Shilts, 1975). It also should be mentioned that if there has been a decrease in evapotranspiration from mudboil areas following fire, the soils might more readily reach the liquid limit and become active.

Mineral cycling changes (not only of the essential growth nutrients but also of the heavy metals) are also suspected following fire. More sediment is lost to streams and lakes after fire and since warmer soils are conducive to microbial activity and further organic matter oxidation, more mineral elements will become available. It is not known at the present time if these nutrient levels are sufficiently great to raise the productivity of nutrient-poor arctic lakes.

Biogeochemical research for heavy metals on plants recovering after fire may be productive if the nutrients released by fire are more readily accumulated. A word of caution should be expressed as well because little information is available for high temperature effects on organic matter-mineral complexes.

References

Cochrane, F.R. and Rowe, J.S.
1969: Fire in the tundra at Rankin Inlet, N.W.T.; Proc. Ann. Tall Timbers Fire Econl. Conf., v. 9, p. 61-74.

Cody, W.J.
1964: Reindeer range survey 1957 and 1963; Plant Res. Inst., Can. Dept. Agric., Ottawa, 15 p.

Cody, W.J. (cont'd)
1965: Plants of the Mackenzie River delta and reindeer grazing preserve; Plant Res. Inst., Can. Dept. Agric., Ottawa, 56 p.

Heginbottom, J.A.
1972: Some effects of a forest fire on the permafrost active layer at Inuvik, N.W.T.; Proc. Seminar on Permafrost Active Layer, N.R.C. Tec., Mem. 103, p. 31-36.

Mackay, J.R.
1970: Disturbances to the tundra and forest tundra environment of the western Arctic; Can. Geotech. J., v. 7, p. 420-432.

Shilts, W.W.
1974: Relationship of patterned ground to sediment properties, zone of continuous permafrost, Keewatin, N.W.T.; Geol. Soc. Am., Abst., v. 6, p. 545.

1975: Tundra fires, southeast District of Keewatin; *in* Report of Activities, Part B, Geol. Surv. Can., Paper 75-1B, p. 187-195.

Shilts, W.W. and Boydell, A.N.
1974: Terrain mapping in the Churchill-Chesterfield Inlet corridor, District of Keewatin; *in* Report of Activities, Part A, Geol. Surv. Can., Paper 74-1A, p. 253-256.

Wein, R.W.
1974: Recovery of Arctic vegetation after fire; Environmental-Social Committee, Northern Pipelines, Rept. 74-6, 63 p.

1975: Vegetation recovery in Arctic tundra and forest-tundra after fire; Arctic Land Use Res. Pro., D.I.N.A., ALUR 74-75-62, 115 p.

Wein, R.W. and Bliss, L.C.
1973: Changes in Arctic *Eriphorum* tussock communities following fire; Ecology, v. 54, p. 845-852.

A STUDY OF STRATIFORM COPPER DEPOSITS IN MIDDLE CARBONIFEROUS STRATA OF NEW BRUNSWICK AND NOVA SCOTIA

E. M. R. Research Agreement 1135-D13-4-250/75

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A program to investigate the occurrence of copper mineralization at the Pennsylvanian-Upper Mississippian contact in the Maritime Provinces was initiated in 1974 (Brown, 1975). A second visit to the field was made during the past summer as a follow-up to this project. Particular attention was given this year to sedimentological features which might explain the environment of host-rock deposition and post-depositional changes which may have had contemporary or subsequent influences on mineralization. Discussions and field examinations with Dr. Peter McNab, post-doctorate fellow at McMaster University, were especially helpful in this investigation.

Regional sedimentological aspects of the Carboniferous strata in the Maritime regions have received considerable attention in the recent literature (e.g., Van de Poll, 1972), but the occurrence and possible metallogenic significance of some smaller features had not been appreciated by the author until now. For example, the transition from Mississippian to Pennsylvanian beds, which marks a distinctive change from native metal to sulphide facies mineralization, is not characterized by a unique contact of the brick-red Maringouin Formation with overlying buff to grey Shepody beds. Instead, a typical stratigraphic column would show several recurrences of Maringouin-type red beds alternating with basal Shepody-type beds. Copper sulphide mineralization, where present, seems most common to the lowest stratigraphic occurrence of sulphide rich Shepody units, with lesser amounts of copper at successively higher repetitions of the Maringouin-Shepody contact.

Repetitions of the Maringouin-Shepody contact have made exploration by drilling for Dorchester type mineralization (copper sulphides in the basal Shepody) much more difficult than if a single unambiguous contact existed. Rapid lateral facies changes and even minor structural displacements commonly prevent reliable correlations of stratigraphy between drill holes, and it is possible, therefore, to err in the identification of the "ore contact". Minor occurrences of mineralization at upper Maringouin-Shepody contacts may also mislead a drilling program. Consequently there is a very practical value to a better understanding of the Maringouin-Shepody transition.

The Maringouin Formation, generally considered Mississippian (Upper Windsor) in age, is characterized by brick-red fanglomeratic sediments ostensibly lacking in organic debris in the Cape Maringouin area. The Shepody beds of probable fluvial origin contain abundant remains of plants (e.g., *Calamites*), commonly coalified and partially pyritized. While the Shepody strata may qualify as Pennsylvanian beds on the basis of their fossil content, the so-called unfossiliferous Maringouin red beds should also be examined carefully as potentially Pennsylvanian in age for these strata do contain fossil impressions of plant debris. Diagenetic oxidation has removed the original organic matter and all but eliminated the evidence of prior organic remains in the Maringouin. Possibly the Maringouin and Shepody sediments are more closely related in age than previously considered, and the lithologic contrast between these formations may be due more to local oscillations in the environment of deposition than to any major temporal or climatic changes as has been previously suggested (*see* Schenk, 1969, p. 1060-1061).

The excellently exposed Cape Dorchester coastal section near Sackville, New Brunswick contains two major and several minor carbonate horizons in the upper Maringouin Formation. These horizons are characterized by a gradual upward increase in carbonate over silt content leading from typically brick red silts into mottled greenish and purplish-grey carbonate. The carbonate bed is commonly one foot or more thick while the underlying transition from silt to carbonate is several feet thick. Overlying beds are generally silty or sandy, or may contain more carbonate in the form of nodules. The occurrence of carbonate beds may represent caliche deposits formed at and below the depositional surface in a desert environment during a period of high evaporation and little sedimentation. Nodules within silt presumably resulted from continued fine grained sedimentation during upward interstitial migration of soluble carbonate. One caliche horizon has been highly silicified (in part jasperoid) and might be more properly labelled a silcrete. Study of these possible desert deposits is continuing. The apparent mobility of carbonate and silica in the early diagenetic environment invites speculation on the possible mobility of metals at the time.

References

Brown, A. C.
1975: A study of stratiform copper deposits in Carboniferous strata of New Brunswick and Nova Scotia; *in* Report of Activities, Part A, Geol. Surv. Can., Paper 75-1A, p. 598-600.

Schenk, P. E.
1969: Carbonate-sulfate-redbed facies and cyclic sedimentation of the Windsorian stage (Middle Carboniferous), Maritime Provinces; Can. J. Earth Sci., v. 6, p. 1037-1066.

Van de Poll, H. W.
1972: Stratigraphy and economic geology of Carboniferous basins in the Maritime Provinces; Twenty-fourth Internat. Geol. Congr., Montreal, Guidebook to Field Excursion A60, 96 p.

[1] Ecole Polytechnique, University of Montreal.

ERRATUM

Report of Activities, Part C
Geol. Surv. Can., Paper 75-1C

The following table was omitted from report 2, "Foraminifera and organic geochemistry of two sediment cores from a pock-marked basin of the Scotian Shelf" by G. Vilks and M. A. Rashid. Reference is made to these data on page 6 of Paper 75-1C.

TABLE 2. 1

Preserved plant pigments in Emerald Basin and Chaleur Trough cores.

Emerald Basin				Chaleur Trough			
Core	Core Interval (cm)	Chlorophyll	Phaeophytin	Core	Core Interval (cm)	Chlorophyll	Phaeophytin
8	0	.24	2.22	7B	0	2.19	25.81
	100	.32	3.22		100	1.82	26.44
	200	.20	3.65				
	300	.40	3.29		300	2.02	28.25
	400	.24	3.86		400	2.93	32.50
	500	.24	3.28		500	2.39	22.22
	600	.27	3.42				
	700	.28	3.12	(After Rashid *et al.*, 1975)			
	800	.35	3.98				
	900	.31	2.61				
	1000	.23	2.40				
	1100	.20	2.42				
9	top	.34	3.98				
	100	.23	2.16				
	200	.26	2.47				
	300	.19	2.59				
	400	.16	1.69				
	500	.21	2.09				
	600	.23	2.07				

AUTHOR INDEX

Report number in brackets.