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PAPER 77-1A

REPORT OF ACTIVITIES
PART A

Energy, Mines and Resources Canada

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1977

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INTRODUCTION

The "Report of Activities", has in the past few years, become an important means whereby the results of the Geological Survey's many scientific activities are made available to industry, the scientific community and the general public. Each of the three parts is published within about two months of the submission of the manuscripts, a fact of considerable importance to many users and a source of satisfaction to those responsible for the preparation and production of the reports. This volume, Part A, is of particular interest to a wide audience because it contains much new information resulting from the 1976 field program. Most of the papers making up this volume are concerned with studies in Terrain Sciences, Stratigraphic and Structural Geology, Precambrian Geology, Cordilleran Geology, and Mineral Deposits.

By means of field and laboratory investigations the geological characteristics of mineral deposits and their relationships to their geological environments are studied. The results of these studies enable the Geological Survey to determine features favourable to the occurrence of mineral deposits and to develop guides to assist in mineral exploration. Data on mineral deposits is of course also essential in making assessments of Canada's mineral resources and to development of programs to manage these resources, one of the responsibilities of the Department of Energy, Mines and Resources. Reports 5 to 8 present the results of studies designed to advance our understanding of the occurrence of uranium and form part of our major effort directed to this important energy resource. Report 1 presents the results of a continuing study designed to relate the nature and distribution of the mineral deposits in the northern Cordillera to its geological features. This study is designed to integrate commodity metallogeny studies with regional geology. Volcanogenic massive sulphide deposits are an important source of Canada's lead and zinc production (and to lesser extent copper and silver) and are the subject of ongoing studies. Report 2 gives some interesting relationships between grade and tonnage including the fact that lead grade varies inversely as copper and directly as zinc and that 80% of the deposits yet to be found should be between 0. 1 and 10 million tons.

The Geological Survey's Cordilleran and Pacific Margin subdivision, located in Vancouver, comprises a staff of about 25 who are concerned with geological studies in the Cordilleran Orogen designed to elucidate and describe the general composition, structure, origin and geological evolution of the Cordilleran Region. Reports 45 to 60 and 87 and 88 are concerned with this type of work. Such studies provide basic data for helping to assess mineral potential, to guide mineral exploration and to aid in the orderly development of land utilization. Current work is directed to two interrelated objectives – completion and updating of the reconnaissance phase of regional investigations to provide a broad geological and tectonic framework for the region, and detailed studies of specific problems, to further our understanding of the nature and sequence of geological processes with particular reference to the formation and localization of mineral deposits.

Copper mineralization in the Proterozoic basins of the Mackenzie Mountains has been known since 1962. Report 46 presents some interesting observations on the tectono-stratigraphic framework of the Redstone copper belt that should measurably assist in mineral exploration in this area. The Geological Survey has been conducting an ongoing geological reconnaissance of the Coast Mountains between southeast Alaska and Vancouver with the objective of elucidating the main events in the geological history of the Coast Crystalline Belt and of developing an understanding of the processes governing the formation of plutonic rocks in such orogenic belts. Despite most difficult weather conditions this work was continued in 1976 and Report 55 gives some of the results.

As part of its assessment of renewable energy resources, the Department of Energy, Mines and Resources has undertaken studies of geothermal deposits. The Geological Survey is making an inventory of the distribution, nature and geological setting of hotsprings in Canada and the chemistry of their waters. Report 58 describes the geochemistry of the Mount Meagher hotsprings area of British Columbia.

The Geological Survey began reconnaissance mapping of the Canadian Shield in 1842 using canoes and foot-traverses and completed the task 131 years later using helicopters. Much of the current work on the Shield is being done within the Northwest Territories and consists of upgrading the geological mapping of those areas in which supracrustal rocks have been preserved. These are areas of high economic interest and knowledge of their geology is essential to any economic assessment of the mineral potential of the Shield. The 18 contributions to our knowledge of Precambrian geology found in this publication include both the results of areal mapping and detailed studies. The area of the East Arm of Great Slave Lake has been interpreted as a faulted rift arm but many aspects of *the structural and magmatic development of the feature were poorly understood because of the lack of systematic large-scale mapping. Report 25 gives some of the results of a thorough remapping of the Proterozoic rocks of this area carried out in 1976 for publication at 1: 50 000 scale. The report presents new data and new concepts and is a significant contribution to the geology of the area. Reports 36 and 89 describe aspects of Archean geology from northwestern Ontario and reports 37 and 38 give details on rocks of similar age from the Abitibi-Kirkland Lake area of Ontario and Quebec. Reports 40 to 44 and 83 are progress reports on the upgrading of areas in Quebec and the Northwest Territories. The results of an ongoing study of the regional metallogeny and volcanic stratigraphy of the Superior Province are given in report 39. It is interesting to note that the Red Lake belt of northwestern Ontario has some features strikingly similar to those found in the Abitibi belt of the Timmins region.*

Stratigraphic, structural and paleontological studies are essential to understand the geological framework of the sedimentary basins of western and Arctic Canada. Such knowledge is especially important in ascertaining our coal, oil and natural gas resources. Reports 22 and 23 present results of studies carried out as part of "Operation Boothia" a study on Somerset Island and Boothia Peninsula designed to produce bedrock geological maps on a scale of 1: 100 000 or 1: 250 000 and to provide information essential for evaluating proposals for construction of a natural gas pipeline. A study was begun in 1975 to establish the distribution and continuity of coal seams in the Kootenay Formation of Alberta and British Columbia. This was completed in 1976 and the results are given in report 21. Two significant directions of faulting are present in Virginia Falls map-area, District of Mackenzie. The conclusions of a study of these faults, given in report 24, are of interest because the Cadillac lead-zinc deposits occurs in apparent association with a fault that parallels the prominent north-northeast direction of faulting. Similar associations of fault direction and stratigraphy are obvious exploration targets.

A considerable part of the Geological Survey's effort is directed towards providing geoscientific data and interpretive information on the surficial geology and geomorphic processes of the Canadian landmass and an evaluation of geotechnical aspects of bedrock geology that may affect the engineering use of the terrain. In order to meet these requirements several objectives must be met: systematic coverage of the surficial geology of the Canadian landmass; identification and assessment of the occurrence and magnitude of natural terrain hazards; and provision of the geological information needed to assist in the use of, maintenance and restoration of our physical environment. Because large parts of the country are unmapped, many of the activities reported in this publication reflect the response to current issues. Continuing exploration for energy resources in the Arctic Islands, the possibility that pipelines might be constructed and increased mineral exploration activity in the District of Keewatin require terrain information from a range of disciplines.

Reports 14, 79, and 95 give results of inventory mapping carried out in northern areas last summer and numbers 10, 17, 66, and 67 report on geomorphological studies made in the Arctic. Reports 93 and 94 show how studies of surficial materials can be applied to the search for mineral deposits – the technique of drift prospecting. As nuclear energy becomes increasingly important in planning Canada's future development the problem of disposing of radioactive waste becomes a matter of increasing concern. The Geological Survey is participating in a Departmental program supported by Atomic Energy of Canada Ltd. designed to evaluate igneous rock types as potential host rocks for radioactive waste isolation. These studies complement an evaluation of salt deposits also being carried out by the Survey. Reports 9, 80, and 81 are devoted to the appraisal of several igneous rock bodies and point out potential problems that may militate against their use.

Material for this Report of Activities was submitted by authors to their divisional offices by October 15. The papers were read by members of the divisional management and forwarded to the divisional directors for final reading and approval. Scientific editing was carried out between November 1st and 16th, production editing, typing, proofreading and preparation of camera-ready copy were done between November 2nd and 30th. Printing was scheduled to permit a mid-January release. Although close scheduling precluded grouping related reports in the text this relationship is shown by the "Contents" section.

As the "Report of Activities" is probably one of the most widely circulated and generally available Geological Survey publication, a list of its senior management is included in order to acquaint the general public with the names of those responsible for the Geological Survey's research program.

R. G. Blackadar
Chief Scientific Editor

REPRINTS

A limited number of reprints of the papers that appear in this volume are available by direct request to the individual authors. The addresses of the Geological Survey of Canada offices follow:

601 Booth Street,
OTTAWA, Ontario
K1A 0E8

Institute of Sedimentary and Petroleum Geology,
3303-33rd St. N.W.,
CALGARY, Alberta
T2L 2A7

British Columbia Office,
100 West Pender Street,
VANCOUVER, B.C.
V6B 1R8

Atlantic Geoscience Centre,
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B2Y 4A2

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1. REGIONAL METALLOGENY OF THE NORTHERN CORDILLERA

Project 740098

K. M. Dawson
Regional and Economic Geology Division, Vancouver

Field work in 1976 encompassed several mineralized areas in Yukon and Mackenzie, with some emphasis given to lower Paleozoic black shales in eastern Selwyn Basin. The Macmillan Pass region in southeastern Niddery Lake map-area (105 O/1, 2) was singled out for particular study by virtue of a recent discovery in the region of large and potentially economic stratiform barite and lead-zinc-barite deposits.

Geological Setting

Host rocks to these stratiform deposits are black shales, in part tuffaceous and pyritic, of the informally designated Devono-Mississippian 'Black Clastic' group. Shales and cherty clastic rocks of the group, which blankets most of Selwyn Basin, correlate with the Besa River Formation of the southern Mackenzie and northern Rocky Mountains, and with the Canol and Imperial formations of the northern Mackenzie and Richardson mountains (Blusson, 1976; pers. comm., 1976).

A generalized geological map (Fig. 1.1) shows the stratiform barite and lead-zinc-barite deposits in the Macmillan Pass area. Figure 1.2 presents a generalized stratigraphic column for the area.

Shales and associated strata of the 'Black Clastic' group are underlain by another extensive shale unit of early Ordovician to middle Devonian age assigned to the Road River Formation. Within Selwyn Basin Road River sediments comprise two broad lithofacies: a central chert and variegated shale unit of deep water character, and a flanking carbonaceous shale, shaly limestone and calcarenite group that changes laterally eastward to carbonates in the Mackenzie Mountains (Blusson, 1976; pers. comm., 1976). Carbon-rich Road River shales host a belt of large stratiform lead-zinc deposits at Howard's Pass in Nahanni map-area (105 I/6, 11, 12) and metamorphosed possible equivalents host the similar Grum-Vangorda-Swim deposits of the Anvil district (105 K). Apparently throughout Selwyn Basin Road River strata are underlain by Cambro-Ordovician 'wavy-banded' silty limestone which is in turn underlain by Proterozoic to Cambrian clastic rocks.

'Black Clastic' strata are overlain, apparently conformably, by relatively thin erosional remnants of Carboniferous shale, quartzite and carbonate which extend discontinuously across northeastern Selwyn Basin.

The 'Black Clastic' Group

In Macmillan Pass area the 'Black Clastic' group attains both its maximum observed thickness of about 1550 m and its greatest lithological complexity where Blusson (pers. comm., 1976) divides the group into lower and upper units that correspond respectively to the Canol (Bassett, 1961) and Imperial (Hume and Link, 1945) formations of the northern Mackenzie and Richardson mountains (*see* Fig. 1.2). The lower unit undergoes marked thickening at Macmillan Pass coincident with the appearance of stratiform barite and lead-zinc-barite deposits. In addition, black shales and siltstones in the lower part of the 'Black Clastic' group at Macmillan Pass contain a higher tuffaceous and cherty clastic component than the soft, black bituminous shales of the Canol formation, reflecting the relatively rapid deposition of these clastic and pyroclastic sediments within an east-west graben-like trough described by Blusson (1976, 1974). Where not removed by erosion, upper 'Black Clastic' siltstones, sandstones and shales correspond closely in total thickness (average 750 m) and lithology to the Imperial Formation.

In the Macmillan Pass area a basal 'Black Clastic' sequence of rhythmically interbedded siliceous black shale and siltstone about 200 m thick grades upward through tuffaceous shale and greywacke to sandstone, and locally, to chert pebble conglomerate. Both conglomerate and an overlying silvery weathering laminated black shale unit thicken abruptly in the vicinity of the TOM and JASON stratiform Pb-Zn-Ba deposits (Fig. 1.1) which, like several other bedded barite deposits in the area, occupy the same stratigraphic position near the base of the silvery weathering shale. Rapid thickening of these two units, from about 50 m to over 800 m combined, occurs at the eastern end of an elongate, possibly fault-bounded trough of 'Black Clastic' sediments which extends westward across Niddery Lake map-area (Blusson, 1974, 1976). The silvery weathering shale, which resembles typical Canol Formation shales to the north, is relatively tuffaceous, pyritic and carbonaceous within the stratigraphic interval containing base metal and barite beds. In addition, the mineralized horizon at the TOM and JASON deposits is pervasively silicified.

Barite and lead-zinc deposition apparently occurred during a period of quiescent sedimentation that followed or accompanied a period of uplift and erosion of chert-bearing Road River strata to the west and southwest. Tuffites may be distal to explosive subaerial volcanism of the same provenance.

Upper 'Black Clastic' sediments are coarser grained than underlying Canol Formation equivalents, and include coarsely cross-laminated sandstone with conglomerate lenses. The thick, resistant, reddish-brown weathering beds apparently were deposited in a shallow water and/or high energy environment during continuing uplift and erosion in late Devonian to Mississippian time. No mineral occurrences have been observed in this cherty clastic sequence, however some pyritic siltstone beds contain distinctive spherical pyrite concretions similar to those reported in the Imperial Formation by Hume (1954, p. 40-47).

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Geol. Surv. Can., Paper 77-1A (1977)

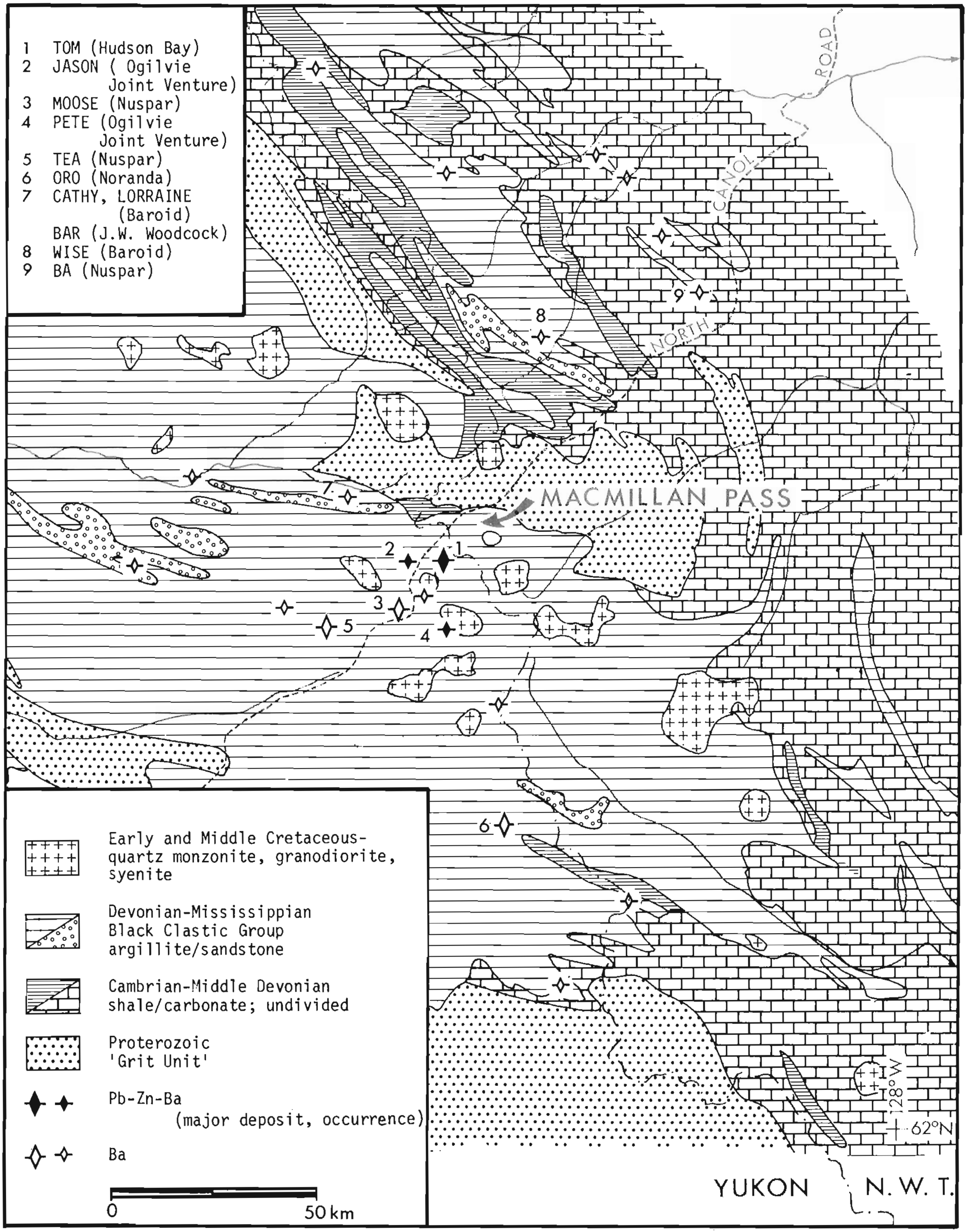

Figure 1.1. Stratiform Ba and Pb-Zn-Ba in (Dev.-Miss.) Black Clastic Group.

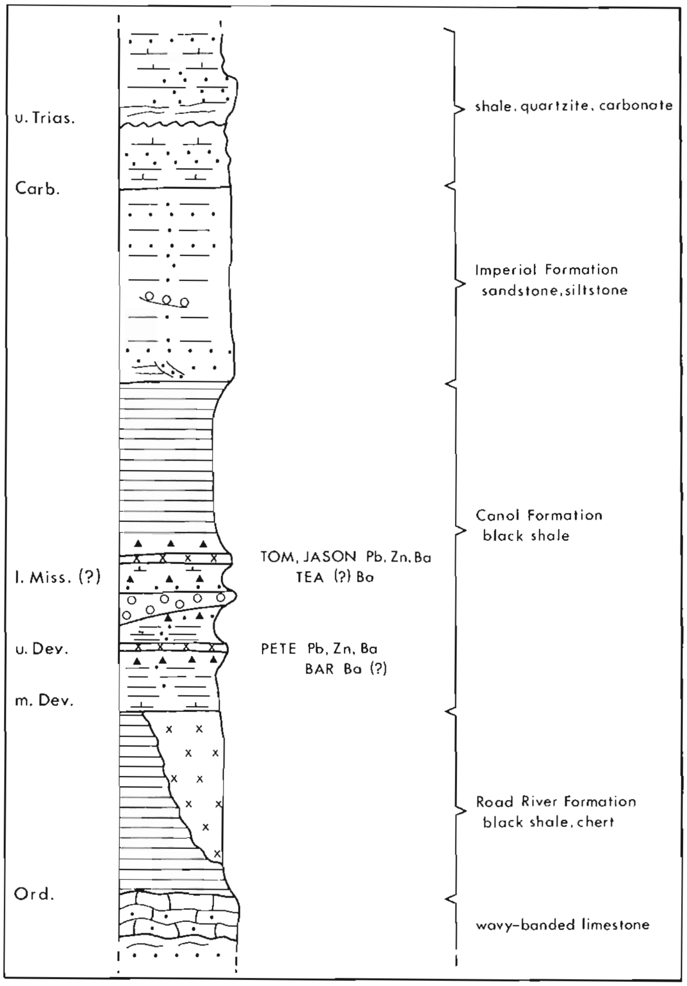

Figure 1.2. Generalized stratigraphic column and mineral deposits, Macmillan Pass area.

Barite Beds

A barite horizon is almost ubiquitous in Devono-Mississippian shales of the northern Cordillera. In almost all sections examined outside of Macmillan Pass area thin beds and lenses of barite could be found within the silvery weathering shale unit or its equivalent a few metres or tens of metres below the upper clastic beds. Barite beds of varying thickness constitute a broad metallogenic belt up to 160 km wide that follows the trend of Devono-Mississippian shales 300 km southeastward to Coal River map-area (95 D) and 400 km northwestward to the Richardson Mountains (Blusson, pers. comm., 1976). The belt may extend into northeastern British Columbia.

The barite horizon is widespread, but only in the eastern Selwyn Basin region (Fig. 1.1) have large amounts of barite been observed in thick, bedded sequences. Within this region, observed associations of Pb-Zn with bedded barite are restricted to the Macmillan Pass area, perhaps reflecting local sedimentary or volcanogenic controls favouring their co-deposition. A smaller, possibly correlative barite horizon in mid-Mississippian black slates has been recognized by Tempelman-Kluit (1976) in the Pelly Mountains south of Ross River.

The form and thickness of barite beds are not consistent over the extensive belt described above. In most Devono-Mississippian sections examined the 'barite zone' consists of tiny concordant barite lamellae dispersed through a thickness of several metres of black shale. Lamellar barite may grade to nodular or ovoid forms of 5-10 mm diameter with radial internal structure, that may, in turn, coalesce into coarse to fine laminated beds. The relatively resistant barite beds commonly resemble light to dark grey thinly bedded limestone, but may range from white to black. In general, bedded barite occurrences are 3 to 10 m thick and extend for several hundred metres along strike. The TEA deposit of Welcome North Mines Ltd. currently being developed by Yukon Barite (Fig. 1.1, No. 5) contains the thickest observed barite sequence, in excess of 100 m. Other bedded barite deposits with significant indicated reserves are Noranda's ORO, Nuspar's MOOSE and Baroid's WISE (Fig. 1.1, Nos. 6, 3 and 8 respectively).

Stratiform Pb-Zn-Ba Deposits

Two potentially economic Pb-Zn-(Ag) deposits, the TOM and JASON (Fig. 1.1), occupy the same stratigraphic horizon in barite beds at Macmillan Pass. The mineral assemblage in common is a thinly laminated sequence of white barite-pale sphalerite-galena-pyrite. Baritic ore may be overlain by or grade laterally to a pyritic assemblage of pyrite-sphalerite-galena which contains little or no barite. Mineralized beds attain a reported thickness of 40 m, but 5 to 15 m is more common. Unmineralized barite beds in the eastern Selwyn Basin region (Fig. 1.1), commonly are underlain by light grey micritic limestone beds a few metres thick that grade progressively upward to barite. Lead-zinc deposits occupy a silicified, pyritic interval in the silvery-weathering shale unit a few tens of metres above the chert pebble conglomerate bed. Thin beds of tuffaceous and cherty siltstone are relatively abundant in the same basal interval.

The TOM property, discovered by Hudson Bay Exploration and Development Company Limited in 1951 and developed through 1970, contains reserves of 8.6 million tons of 8.4% Zn, 8.1% Pb and 2.75 oz/ton Ag (Craig and Laporte, 1972). A bed of baritic ore had been traced at least 1500 m northwestward along one limb of a gently plunging monocline. The JASON and PETE properties were staked in 1974 and 1975 by Clyde Smith for Ogilvie Joint Venture (Brinco, Mitsubishi, Ventures West Capital). Lithologies and mineral assemblages at JASON are believed, by company geologists, to reflect the same horizon as at TOM, 5 km northeast (Sinclair *et al.*, 1976).

The PETE Pb-Zn-Ba occurrence, a similar but thinner and apparently stratigraphically lower mineralized horizon, is 16 km southeast of JASON. Diamond drilling is underway at the JASON property at time of writing.

The host black shale was deposited in a trough-like structure apparently formed by subsidence along a synsedimentary fault(s). The conformable nature of the ore beds, taken in conjunction with the relatively high iron (pyrite) and carbon content of the host rock supports the genesis of Pb and Zn sulphides as biogenic precipitates in an euxinic seafloor environment. However, the Pb-Zn-barite-silica assemblage, reminiscent of Rammelsberg ore (Hannak, 1968), is more logically attributed to an external source, possibly volcanic exhalations.

Age of Deposits

Limited fossil collections and determinations made to date show the age of one Pb-Zn-Ba deposit to be late Devonian, but some other barite and Pb-Zn-Ba deposits may be Mississippian. A limestone bed immediately underlying the Pb-Zn-Ba horizon on the PETE claims (Fig. 1.1, No. 4) contains late Devonian conodonts (B.E.B. Cameron, pers. comm., 1976). A limestone bed about 100 m stratigraphically below witherite-barite beds on the BAR claims (Fig. 1.1, No. 7) contains early to middle Devonian conodonts (Cameron, pers. comm., 1976). In the same area, but 1 km to the south on the CATHY claims, Blusson (pers. comm., 1976) noted late Devonian corals in reefal carbonates underlying barite beds. A late Devonian age of these barite beds is indicated. Limestone beds within the thick TEA barite sequence (Fig. 1.1, No. 5) contain Devono-Mississippian conodonts and a brachiopod tentatively assigned to the same broad range by Cameron (pers. comm., 1976). Carbonized wood of indeterminate genus was collected a few metres below mineralized beds at TOM by Sangster (1971), and similar material was found at about the same horizon at JASON by C.L. Smith (pers. comm., 1976). These deposits may be Mississippian, as may the widespread nodular and lamellar barite occurrences higher in the Canol Formation, but definitive fossil evidence is lacking.

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SOME GRADE AND TONNAGE RELATIONSHIPS AMONG CANADIAN VOLCANOGENIC MASSIVE SULPHIDE DEPOSITS

Project 650056

D. F. Sangster
Regional and Economic Geology Division

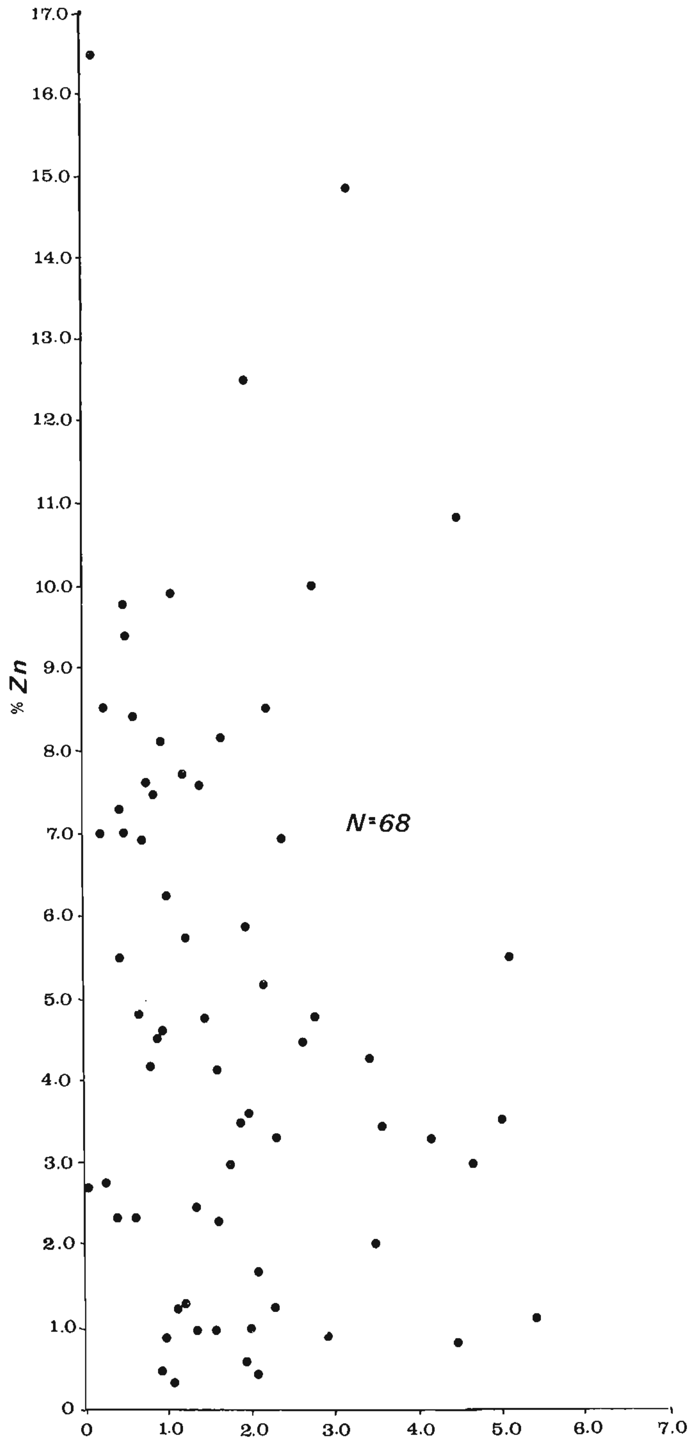

Figure 2. 1. Zn vs Cu grade for Archean deposits.

Over the past thirteen years, volcanogenic massive sulphide deposits have supplied between 60 and 75 per cent of Canada's zinc production and between 20 and 40 per cent of her lead production as well as significant proportions of copper and silver. Of the approximately 200 known deposits of this type in Canada, about 70 per cent are of Precambrian age and the geological characteristics of this large group were reviewed by the author in a previous publication (Sangster, 1972). The present report attempts to summarize some of the grade and tonnage data of the 200 deposits for which data are available. The best available reserve and production data were combined to provide a measure of the total size and grade of the deposits. Variations in the recovery efficiency of the metals were ignored, largely because they are unknown to the author, but also in the hope that they would not affect gross trends in the data. Incomplete reporting of reserves is a chronic problem in exercises of this type and so the author omitted deposits that he felt were not fully represented by published reserves.

The author is grateful to his colleague, Dr. R.I. Thorpe, for the use of data largely compiled by him, and to Messrs. D. Garson and G. Wine who drafted the diagrams.

For comparative purposes, the data were grouped according to whether the deposit is of Archean, Proterozoic, or Phanerozoic age. Essentially, three simple statistical plots were performed on each age group: (1) metal grade vs metal grade (2) grade vs tonnage (3) grade and tonnage frequency histograms. Although more than 30 plots were assembled, only a dozen representative ones are shown here and reference is made to the others throughout the text. Each data point in the plots represents a single geological deposit.

1. Metal vs Metal

For most volcanogenic sulphide deposits, particularly those in the Precambrian, zinc and copper are the most abundant of the three metals Cu, Pb and Zn. Figs. 2. 1 and 2. 2 are respectively, Zn vs Cu plots for Archean and Proterozoic deposits. In both cases, the scatter is considerable and no obvious correlation is discernible. A similar scatter was obtained for Phanerozoic Cu vs Zn (not shown) and although a mathematical correlation between the two elements might be demonstrable through computer analysis, it appears that copper and zinc grades in massive sulphide deposits vary independently of each other.

Lead is relatively abundant only in Phanerozoic massive sulphide deposits and the antipathetic relationship between copper and lead in these deposits has been

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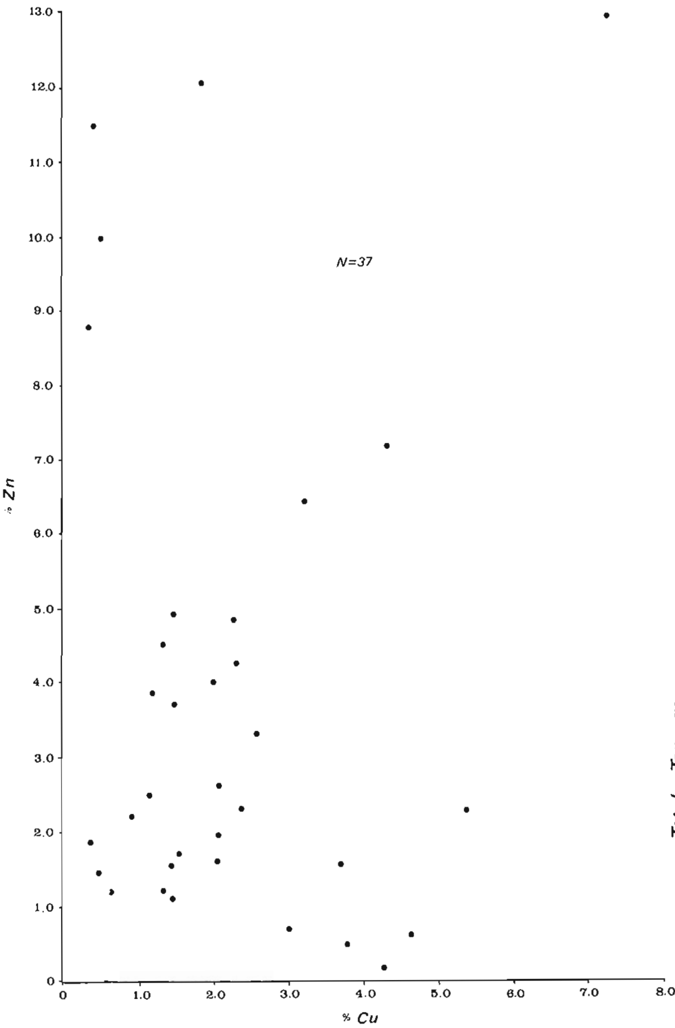

Figure 2. 2. Zn vs Cu grade for Proterozoic deposits.

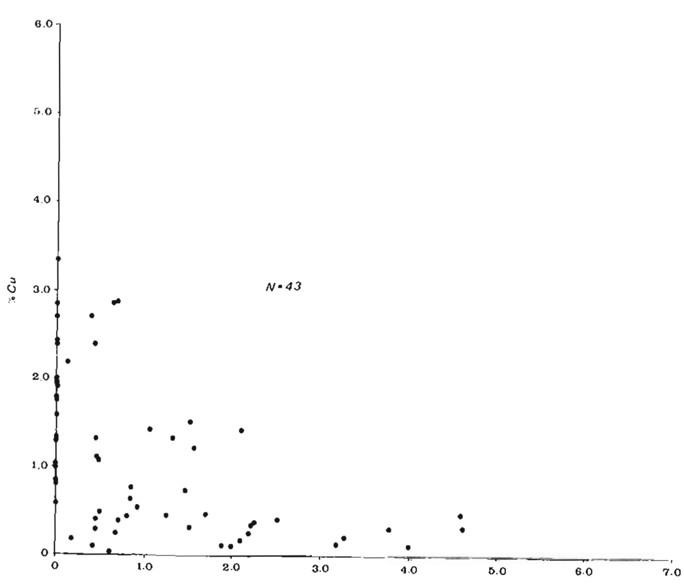

Figure 2. 3. Cu vs Pb grade for Phanerozoic deposits.

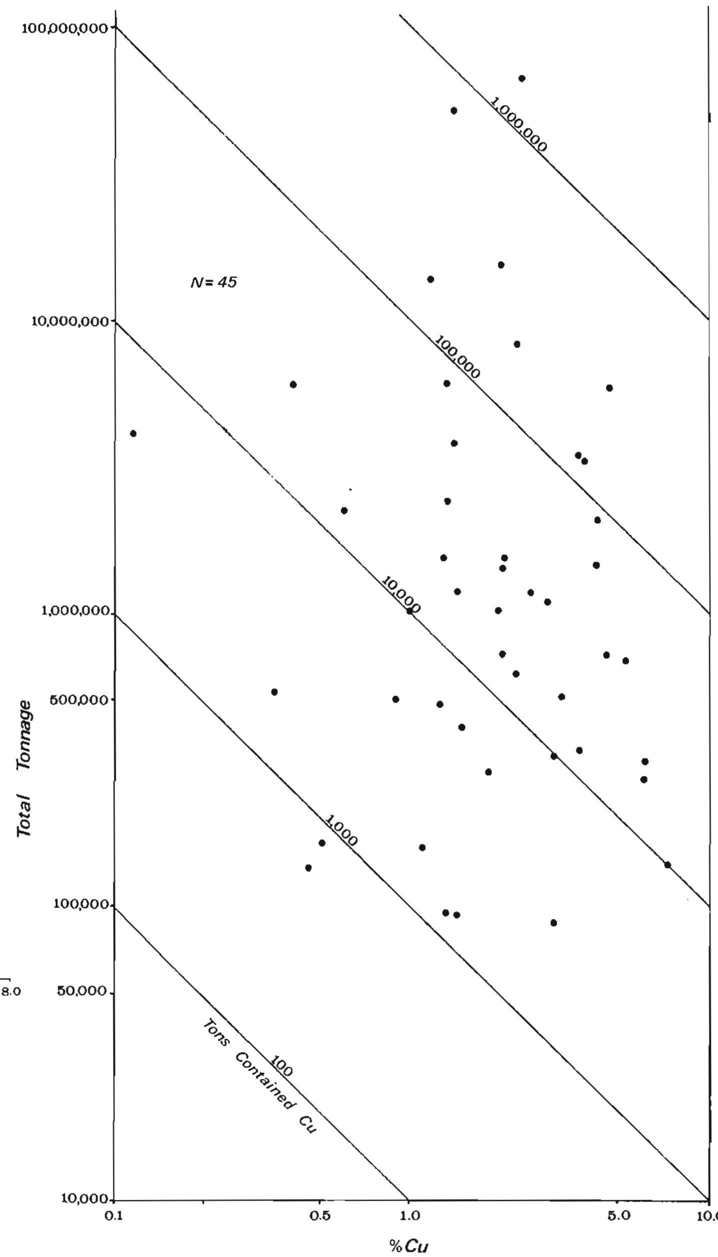

Figure 2. 4. Ore tonnage (short tons) vs Cu grade for Proterozoic deposits.

noted previously (Sangster, 1972, Fig. 8). To illustrate this somewhat more clearly, Cu-Pb data for the Phanerozoic group are presented in Figure 2. 3; it is clear from this diagram that there is an inverse correlation between the two elements. The antipathetic relationship between copper and lead is in contrast to a rather well-defined correlation between zinc and lead in Phanerozoic volcanogenic massive sulphide deposits. A plot of 43 data-points (not shown) revealed a reasonably linear trend with a Zn: Pb ratio of 3: 1. Stanton (1972, Fig. 15-12b), showed a similar linear relationship for 21 stratiform ore deposits of unspecified type, the only difference being that Stanton's data gave a Zn: Pb ratio of only 2: 1.

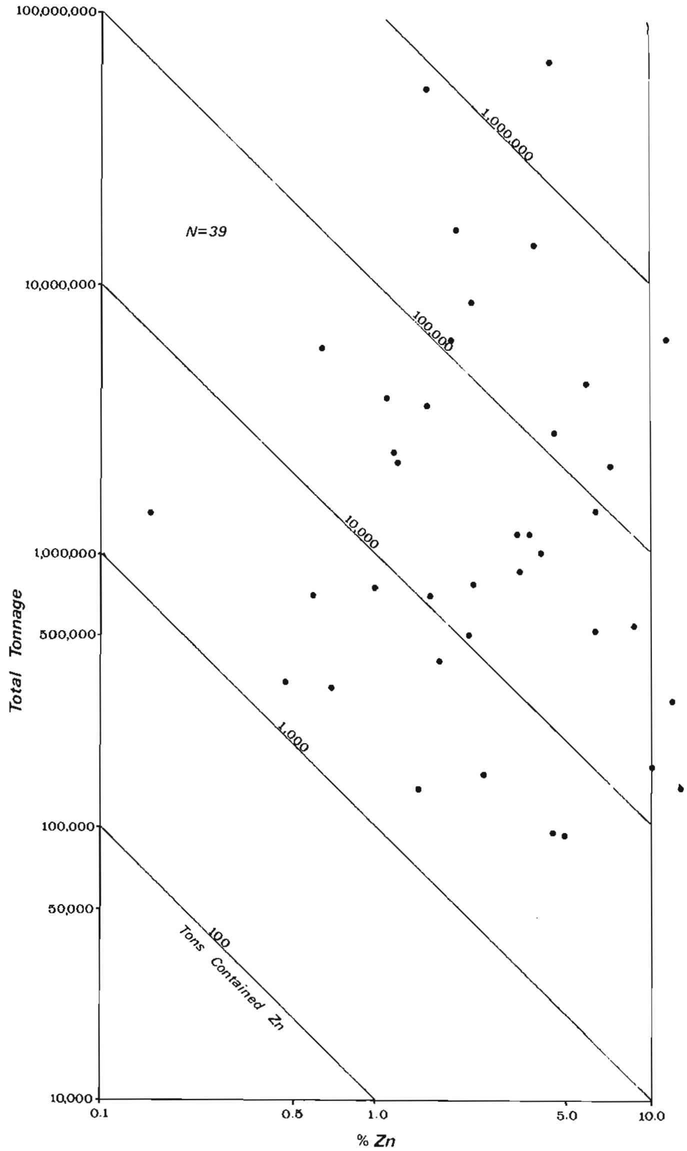

Figure 2.5. Ore tonnage (short tons) vs Zn grade for Proterozoic deposits.

2. Grade vs tonnage

In a recent publication Singer *et al.* (1975) demonstrated a negative correlation between ore tonnage and copper grade for 146 massive sulphide deposits. The scattergram (Fig. 2.4) for 45 Canadian Proterozoic deposits of the same type shows a distribution similar to that shown in Figure 3 of Singer *et al.* (1975). Their data, as well as the author's, show a distinct cut-off in data points in the low-grade, low-tonnage field (below about 3000 tons of contained copper) presumably due to economic, rather than geologic, reasons. Although the scattergram (Fig. 2.4) does not reveal an obvious correlation to the eye, a computer analysis

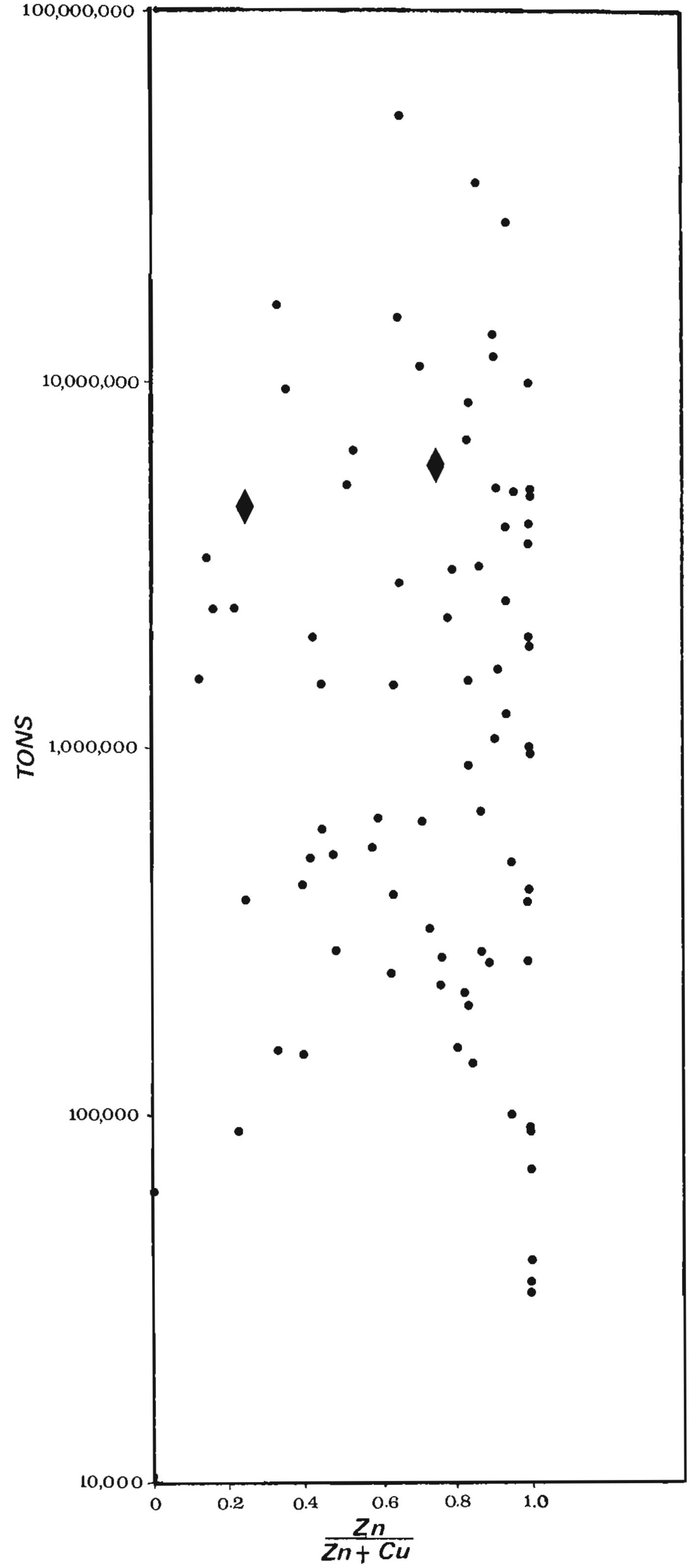

Figure 2.6. Ore tonnage (short tons) vs Zn/Zn+Cu ratio for Archean deposits. Diamond on left side of diagram represents average size (4.7 million tons) of deposits with a ratio less than 0.5, right-hand diamond is average (6.2 million tons) of deposits with ratio greater than 0.5.

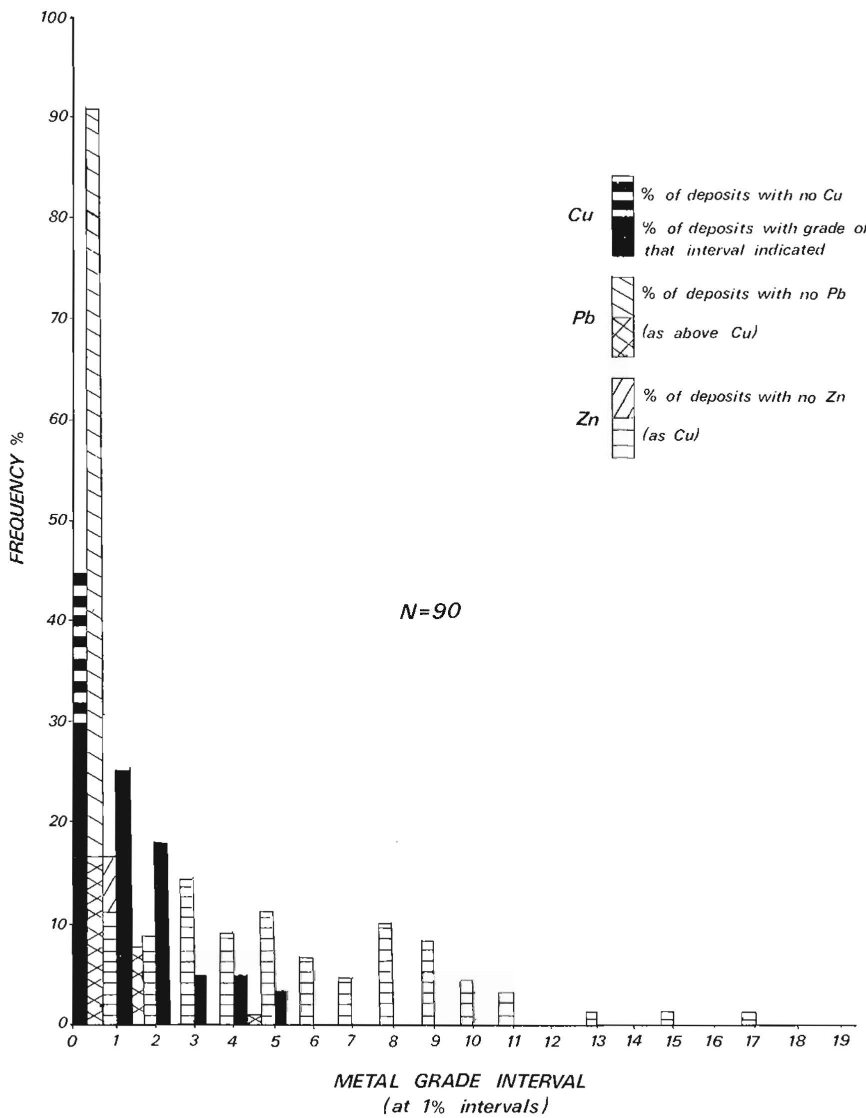

Figure 2. 7

Frequency of occurrence of metal grades in Archean deposits. In the range 0-1%, the upper portion of each bar show the percentage of deposits reporting no values in that metal (i. e. zero per cent). The lower portion of the bars denotes percentage of deposits with metal grade greater than zero but less than one per cent.

of the data might do so and more sophisticated statistical analyses are underway at present.

The plot of zinc vs total ore tonnage for Proterozoic deposits also shows a broad scatter of points (Fig. 2. 5). Again, a computer analysis may reveal correlations not evident to the eye but at least it may be concluded at this point that the two main elements of economic interest in massive sulphide deposits, copper and zinc, do not reveal a strong correlation with ore tonnage (similar plots for Archean and Phanerozoic showed an analogous distribution of points).

Noting the correlations between the elements mentioned previously (e. g. Cu vs Pb and Zn vs Pb), one might suspect that any correlation between one element and ore tonnage might be masked by the correlation of that element and one of the others. In an attempt to overcome this interelement "interference", ore tonnage was plotted against combined (Cu+Pb+Zn) grade. In all three age groups, the data show a scatter similar to that shown in Figures 2. 4 and 2. 5 and no obvious correlation is evident. However, one interesting aspect is revealed and that is: regardless of age, a consistent 88 per cent of all deposits in each age group contains combined Cu+Pb+Zn grades of less than 10 per cent. Because this constitutes an upper rather than a lower cut-off, and is consistent regardless of the Cu: Pb: Zn ratios in the more than 200 deposits sampled, it is tempting to regard it as a geological, rather than an economic, cut-off. The same data reveal that the most likely combined grade (Cu+Pb+Zn) is in the 5-6 per cent range.

To test a geological "hunch" of the author's, i. e. that the larger deposits seemed to be more zinc- than copper-rich, ore tonnage was plotted against Zn/Zn+Cu ratio for 83 Archean deposits (Fig. 2. 6). When the points are divided into two groups, those above and those below a Zn/Zn+Cu ratio of 0. 5, the uneven distribution of metal ratios in massive sulphide deposits becomes evident i. e. between 70 and 80 per cent of the deposits contain more zinc than copper, a property previously noted by the author (Sangster, 1972, p. 14). Ignoring, for the moment, the disparity in population

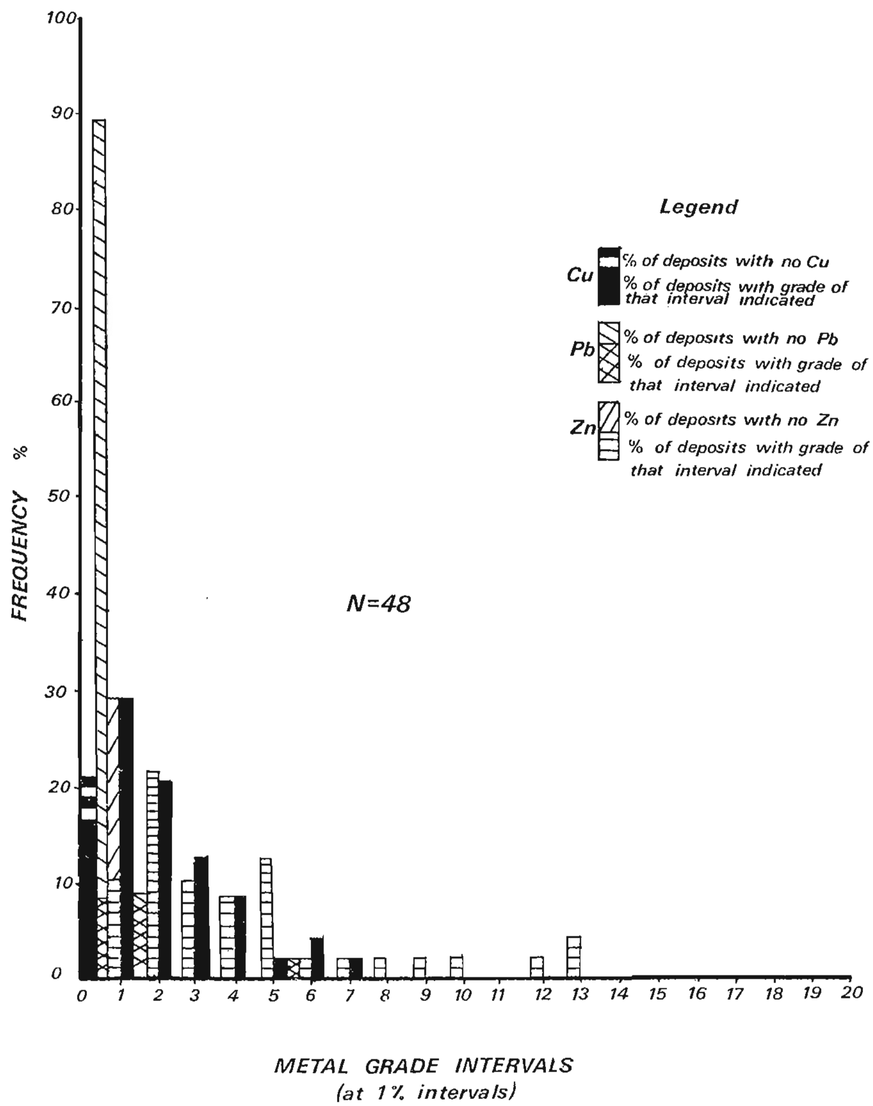

Figure 2. 8.

Frequency of occurrence of metal grades in Proterozoic deposits. In the range 0-1%, the upper portion of each bar show the percentage of deposits reporting no values in that metal (i. e. zero per cent). The lower portion of the bars denotes percentage of deposits with metal grade greater than zero but less than one per cent.

between the two groups, the average size (rather than the median size) of the relatively copper-rich deposits (i. e. those below a ratio of 0. 5) is 4. 7 million tons. Average size of the zinc-rich deposits, on the other hand, is 6. 2 million tons and the range in size of the zinc-rich portion is, in general, larger than that for the copper-rich deposits. It is possible that economic factors might have influenced this distribution as well because a small copper-rich deposit is presumably more amenable to exploitation than a zinc-rich deposit of the same size because of the price differential between copper and zinc.

3. Grade and tonnage frequency histograms

Perhaps of most interest to those in the mineral industry are the histograms of grade and tonnage for the three age groups. The dramatic decrease in the relative number of deposits with less than 1 per cent Pb from about 90 per cent in the Precambrian to about 65 per cent in the Phanerozoic, together with the increase in the range of lead grades, is evident from Figures 2. 7, 2. 8 and 2. 9. The rise in copper grade in the Proterozoic (Fig. 2. 8) relative to that in the Archean and Phanerozoic may be more apparent than real. The Proterozoic data are largely from deposits in the Flin-Flon – Snow Lake belt where cut-off grades may be above normal because of relatively high transportation costs between mine and mill. Zinc, as usual, shows the widest range in grade, commonly between 0-10 per cent whereas the other two metals each generally range only between 0-5 per cent. Taking all factors into account, a "best guess" estimate of the most likely grades to be encountered in volcanogenic massive sulphide deposits would be roughly 4% Zn, 1% Pb, and 1% Cu, in harmony with the "most likely" 6% total metal grade alluded to previously.

Grade, of course, means little unless it is coupled with tonnage. One comment that is frequently made when referring to massive sulphide deposits i. e. "all the large ones are in the Archean", is not supported by the data available. Examination of the tonnage data in the three age groups (Fig. 2. 10, 2. 11, 2. 12) reveals a

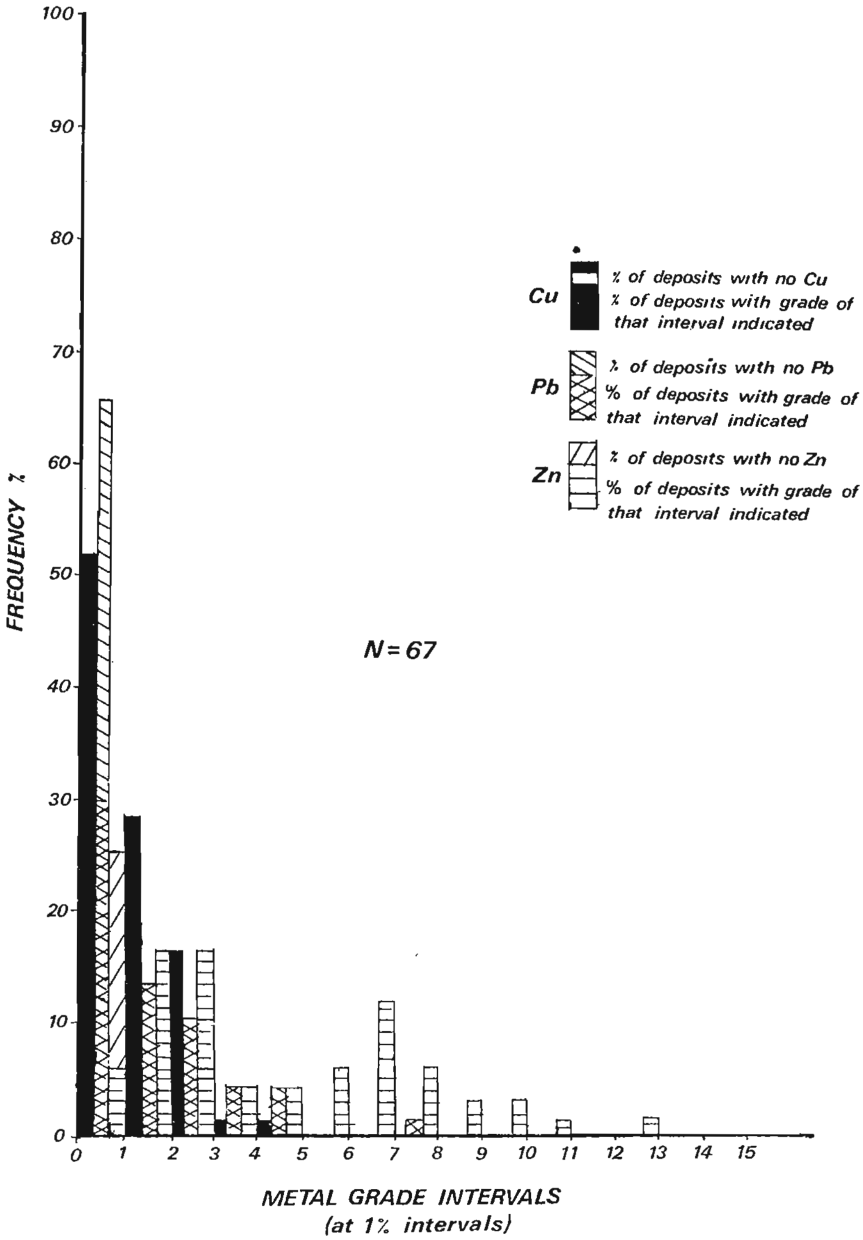

Figure 2. 9.

Frequency of occurrence of metal grades in Phanerozoic deposits. In the range 0-1%, the upper portion of each bar show the percentage of deposits reporting no values in that metal (i. e. zero per cent). The lower portion of the bars denotes percentage of deposits with metal grade greater than zero but less than one per cent.

number of interesting features: (1) in each group, the data define a log-normal distribution; (2) the proportion of deposits in the different size ranges is remarkably similar in the three age groups i. e. there is no well-defined tendency for one age group to contain, on the average, smaller or larger deposits than the other two groups.

Data for this study were entirely supplied by the mineral industry and, as is usual when one does not collect original data, one is constrained by the data available. The data, of course, are controlled by economic as well as geological factors and therefore it is unwise to draw sophisticated geological conclusions from them. For the most part, the deposits represented by these data represent only that part of the geological population of deposits that has been discovered and/or exploited.

Nevertheless, with these and other caveats in mind, certain general trends may be tentatively discerned in these plots: (a) when exploring for massive sulphide deposits in volcanic terrain of any age in Canada, 80 per cent of the deposits being sought will fall in the size range 0. 1 to 10 million tons and roughly half of these will be less than one million tons in size; (b) using age, rather than, for example, depositional environments, as the only criterion, the most likely average grade to be encountered will be in the order of 4% Zn, 1% Pb, and 1% Cu; (c) larger deposits may be zinc-rich relative to copper but beyond that, correlation of metal grade (rather than metal ratio) with tonnage is not obvious and the two may be independent; (d) lead grade varies inversely as copper and directly as zinc.

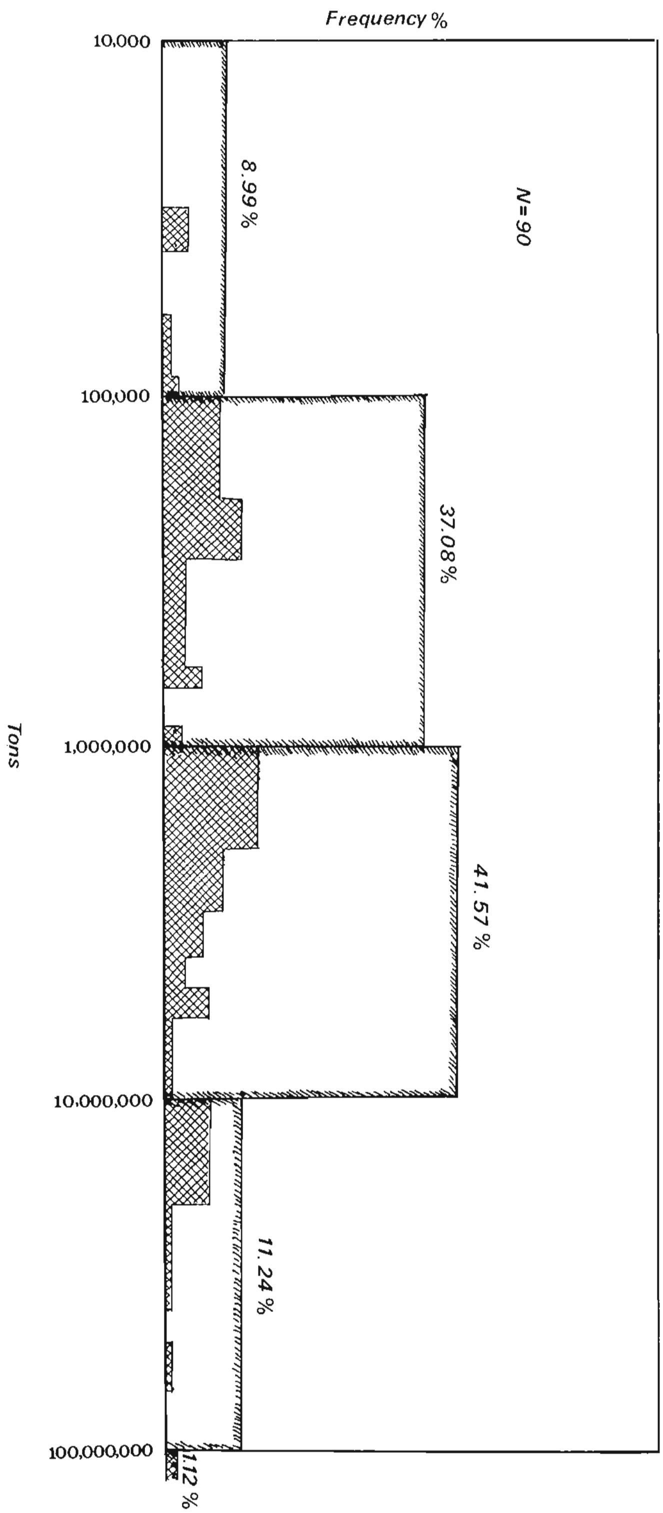

Figure 2.10. Frequency of occurrence of ore tonnages in Archean deposits.

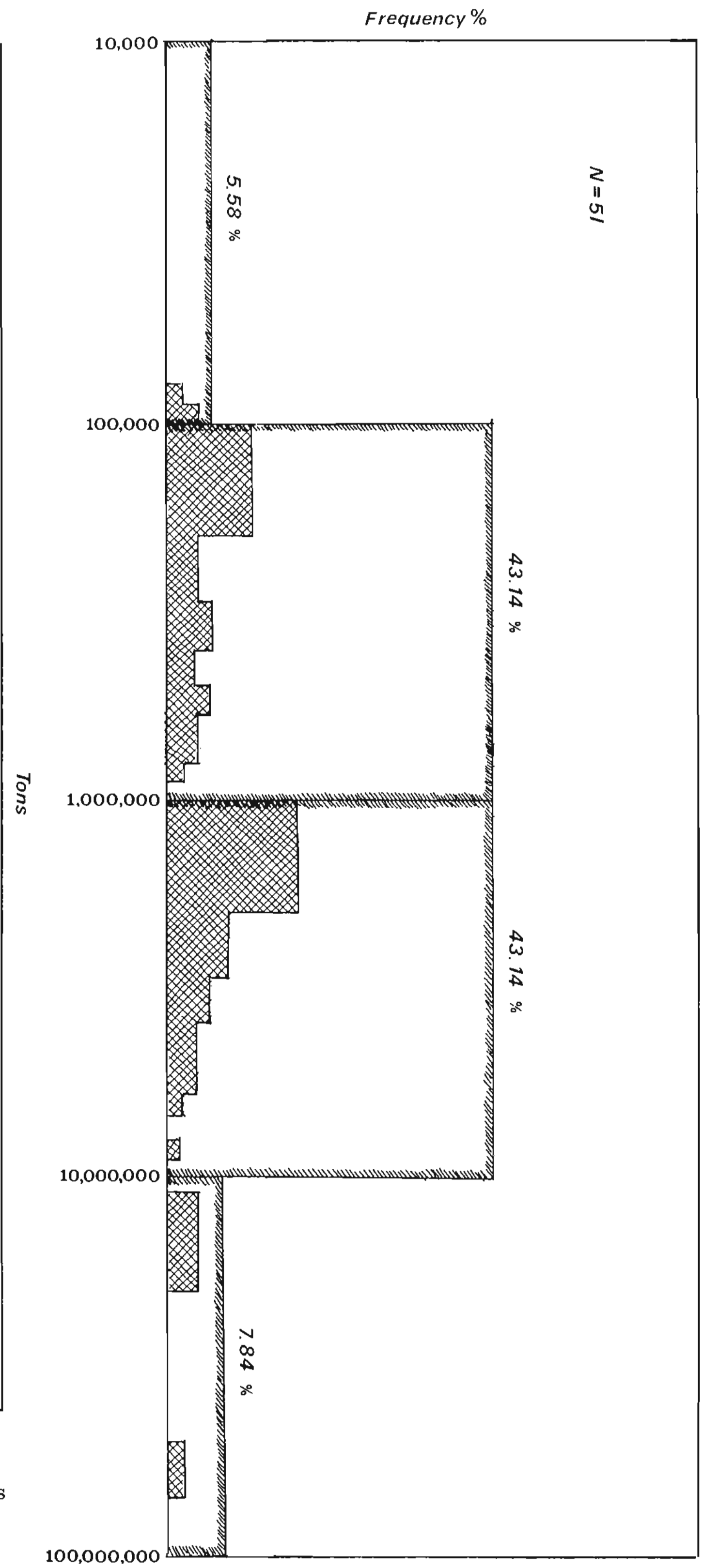

Figure 2.11. Frequency of occurrence of ore tonnages in Proterozoic deposits.

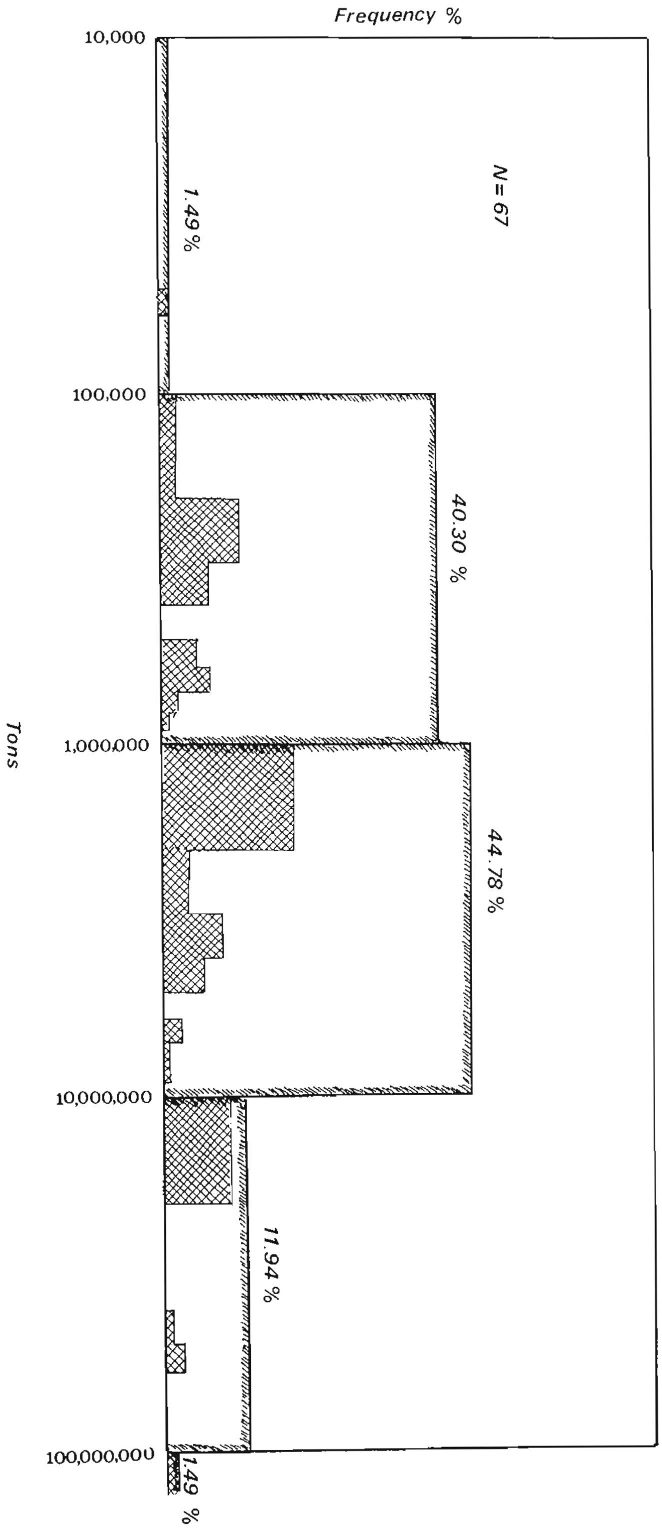

Figure 2.12. Frequency of occurrence of ore tonnages in Phanerozoic deposits.

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3. A PRELIMINARY STUDY OF IRON OXIDE AND MANGANESE OXIDE UNITS ASSOCIATED WITH VOLCANOGENIC SULPHIDE DEPOSITS, SHERBROOKE AREA, QUEBEC

Project 650056

D. M. Watson and D. F. Sangster
Regional and Economic Geology Division

The Sherbrooke area of the Eastern townships of Quebec has been of interest to geologists since the early 1800's. Logan visited the area in the 1850's, and between 1859 and 1866 it was the most active mining area in the two Canadas. This area was most recently mapped by Lamarche and St. Julien in 1962 and 1963 (St. Julien and Lamarche, 1965).

The major metallic deposits in the area, the Albert, Eustis, Capel and Suffield mines (Fig. 3.1), were worked intermittently until 1953 at which time the Suffield mine ceased operations. In recent years there has been a minor amount of exploration activity in the area, particularly near the old Suffield property.

Despite the geological and mining activity in the region, the relationship of the various units, particularly those rich in manganese and iron oxide, to the orebodies is as yet undefined. Manganese and iron-rich rocks (possibly exhalites), are in contact with the orebodies at the Suffield, Howard, King, and Silver Star mines and are close to, if not actually in contact with, ore at the Albert and Capel mines.

The purpose of this study is to conduct a petrographic and chemical examination of these iron and manganiferous rocks to determine if there is any relationship between them and the massive sulphide deposits of the area, and to determine whether or not these relationships would be useful in exploration for further ore.

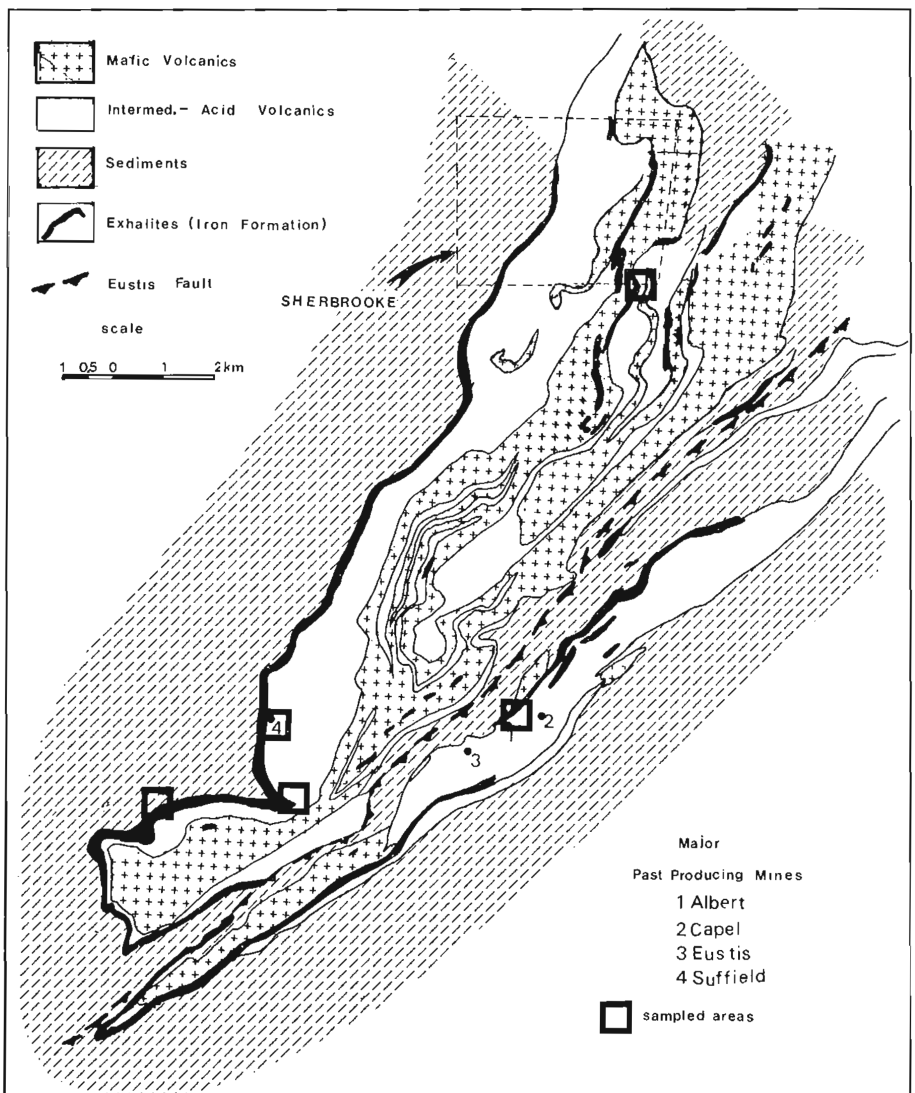

Figure 3.1.

Generalized geology of the Sherbrooke area, showing sampled area (adapted from St. Julien and Lamarche, 1965).

From: Report of Activities, Part A;
Geol. Surv. Can., Paper 77-1A (1977)

The senior author visited the area in early August, 1976, in order to become familiar with the geology of the district. A series of samples from all of the various units present was collected. To aid in field identification of the manganiferous horizons, a "manganese field test" was developed as follows: 1) add 2-3 drops of concentrated sulphuric acid (H_2SO_4) to a few chips of sample in a glass tube or vial; 2) add 2-3 drops of 5% silver nitrate solution ($AgNO_3$); 3) add a few grains of ammonium persulphate ($(NH_4)_2S_2O_8$); 4) development of a pink to deep red colour indicates the presence of manganese.

Preliminary examination of these first samples showed: (1) the rocks which are possibly exhalative (hematite and magnetite iron-formations and manganiferous cherts) were very distinctive and could be easily recognized in the field; (2) the unit described by St. Julien and Lamarche as "cherts, siltstones, sandstones, and thinly bedded rocks" (St. Julien and Lamarche, 1965), everywhere contains manganese in detectable quantities. Samples of rocks adjacent to (at times in contact with) these rocks, showed no signs of manganese when tested with the field kit. The hematite-jasper and magnetite-rich iron-formations also contained detectable manganese as did some of the sulphides from the Albert, Capel and Suffield mines.

In early September, a second trip was made to collect samples from the iron-formations and manganiferous rocks. Approximately 150 samples were obtained, the largest number being from the vicinity of the Suffield mine, because here the manganese-rich horizon lies in contact with the sulphide deposit and is well exposed. A total of 65 samples were collected across 50 feet perpendicular to the strike and for 500 feet along the strike from the ore zone.

In addition, a similar series of samples was collected in the vicinity of the Albert mine, and at several other places on this horizon, that were not in the vicinity of known orebodies.

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4. MINERALOGY OF SERPENTINIZED ULTRAMAFIC ROCKS AND ASSOCIATED NICKEL DEPOSITS

Project 750034

J.M. Duke
Regional and Economic Geology Division

Many nickel sulphide deposits are associated with olivine-rich ultramafic igneous rocks which have undergone partial or complete serpentinization. The grades of such deposits are usually in the range from 0.4 to 4.0 per cent nickel, whereas barren peridotites and dunites typically contain between 0.1 and 0.4 per cent. It is a commonly held view that these 'background' concentrations of nickel occur mainly in solid solution in silicate and oxide minerals and therefore are not recoverable by conventional metallurgical processes. This metallurgical consideration has discouraged the exploration and development of low grade (i.e. less than about 1 per cent) nickel deposits in serpentinized ultramafic rocks despite the fact that the distribution of nickel among the constituent minerals of these rocks has not been adequately documented in the literature. Eckstrand (1971) has discussed some of the implications of this problem. The mineralogy and nickel distribution in a number of Canadian occurrences of nickeliferous serpentinite are being studied to supplement the commodity investigation of nickel by O.R. Eckstrand.

The investigation will focus initially on the Dumont ultramafic body in northwestern Quebec, which is the best known Canadian example of the Mount Keith type of nickel sulphide deposit (Eckstrand, 1973, 1974, 1975; Naldrett and Arndt, 1975). It is a steeply-dipping lensoidal body with a true thickness of 1000 to 2000 feet along most of its 4.5-mile strike length. The lens is reported to contain about 500 000 000 tons of material averaging 0.327 per cent nickel to a depth of 1000 feet with a central, higher grade zone averaging 0.646 per cent (Northern Miner, July 13, 1972). The lens comprises a central core of dunite which grades into wehrlitic peridotite towards both upper and lower contacts. The rocks have undergone different degrees of serpentinization with the proportion of magmatic minerals (olivine + chromite ± calcic clinopyroxene) ranging from 0 to 90 modal per cent. Talc-carbonate alteration occurs locally in marginal zones of the body.

The diversity in the nature and degree of alteration, together with the excellent preservation of primary textures and the minimal effects of deformation make the Dumont ultramafic lens particularly suitable for study. It will be possible to determine the distribution of nickel among a wide variety of primary and secondary mineral species and, moreover, to see to what extent the distribution is dependent on the degree of alteration.

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5. CONCEPTUAL MODELS FOR URANIUM DEPOSITS AND AREAS FAVOURABLE FOR URANIUM MINERALIZATION IN CANADA

Project 750010

V. Ruzicka
Regional and Economic Geology Division

An understanding of the geology and genesis of uranium deposits is fundamental to the discovery of new deposits and to the efficient exploitation of known deposits.

A general model simulating formation of uranium deposits (Fig. 5.1) postulates:

(1) Primary concentration of uranium and its geochemical availability (i.e. source);

(2) Mechanisms of transport of uranium from the site of origin or pre-concentration to the site of re-deposition; this can involve either mechanical or chemical media;

(3) Concentration and deposition of uranium which can occur in structural and lithological traps due to: differentiation, metasomatism, precipitation (due to P-T changes, redox changes, evaporation, etc.), sedimentation, absorption, etc.;

(4) Modification of uranium deposits (due to metamorphism, during diagenesis, by oxidation, accretion, etc.);

(5) Preservation of uranium deposits from destruction (retention of sufficient uranium to comprise an exploitable deposit).

The majority of the world's known and presently economic uranium resources are of the following genetic types: (a) in Lower Proterozoic quartz-pebble conglomerates (syngenetic conglomerate deposits), (b) in veins and related deposits, (c) in granitic-pegmatitic rocks, (d) in sandstones, and (e) in supergene deposits. Deposits of other types are generally low grade and/or small. Therefore attention will be paid to conceptual models simulating formation of the above mentioned types of uranium deposits.

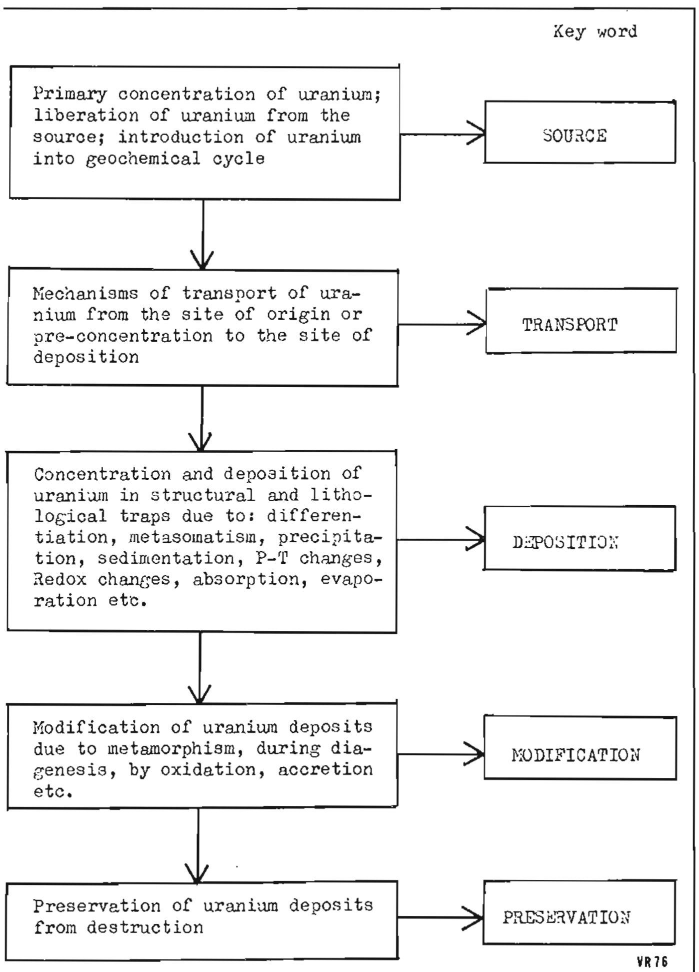

Figure 5.1. A conceptual genetic model simulating formation of Uranium Deposits.

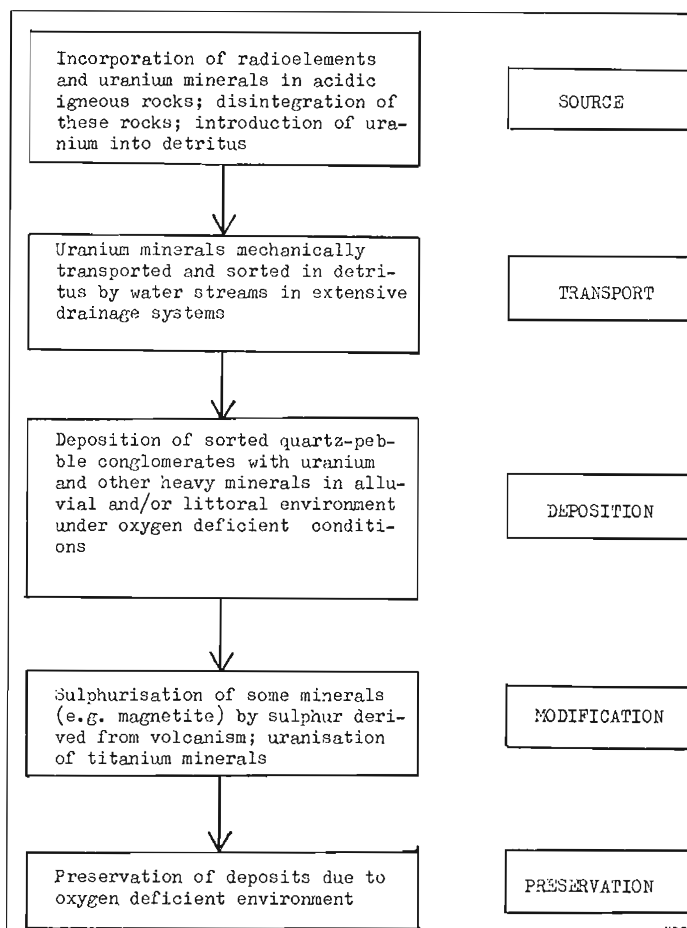

Figure 5.2. A conceptual model simulating formation of Uranium Deposits in Huronian Conglomerates.

From: Report of Activities, Part A;
Geol. Surv. Can., Paper 77-1A (1977).

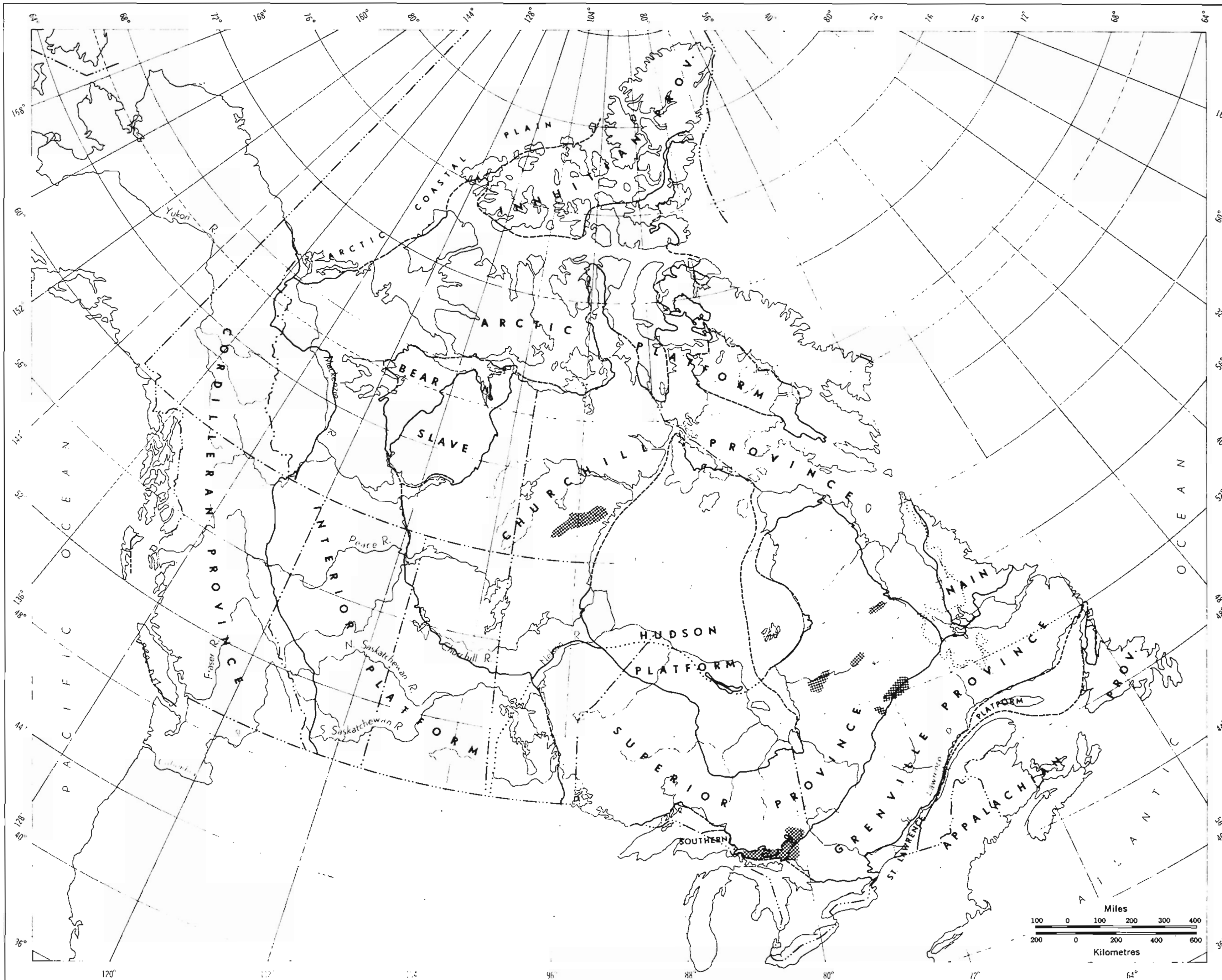

Figure 5.3. Areas favourable for syngenetic conglomerate deposits in Canada.

The conceptual model simulating processes leading to formation of syngenetic conglomerate deposits (Fig. 5.2) postulates the source of the deposits to be incorporation of radioelements and uranium minerals in acidic igneous rocks, disintegration of these rocks and the release of uranium minerals into resultant detritus. According to this model, uranium minerals are mechanically transported and sorted in detritus by stream waters in extensive drainage systems. Deposition of sorted quartz-pebble conglomerates with uranium and other heavy minerals occurs as a rule in an alluvial and/or littoral environment. Disintegration of the source rocks, transportation of the detritus and deposition of the sorted detritus takes place under oxygen deficient conditions, which probably existed during Lower Aphebian time (Roscoe, 1969). Modification of the mineralization occurs during the diagenetic stage of sedimentation. At least two kinds of alteration of the original minerals can be observed: sulphurization of some minerals (e.g. magnetite) by sulphur produced apparently by volcanism; uranization of titanium minerals. Preservation of deposits is possibly due to an oxygen-deficient environment caused by burial, presence of pyrite and solidification of the deposits.

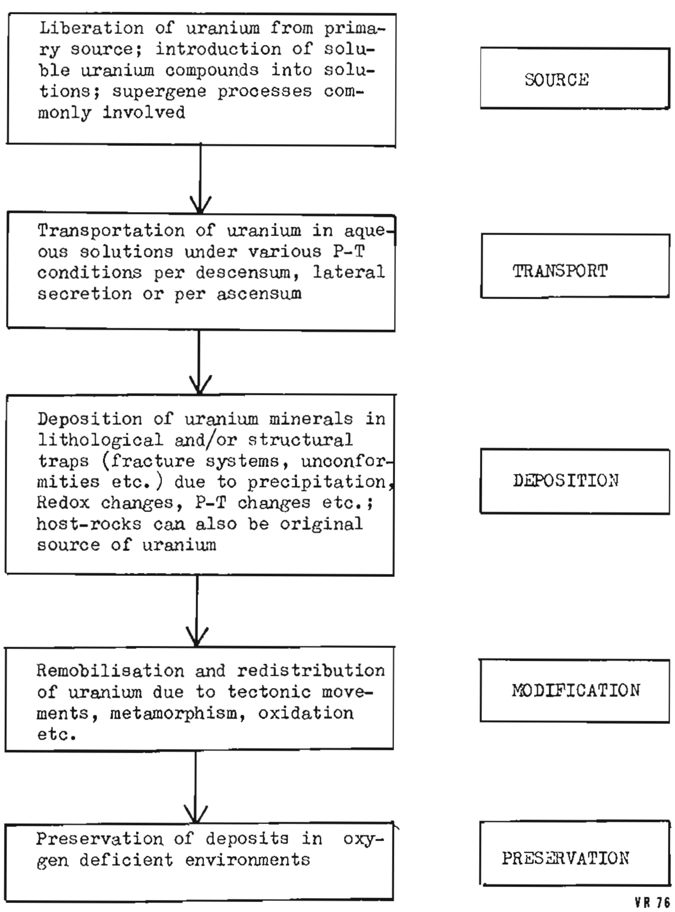

Figure 5.4. A conceptual model simulating formation of veins and related Uranium Deposits.

In Canada, the Lower Proterozoic uraniferous conglomerates of the Elliot Lake — Agnew Lake deposits exhibit certain features that can be used as a guide for further exploration, either in areas surrounding the identified deposits, or in analogous lithostratigraphic formations deposited under oxygen deficient conditions (e.g. the Matinenda Formation), elsewhere in the Shield (Fig. 5.3).

Besides the Elliot Lake — Agnew Lake area, the conglomerates occur in the Cobalt Embayment north and northeast of Sudbury. A few years ago radioactive occurrences in the Papaskwassati Basin of northern Quebec were targets of exploration for deposits of this type, but results show that these radioactive strata are somewhat younger than the Matinenda Formation of the Elliot Lake area and therefore are not comparable. During a complex magnetic, electromagnetic and radiometric survey in the early 1970's, INCO discovered uranium mineralization east of James Bay, in the vicinity of Sakami Lake (Robertson, 1974). However, correlation of these sediments with those in the Elliot Lake — Agnew Lake area is uncertain. Some uranium and gold was found in the pre-Hurwitz sediments of the Montgomery Lake Formation in the Padlei Fold Belt west of Hudson Bay (*ibid.*). The low uranium and gold contents in these conglomerates reflect the nature of the source rocks in their provenance area, which differs from that in the Elliot Lake — Agnew Lake area.

The conceptual genetic model simulating formation of veins and related types of uranium deposits (Fig. 5.4) postulates a pre-concentration of uranium in the primary source, liberation of uranium from this primary source, and introduction of uranium into solutions. Supergene processes are commonly involved in the transformation of insoluble compounds of tetravalent uranium into water-soluble compounds of hexavalent uranium. The host rocks themselves can also be one possible source of uranium. Transportation of uranium in aqueous solutions occurs under various P-T conditions during their descent, lateral movement or ascent. The mechanisms by which uranium is transported are complex and still poorly understood. At low temperatures, uranium is commonly transported by organometallic (chelate) complexes, or by CO_2-rich fluids in the uranyl carbonate ion state and precipitated as pitchblende which can itself originate from partial reduction of hexavalent uranium compounds to tetravalent (Moreau, 1974). (Pitchblende is commonly a mixture of oxides of hexavalent and tetravalent uranium.) Under low temperature conditions only uranium becomes mobile and thorium remains at the site of origin. At high temperatures, i.e. above 500°C, in the domain of igneous processes (magmatism and volcanism), uranium is transported in the tetravalent or hexavalent state commonly in the form of halides and is accompanied by thorium and the rare-earths. It is precipitated in the form of uraninite or in U-Th-REE minerals. Deposition of uranium takes place in lithological and/or structural traps (fracture systems, unconformities, etc.) due to precipitation brought about by redox changes, P-T changes (e.g. drop in CO_2 pressure) etc. A hypothetical

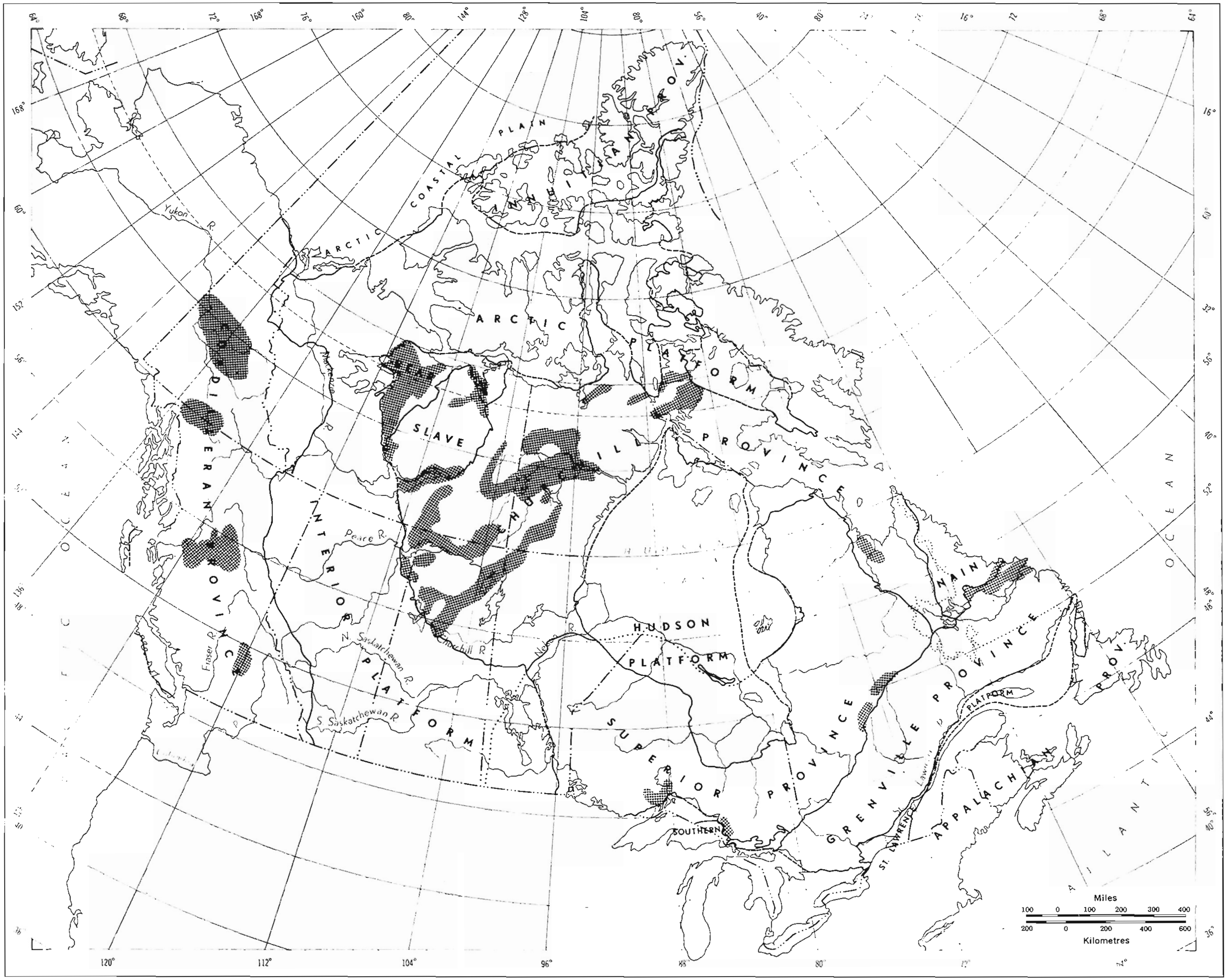

Figure 5. 5. Areas favourable for vein and related types of uranium deposits in Canada.

sequence of processes involved in formation of a pitchblende-rich orebody can be deduced from the "D" uranium deposit at Cluff Lake in the Carswell structure of northern Saskatchewan (Ruzicka, 1975). (1) The uranium, gold and other metals were presumably liberated by weathering from the pre-Athabasca complex (which can be tentatively correlated with the Tazin Group of the Beaverlodge area). A thick regolith occurs on top of these basement rocks, i. e. beneath the distinct pre-Athabasca unconformity; (2) the uranium is concentrated mainly in pelitic rocks, which also contain a high amount of bitumen, at the base of the Athabasca Formation and in the basement complex. The uranium was apparently remobilized from the pelites and redistributed along faults and fractures which served as structural traps. Deposits in the Wollaston Lake fold belt (e. g. Rabbit Lake, Key Lake) are also related to the pre-Athabasca unconformity. It is interesting to note that the pre-Athabasca unconformity is a time correlative of the pre-Kombolgie unconformity in the East Alligator River region of Northern Territory of Australia, where the world famous uranium deposits Ranger I, Ranger III, Koongarra, Jabiluka I and Jabiluka II occur. Areas favourable for veins and

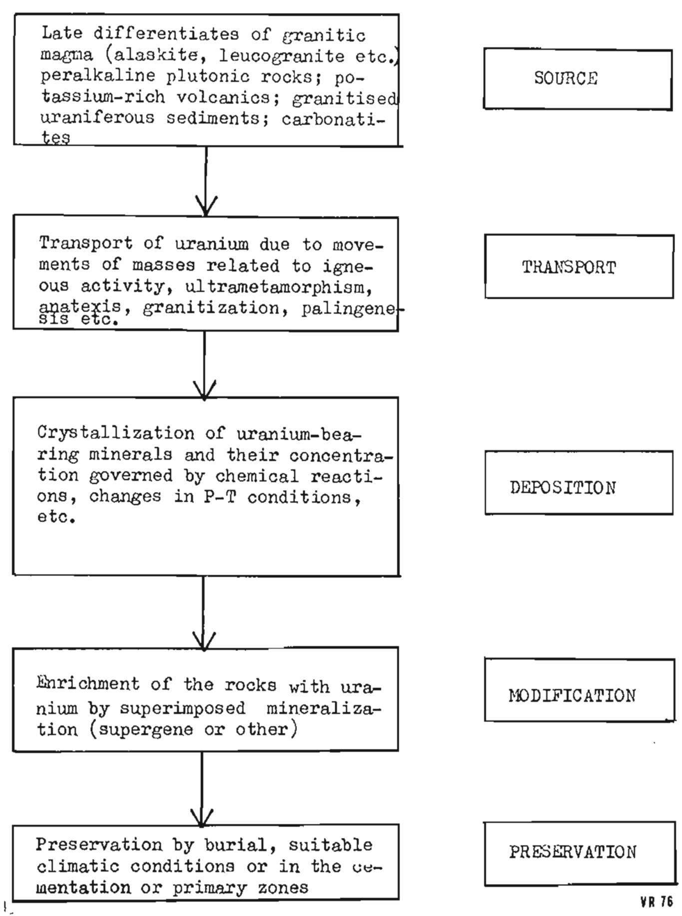

Figure 5. 6. A conceptual model simulating formation of orthomagmatic and anatectic Uranium Deposits in Granitic-Pegmatitic rocks ("Porphyry"-Uranium).

related types of uranium deposits in Canada are shown on Figure 5. 5. One of the most promising areas in Canada is that along the pre-Athabasca and pre-Martin unconformities. It includes rocks of the Wollaston Lake and Tazin belts. Similar geological environments can be found along the unconformities beneath the Thelon Basin and Bathurst Embayment in Northwest Territories and, in a little younger environment, along the pre-Sibley unconformity in Ontario.

The conceptual model simulating formation of orthomagmatic and anatectic uranium deposits in granitic-pegmatitic rocks postulates as source late differentiates of granitic rocks (alaskite, leucogranite, etc.), peralkaline plutonic rocks, potassium-rich volcanics or granitized uraniferous sediments (Fig. 5. 6). Uranium is remobilized and reconcentrated due to igneous activity, ultrametamorphism, anatexis, granitization, palingenesis etc. The mineralization can be further enriched in uranium by supergene processes. It is believed that secondary mineralization is an important factor in the formation of the "porphyry-uranium" (Armstrong, 1974) type of deposits. For example in the Rössing uranium deposit in South-West Africa, approximately 55% of the uranium is in uraninite, 5% in betafite and 40% in secondary uranium minerals (Backström, 1970). The average grade of this deposit is 0. 7 lb. U_3O_8 per short ton. Preservation of such mineralization is possible by burial or by conditions preventing depletion of uranium minerals.

In Canada, environments favourable for this type of uranium deposit can be found in granitic-syenitic terranes of the Canadian Shield and in the Cordillera (Fig. 5. 7).

The conceptual model simulating formation of uranium deposits in sandstones (Fig. 5. 8) postulates as a source, silicic crystalline and/or tuffaceous rocks containing labile uranium or other source material with uranium soluble under supergene conditions. (Some students of the genesis of this type of deposit suggest the source to be saline, residual brines, where uranium was concentrated after precipitation of more soluble Uranium is transported in aqueous solutions through permeable sandstone; the aquifer commonly exhibits a "plumbing" system and is bounded by impermeable beds. (The "brinal" theory postulates that uranium-bearing brines were expelled from the evaporites during burial and compaction and moved to overlying permeable sandstones (Dunsmore, 1975)). Host rocks are, as a rule, nonmarine sandstones with biogenic carbon. Reductants such as pyrite, H_2S, petroleum, humic acids and bitumens, cause precipitation of uranium along redox fronts. Changes in the redox front, hydrodynamic conditions, tectonic movements, etc. can cause migration/accretion of mineralization. Its preservation can be due to a closed hydrodynamic system, a reducing environment, and other factors (Alder, 1974).

Uranium deposits in sandstones do not as yet contribute to Canada's uranium reserves. However, indications of this type of mineralization exist in some of the Phanerozoic basins containing continental sandstones, where geochemical conditions suitable for

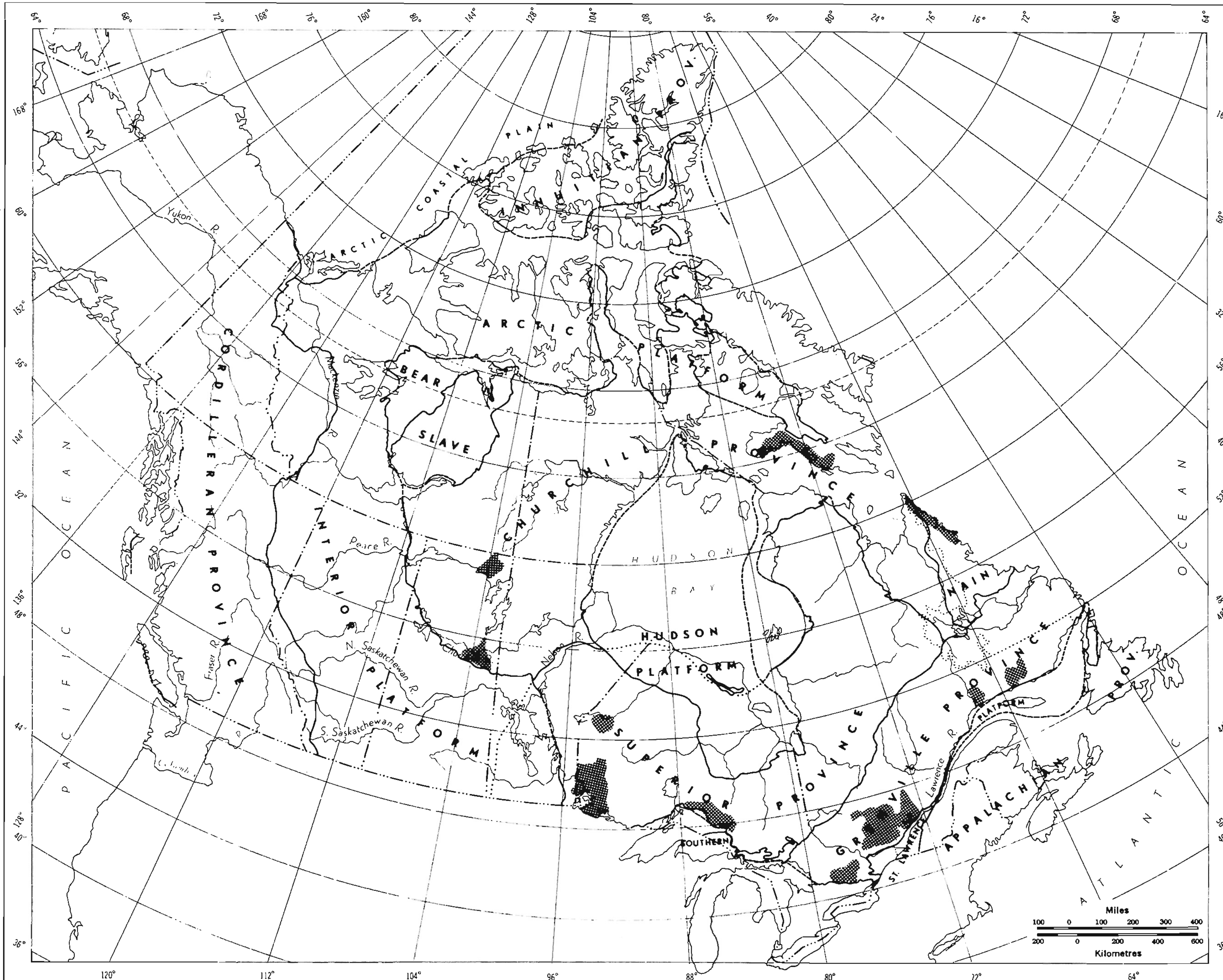

Figure 5. 7. Areas favourable for magmatic/anatectic and related uranium deposits in Canada.

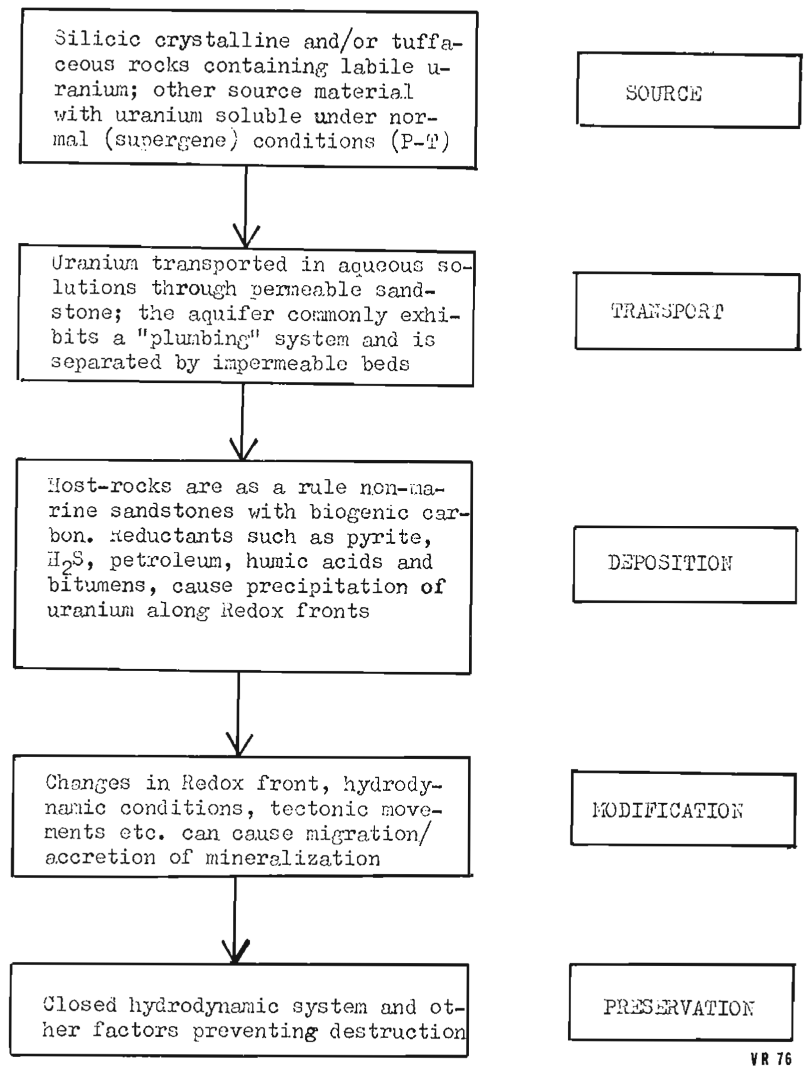

Figure 5. 8. A conceptual model simulating formation of Uranium Deposits in Sandstones.

generation of uranium-bearing fluids, redox fronts in permeable strata, and distinct lithological and/or structural traps favourable for deposition of uranium occur at present or occurred in the past.

Areas favourable for the sandstone type of uranium deposits in Canada are shown in Figure 5. 9.

The conceptual model simulating formation of supergene uranium deposits (Fig. 5. 10) postulates as source granites, tuffs or other rocks containing uranium in the ppm range. Uranium is transported in dissolved forms in surficial water or in groundwater.

Semi-consolidated materials (duricrusts: calcrete, gypcrete, ferricrete) filling depressions, basins and old stream channels are favourable for deposition of uranium mineralization. Changing climatic conditions and hydrodynamic factors can modify these deposits. Preservation of deposits is controlled by a delicate balance between mobility-stability equilibria of uranium minerals (Premoli, 1976). Calcrete deposits are known in Western Australia (Yeelirrie area), South-West Africa and the Somali Tertiary basin in northeast Africa. The deposits in South-West Africa contain large tonnages of U_3O_8. In Canada similar climatic and geological conditions may have occurred and the existence of fossil calcretes should be considered in exploration of areas where basement rocks have high uranium background values.

The conceptual models simulating formation of the above mentioned types of uranium deposits and the distribution of areas favourable for these genetic types in Canada are subjects of studies at the Geological Survey of Canada. Results of these studies should serve as a basis not only for assessment of Canada's uranium endowment, but also for development of criteria applicable to uranium exploration in this country.

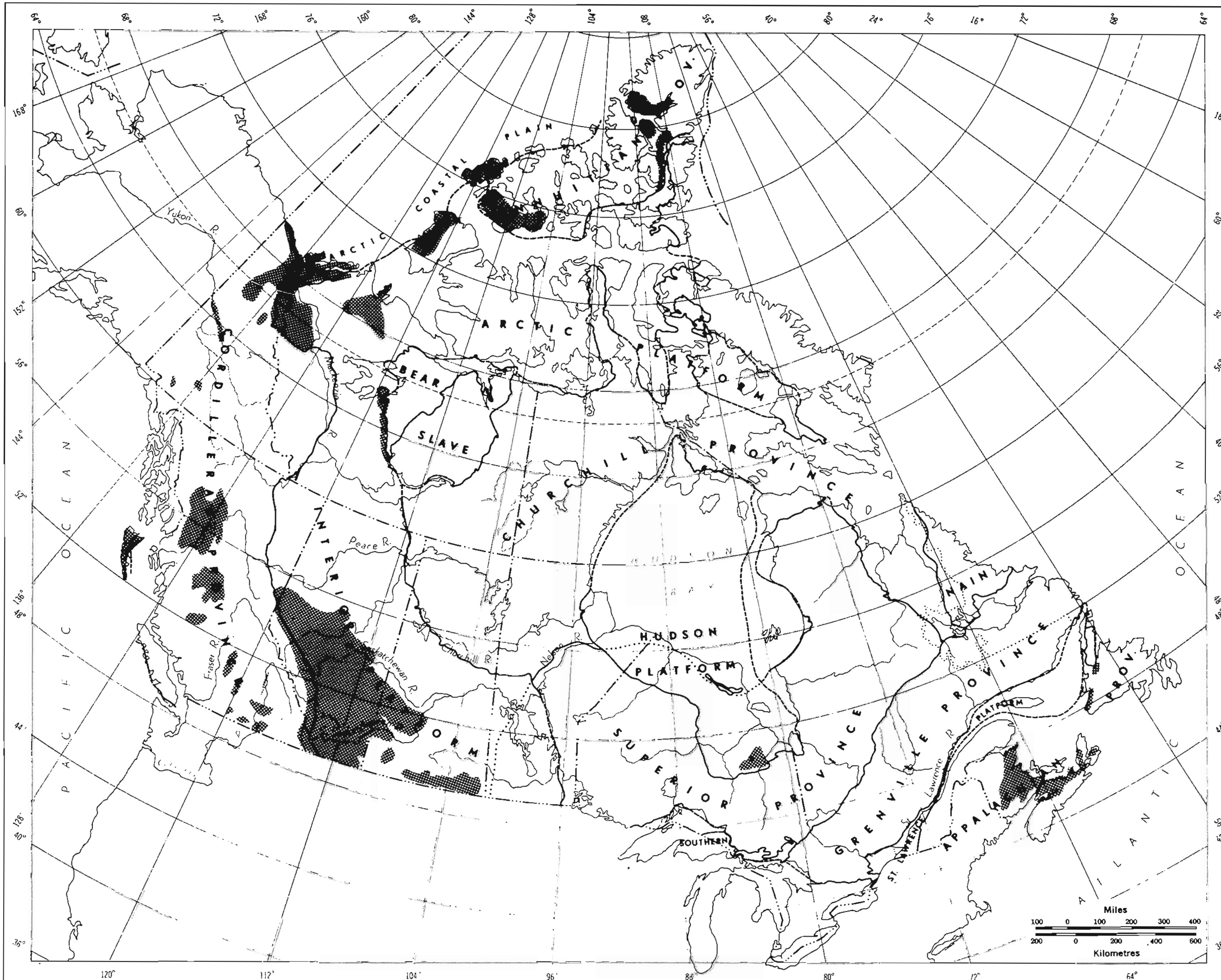

Figure 5.9. Environments favourable for sandstone uranium deposits in Canada.

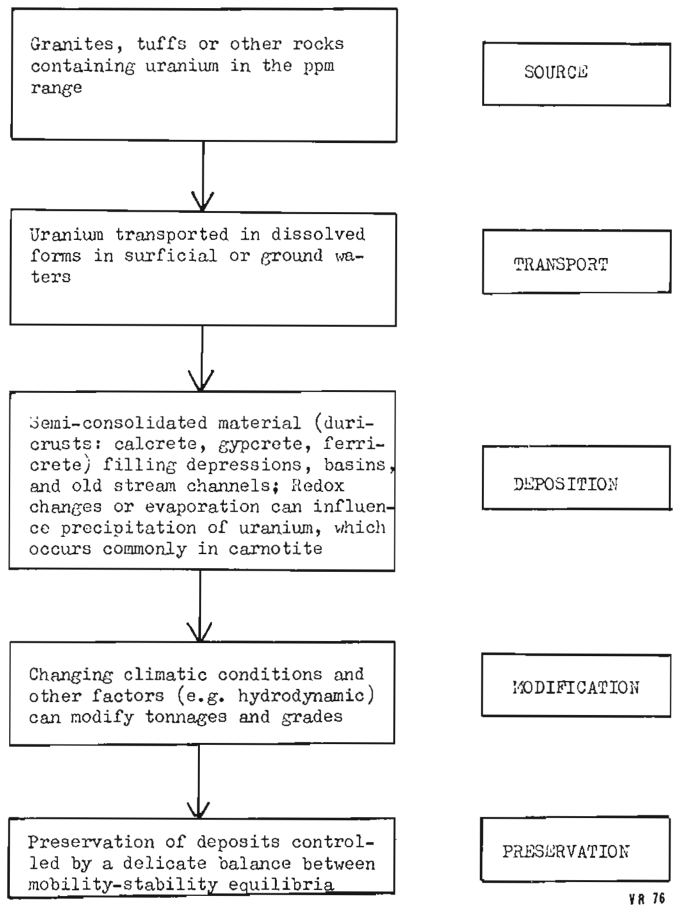

Figure 5.10. A conceptual model simulating formation of supergene Uranium Deposits.

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6. ASSESSMENT OF SELECTED URANIUM OCCURRENCES AND AREAS FAVOURABLE FOR URANIUM MINERALIZATION IN CANADA

Project 750010

V. Ruzicka
Regional and Economic Geology Division

Members of the Uranium Resource Evaluation Section were engaged in both field and laboratory studies of selected uranium occurrences and areas favourable for uranium mineralization in Canada during the spring and summer of 1976. L.P. Tremblay studied uranium occurrences in the Shield areas in Saskatchewan and Quebec; R.T. Bell occurrences in some Phanerozoic basins in the Cordillera and Interior Platform and in some Precambrian complexes in the Yukon; H.E. Dunsmore studied occurrences in Phanerozoic basins in the Atlantic Provinces; A. Miller and M.A. Turay examined selected Shield areas in the Northwest Territories; J.Y.H. Rimsaite studied mineralogy of the Rabbit Lake deposit. The writer assessed uranium occurrences in Ontario, Labrador, and participated in assessment of the Wollaston Lake Belt in Saskatchewan and of Precambrian complexes in the Yukon part of the Canadian Cordillera. N. Prasad and J.A. Kerswill, besides assisting in some of the above mentioned field operations, were engaged in data processing related to building a uranium commodity index (as part of CANMINDEX), and a computer based resource evaluation file (URE-3). Individual detailed reports, where data are not classified as industrially confidential, are included elsewhere in this volume. The studies have been conducted in close co-operation 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, with representatives of other Federal and Provincial organizations and with geologists of exploration and mining companies active in pertinent areas.

Huronian complexes in Ontario

In the Elliot Lake area results of exploration drilling, carried on by several companies, did not change substantially the existing knowledge of distribution of uranium mineralization.

The underground workings on the tenth level of the Quirke II Mine intersected dark grey sheared argillaceous rocks with scattered quartz pebbles, fragments of chert and with layered 'buckshot' pyrite. This rock (not identified before) occurs beneath the basal reef, but above the basement diorite. It appears that it was deposited in a local depression. Study of this rock is underway.

Distribution of Huronian volcanics penecontemporaneous with uraniferous sediments of the Matinenda Formation is the subject of another study. A suite of drill-core samples of Huronian and pre-Huronian volcanics is presently being geochemically and petrologically analyzed in order to find criteria for discrimination between these two ages of volcanic rocks. Solution of this problem could influence estimates of undiscovered uranium resources and consequently exploration strategy in the Elliot Lake basin and elsewhere.

Examination of the Aweres Formation north of Aweres Lake, on highway 556 approximately 1.25 miles north of the turnoff from highway 17, in an area which reportedly contained uranium mineralization (E. Leahy, pers. comm.) revealed that the radioactivity (*in situ*, 2π geometry on TV-1; T_1 = 10 000 cpm, T_2 = 200 cpm, T_3 = 20 cpm) is apparently related to a fracture filling rather than to syngenetic mineralization in conglomerates.

The Huronian complexes between Spanish and Sudbury were targets for uranium prospecting by private prospectors and several exploration companies. However, except for a few additions, mainly in extensions of known occurrences in the broader Agnew Lake area, no substantial changes in the uranium resource estimates are warranted there (Energy, Mines and Resources, 1976).

In the Cobalt Embayment, Aggressive Mining Ltd. explored in Turner Township and Hollinger Mines Ltd. explored in McLennan Township west of Lake Wanapitei. The exploration drilling in Turner Township confirmed the presence of uranium mineralization in argillite beds rather than in conglomerates.

Pre- and post-Huronian complexes in Ontario

Assessment of uranium resources in Ontario, outside of the Huronian complexes, was concentrated mainly in areas with (a) granitic and anatectic, (b) vein and related, and (c) metasomatic, volcanogenic and minor types of uranium mineralization.

(a) The granitic and anatectic type of mineralization was investigated in the Grenville Structural Province, and in the English River and Quetico Gneiss belts. In all of these areas possibilities for enlargement of low grade uranium reserves still exist.

(b) Assessment of vein and related deposits took place in the Montreal River and Lake Nipigon areas. The spatial relationship of uranium mineralization to unconformities appears to be an important factor in localization of uranium in both areas.

In the Montreal River area, zones of kaolinization in granitic rocks were observed in the vicinity of Theano Point-Casey Lake occurrences distributed mainly along diabase dykes. The possibility of a relationship needs further study.

In the Nipigon area, fragments of rocks presumably of the Sibley Formation, were identified by J.M. Franklin of the Geological Survey in faults mineralized by uranium in basement rocks. Uranium mineralization in the Sibley Formation, approximately 10 m above the pre-Sibley unconformity reported previously (Ruzicka, 1976) was identified by H.R. Steacy and A.G. Plant as coffinite (Fig. 6.1). The coffinite rims and replaces pyrite grains and also occurs among these as very irregular masses.

From: Report of Activities, Part A;
Geol. Surv. Can., Paper 77-1A (1977).

(c) Uranium mineralization in Timiskaming sediments and igneous rocks were studied *in situ* in the Kirkland Lake-Larder Lake area. It was found that the radioactivity in the Timiskaming polymictic conglomerates occurs mainly in spots with carbonate veining and therefore is likely of epigenetic origin. Higher gamma-ray spectrometric readings (on TV-1, 2π geometry: T_1 = 85 000 cpm, T_2 = 2250 cpm, T_3 = 65 cpm) were obtained on tuffaceous greywacke. However the highest radioactivity (on TV-1, 2π geometry: T_1 = 100 000 cpm, T_2 = 3000 cpm and T_3 = 100 cpm) was found in carbonate-bearing trachyte. It appears that the trachyte-syenite contacts are the most favourable loci for uranium mineralization in the Kirkland Lake-Larder Lake area.

Uranium-bearing areas in Labrador

Recent studies on the relationship between host rocks and the density of uranium mineralization in the Makkovik-Kaipokok Bay area showed, that despite

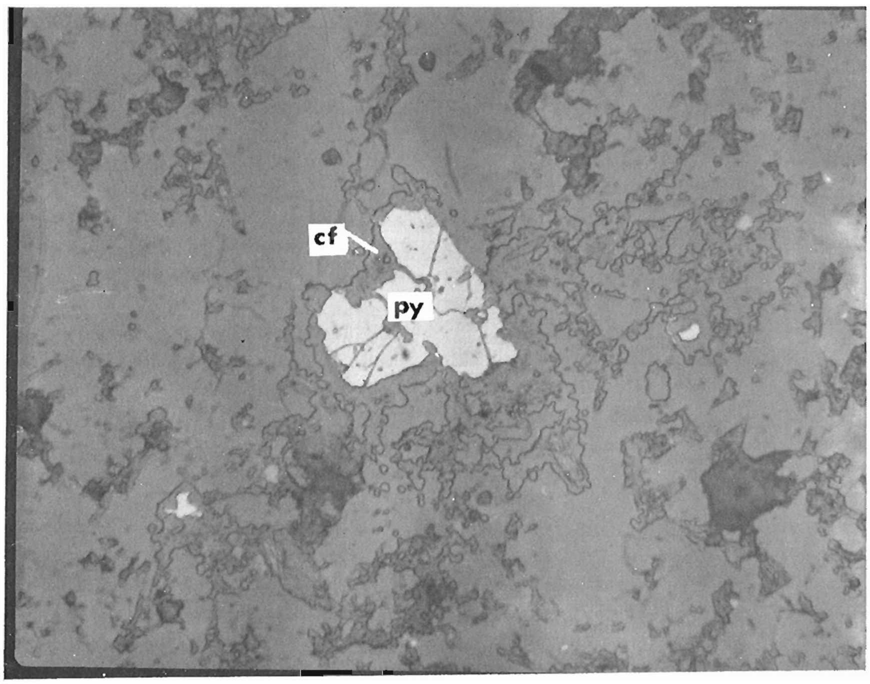

Figure 6.1.

Photomicrograph of a polished section of a mineralized portion of Sibley Formation near Nipigon, Ontario; py = pyrite, cf = coffinite; X 650.

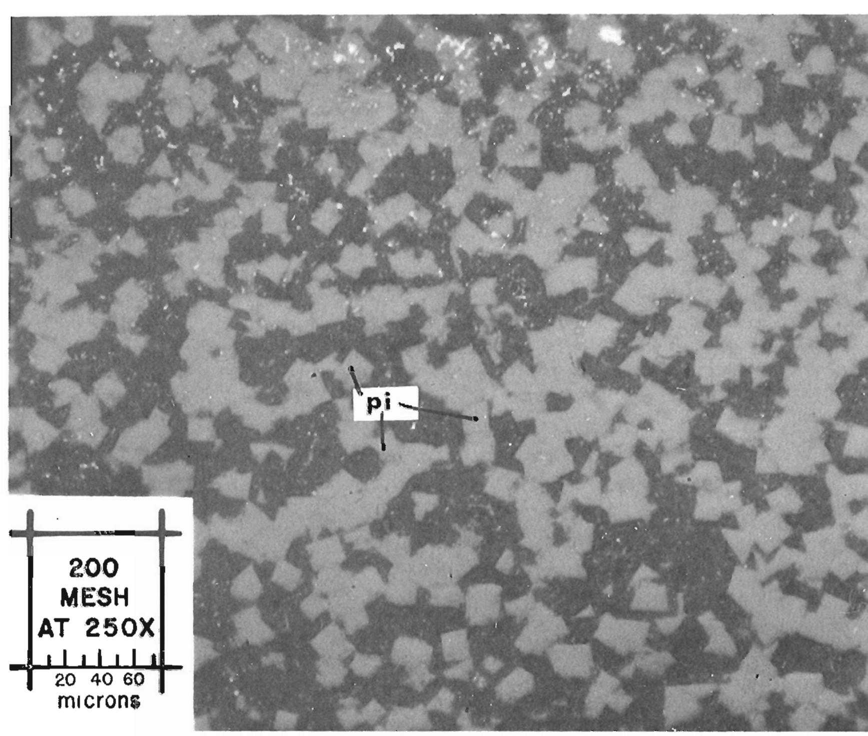

Figure 6.2.

Euhedral pitchblende ("urania") (pi) from the Deilmann orebody, Key Lake deposit, Saskatchewan; reflected light.

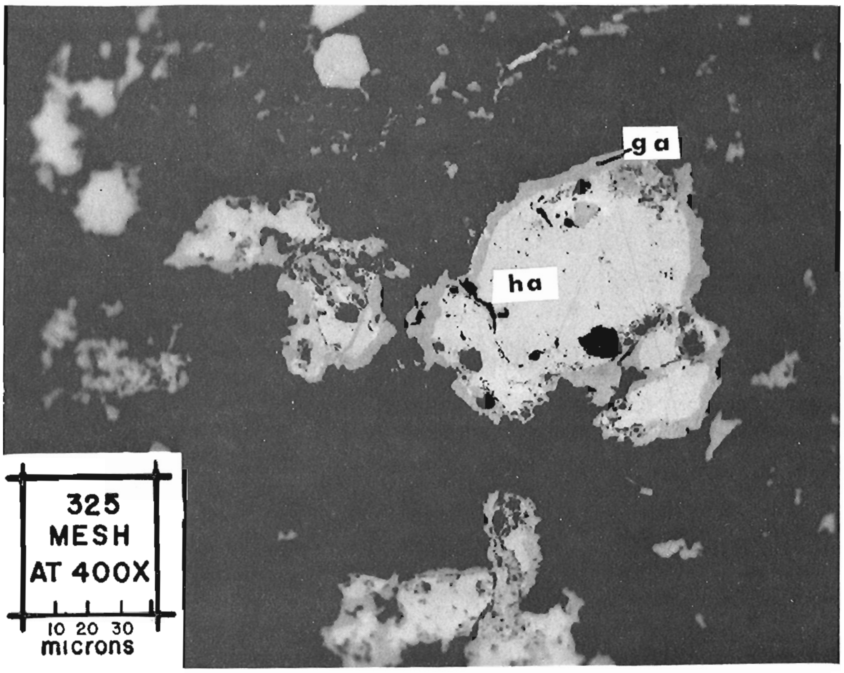

Figure 6.3

Hauchecornite (ha) rimmed by galena (ga) from a high grade uranium ore of the Gärtner orebody, Key Lake deposit, Saskatchewan.

different geological environments and despite different types of uranium mineralization in the Kitts and Michelin deposits, some similarities exist in the mineralization density.

The uranium mineralization in the Kitts belt is confined mainly to a tuffaceous argillite horizon, 10-25 feet thick, which is separated from the underlying 'lower amphibolite' by a cherty quartzite and garnet-hornblende layers and from the overlying 'upper amphibolite' by a narrow band of tremolitic calcareous shale (S. Gandhi, pers. comm.). In the Michelin belt the uraniferous horizon consists of sheared quartzo-feldspathic rocks apparently derived from rhyolitic tuffs or flows, exhibiting hematitic alteration along radioactive zones. Whereas the mineralization in the Kitts deposit occurs in individual lenses of pitchblende, the uranium in the Michelin zone is finely dispersed. It was found, however, that the amount of uranium calculated to occur in volumetrically comparable units is, despite different ore grades, almost identical in both belts. This problem is being studied further.

Examination of the Stormy Lake showing in the Seal Lake Synclinorium (Marten and Smyth, 1975) showed that the radioactivity is confined to sheared quartz-pebble conglomerate and is, apparently, of epigenetic origin.

Some features of mineralization in the Key Lake deposit, Wollaston Lake Fold Belt

Studies on two selected drill-core samples collected from the Gärtner and Deilmann orebodies of the Key Lake deposit revealed mineralogy that is to a certain degree similar to that in the mineralized boulders examined by Watkinson *et al.* (1975).

In addition, euhedral pitchblende (*see* Fig. 6.2) was identified by S. Kaiman of CANMET in the sample from the Deilmann orebody. This very rare form of pitchblende resembles that described as urania by Robinson (1955) from Nicholson No. 4, Fish Hook and Gunnar properties in the Beaverlodge area, Saskatchewan. Furthermore, in the samples from the Gärtner orebody maucherite ($Ni_{11}As_8$) was found closely associated with massive pitchblende. Hauchecornite ($Ni_9Bi_2S_3$) exhibits a rim of galena in the same sample (*see* Fig. 6.3). However, the most common nickel mineral associated with pitchblende in the sample from the Gärtner orebody is gersdorffite (NiAsS).

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GEOLOGY OF SOME URANIUM OCCURRENCES IN WESTERN CANADA

Project 740069

R. T. Bell
Regional and Economic Geology Division

Systematic investigations of uranium deposits, particularly those related to sandstones, were continued in Western Canada during 1976. Uranium occurrences in the Yukon are reported in this publication separately (Bell and Delaney, 1977). Occurrences in British Columbia have been tabulated recently by P. Christopher (1976) to whom I am indebted for his kind co-operation.

R. Anderson and A. Boyer conducted a regional emanometer survey of an 85 square miles area in eastern Cypress Hills north of Eastend, Saskatchewan. At stations with one quarter mile spacing, radon gas was measured in soils overlying five Tertiary and Upper Cretaceous clastic units known to contain uraniferous lignite and fossil bones (Cameron and Birmingham, 1970; Bell, 1975). A radiometric survey of this area was also completed by car this summer.

A brief description of some new occurrences in British Columbia and Alberta follows. A McPhar TV-1 three-threshold scintillometer was used in the field for on-site estimation of U and Th content (in ppm).

Jasper Park, Alberta

About 3. 2 km east of Geike Creek on the Yellowhead highway west of Jasper, the Miette Group of grey quartzose gritty sandstone, conglomerate and light grey shale showed 3 to 4 times background radiation. At one site (52°52'00"N, 118°12'30"W) silty shales showed 7 times background. *In situ* measurement suggests the anomaly is due to U rather than Th.

Bowron River, British Columbia

Thin seams (near old coal mine at 53°49'50"N, 121°55'00"W) and associated fine grained sediments in a 2-m zone about 1 m beneath the main coal seams showed anomalous radioactivity (up to 10 times background) due to U rather than Th. In contrast the main coal seams (Cameron and Birmingham, 1970) are only very weakly radioactive.

Clearwater River, British Columbia

(a) Raft Batholith. The Raft Batholith (unit 20, Campbell and Tipper, 1971) a porphyritic quartz monzonite, near the contact with Tertiary plateau lavas (unit 25, *ibid.*) shows anomalous radioactivity (3 to 5 times background). Several localities associated with thin pegmatite dykes, gouge zones, and felsite dykes show in excess of 10 times background radiation. The most notable is in a thin (5- to 8-cm-thick) pegmatite about 6. 3 km north of the bridge over Clearwater River, on the road on the west side of the river. *In situ* measurements suggest about 400 ppm U and 800 ppm Th.

Other granites (unit 14, *ibid.*), notably along Eakin Creek, show 2 to 4 times background radioactivity.

(b) Joseph Creek. Tertiary boulder conglomerate and associated fine grained sediments (unit 21, *ibid.*) underlying volcanics (unit 22, *ibid.*) just east of the bridge over Joseph Creek (51°27'38"N, 120°08'15"W) near Dunn Lake show anomalous background radiation throughout (5 to 7 times). These sediments are about 13 km north of the coaly sediments at Newhykulston Creek reported on by Cameron and Birmingham (1970).

Kelowna, British Columbia

Tertiary sediments and volcanics and underlying basement rocks were inspected in the region near Kelowna. Near Kallis Creek (49°46'51"N, 119°04'24"W) about 2 m of gravels with about 10 cm of sands and silts at the top are overlain by plateau lavas. These sands and silts are irregularly radioactive for 1 to 4 m along the contact at several localities (10 to 16 times background radiation). The underlying granitic rocks are deeply weathered, bleached, and in places rusty and show spotty radioactivity (up to 10 times background) as much as 10 m below the contact.

Okanagan Falls, British Columbia

(a) White Lake Basin. Radioactive anomalies (about 10 times background) were investigated in the White River basin (Church, 1973). Notable sites are in the Park Rill Member (49°17'07"N, 119°37'38"W), Kitley Lake Member (49°22'53"N, 119°38'10"W) and in both Kitley Lake and Yellow Lake members (49°22'52"N, 119°38'10"W), all of the Marron Formation, and in volcanic members (49°19'26"N, 119°37'14"W) of the White Lake Formation. These anomalies, however, appear to be entirely due to thorium. Sediments of the White Lake Formation were investigated and proven to be barren.

(b) Allandale Lake, British Columbia. A pegmatite dyke (49°23'35"N, 119°21'03"W) in an intrusive plug (Little, 1961) contains betafite, cyrtolite, and perhaps brannerite and euxenite in association with magnetite and hematite. The porphyritic syenite host (Coryell) shows somewhat higher than normal radioactivity.

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Project 750069

R. T. Bell and G. D. Delaney[1]
Regional and Economic Geology Division

Introduction

Emphasis in field work in 1976 on uranium in the northern Cordillera was on the Wernecke Mountains, where uranium occurrences had already been reported (Blusson, 1976).

In light of the types of uranium occurrences described herein, it will be necessary to redefine the stratigraphic units and it will also be necessary to clarify the provenance and genesis of these units.

Perhaps the most spectacular element in this area is the hitherto unrecognized intrusive breccias (Figs. 8.1 to 8.5) within the lower parts of the Precambrian section; these breccias may relate to volcanic centres within it. More importantly, uranium and copper occurrences in the area relate directly to both intrusive breccias and to the volcanic centres.

Precambrian Stratigraphy

The stratigraphy and distribution of the lowermost Proterozoic units (Helikian, perhaps Aphebian) in Wind River, Snake River, Nash Creek and Nadaleen River map-areas (Norris, 1975; Green, 1972, Blusson, 1974) are tentatively revised. The schematic section (Fig. 8.2) based on seven measured sections and cursory inspections of several more, summarizes the lithologic character of the strata in the area. The 14 units may be grouped into three sequences (groups) roughly equivalent to Green's (*ibid.*) unit 1a, the remainder of unit 1 and unit 2.

The lowermost group (A), exceeding 1500 m thickness, outcrops along the Bonnet Plume River from Fairchild Lake (64°55'N, 133°45'W) to Kiwi Lake (65°15'N, 134°40'W) in a band about 9 km wide. It is typified by argillite, siliceous carbonates, calcareous siltstones, lustrous and spotted phyllites, dolomite and minor limestone and quartzite. Some of the fine grained rocks may be of volcanic origin.

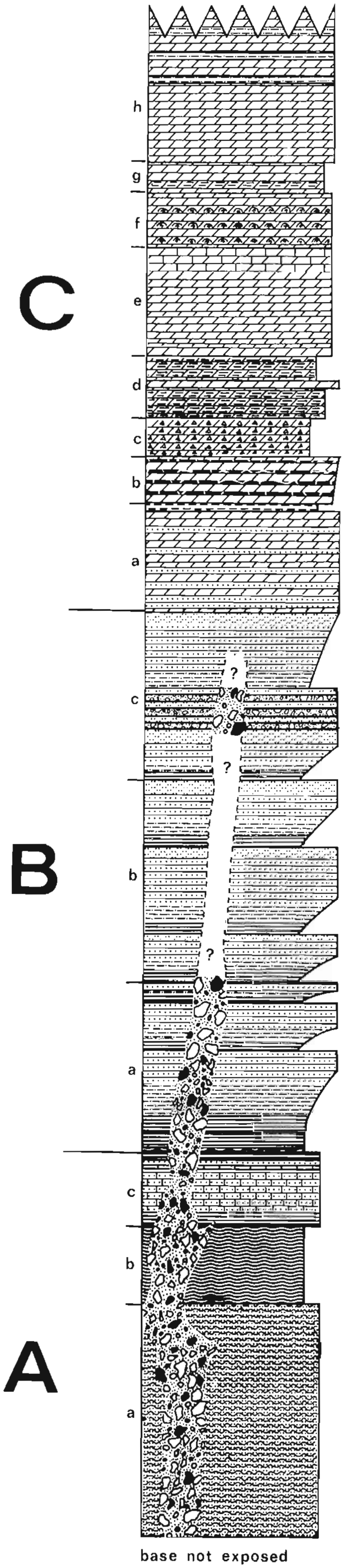
The middle group (B) comprises dark grey to black mudstones, siltstones and quartzose sandstones, more than 2100 m thick and perhaps as much as 4500 m thick. The lower part (Ba) is pyritic, silty mudstone with coarsening-upwards cycles of mudstone-siltstone-sandstone and with a gradual increase in the proportion of the coarser part up-section. The medial part (Bb) is largely a dark grey sandstone sequence, containing slump structures and normally graded sandstone beds. Slump breccias and felsic (reworked?) volcanic breccias associated with dark grey to green mudstones and graded sandstones (Bc) occur near the top of the succession. Some breccias are strongly heterolithic but most contain mainly one or two types of clasts in any one bed. The clasts are predominantly argillite, carbonate-argillite, chert, siltstone (tuffaceous?) and felsite. Matrix is fine argillaceous to tuffaceous material. Most of these breccias are clearly stratiform and commonly graded; some can be traced directly into flow and pyroclastic centres (e.g. at 64°40'N, 134°54'W) where indeed some may be discordant (gas vent breccias). Volcanic rocks may occur in the lower part of this group (B) but none were observed.

Group B is gradational to group C which comprises at least eight (Fig. 8.2) locally mappable units. Lateral facies changes are significant, for example unit Cf probably thickens substantially towards the southwest. Further description will follow at a later date.

Figure 8.1. View looking northeast at Slab Mountain showing castellate weathering of breccia (D) cutting unit Ac; note large (100 m range) breccia blocks at X; note small breccia pipe (d) in centre of photo cutting unit Ab (GSC 2031031).

[1]Department of Geology, University of Western Ontario, London, Ontario.

From: Report of Activities, Part A;
Geol. Surv. Can., Paper 77-1A (1977)



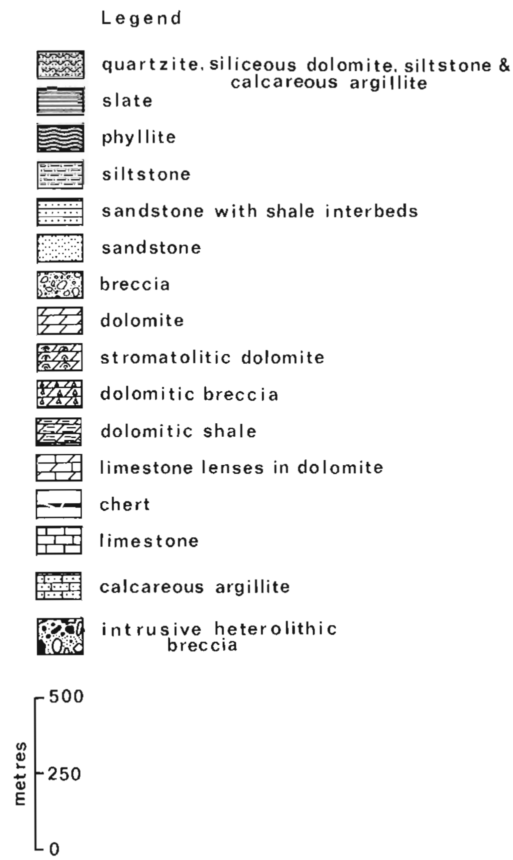


Figure 8. 2. Proterozoic stratigraphy in the vicinity of Fairchild Lake, Yukon Territory.

Paleocurrent data are insufficient to draw important conclusions. Preliminary work on group B suggests sediment transport from the northeast and north.

Intrusive Heterolithic Breccias

Large elongate discordant bodies of heterolithic breccias (Figs. 8. 1 to 8. 5) occur in the zone along Bonnet Plume River where they cut group A and the lower part of group B. At several other localities lithologically similar heterolithic breccias are associated with stratiform, much less heterolithic, volcanic breccias (Bc). None were observed cutting unit C. Dip separation is apparent between the walls of some breccias. The largest breccia bodies occur near Quartet Lakes (65°60'N, 134°23'W) and at Slab Mountain (65°00'N, 134°00'W) Fig. 8. 1, about 4 km by 0. 5 to 1 km. The bodies follow two trends, roughly north-northeast and west-northwest.

Fragments in the breccias range from at least 100 m by 40 m down to finely comminuted matrix. Usually the larger blocks (>2 m) are carbonate or silicic carbonate whereas the smaller fragments represent most lithologies observed within unit A. Fragments of

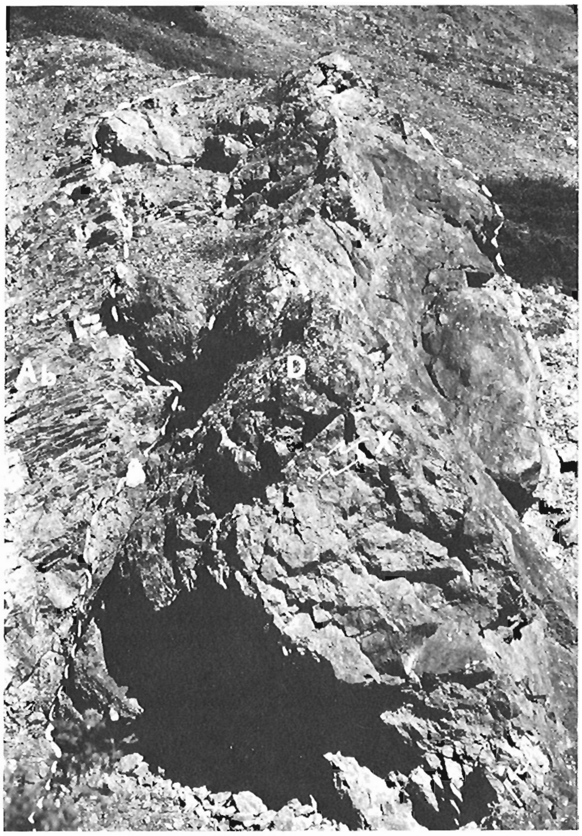


Figure 8. 3. View looking north at the small breccia pipe on Slab Mountain shown in Figure 8. 1; examples of fragments indicated at X; bedding in unit Ab dips gently to the left and to the northwest; prospector's pick for scale in lower right (GSC 203103-H).

felsite (?), greenstone, and metadiorite are present but rare. Perhaps an earlier and lower igneous complex may be inferred.

Breccias and adjacent country rock are commonly carbonated. Disseminated coarse magnetite, specular hematite, ferrous dolomite, and siderite and finely disseminated barite are common. Some barite-magnetite veins are present. Locally the breccias and adjacent wall rocks are silicified and feldspathized and contain uranium mineralization. Cu and Co mineralization (commonly observed as secondary minerals) are also present. The breccias and adjacent carbonated country rock weather with a spectacular castellate form (Figs. 8. 1, 8. 4).

Uranium Occurrences

Uranium (and copper) mineralization is associated with both the largely stratiform breccias of unit Bc (designated VB style) and the discordant exotic breccias along the Bonnet Plume River (designated BP style). It is noteworthy that reports of uranium mineralization in breccias have been very rare in Canada (an exception: Reinhardt, 1972). The occurrences in Wernecke Mountains are listed in Table 8. 1.

Radiometric levels in units A and B are anomalously high, some levels as great as 6 times background. These strata, particularly dark shales and clastic rocks, may be a source for the mineralization associated with the breccias.

Erosion of equivalent strata in the Wernecke and perhaps the Ogilvie Mountains may have contributed uranium to successor basins such as the Bonnet Plume Basin. Metamorphic remobilization of uranium in equivalent Precambrian strata in adjacent terrains could be significant.

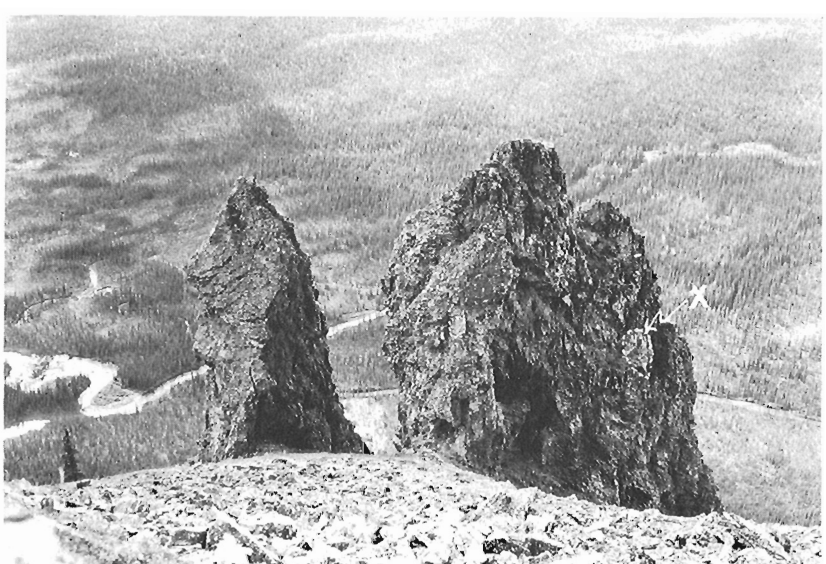


Figure 8. 4

Castellate weathering of breccia near Quartet Lakes; examples of larger fragments indicated; the spires are 15 to 20 m high (GSC 203103-G).

Table 8.1

Uranium Occurrences — Wernecke Mountains

NTS (1)	No. (2)	Long. W. (3)	Lat. N. (4)	Approx. elev. feet (5)	Nature of site (6)	Nature of host (7)	Nature of min. (8)	Mineralogy (9)	Style of occ. (10)	Associated elements or minerals (11)
106D/10	1a	134°56'00"	64°39'55"	4100[1]	a, b	c, e, f, g	a, b	a	VP	Ba, F-S
"	1b	134 55 40	64 40 10	4050	f	c, h	a, b	a	VP	Ba, F-S
"	2a	134 54 10	64 39 45	4000	a, b	c	a	a, e?	VP	Ba, F-S
"	2b	134 54 35	64 39 35	4000	a	e	b	a	VP	Cu, F-S
"	3	134 44 00	64 40 00	5100	b	c, d, f	a, b	a	VP	Cu, Ba, F-S
106D/16	4	134 16 20	64 56 22	4900	b, e	a, b	c	b	BP	--
"	5	134 25 00	64 59 01	5200	c	f	a	a	VP	F-S
"	6	134 25 06	64 59 18	4100	a, c	c, d	a	a	VP	--
"	7	134 02 00	64 59 18	1875	e	a	a, b	a	BP	--
"	8a	134 01 00	64 59 45	3100	c	h	a	a	BP	Cu, F-S
"	8b	134 00 55	64 59 35	2800	c	h	a	a	BP	F-S
"	8c	134 00 55	64 59 30	2600	c	a?	a	a	BP	F-S
"	9a	134 01 14	64 59 59	4150	a	a	a	a	BP	--
"	9b	134 01 03	64 59 57	4000	a	a	a, c	b?	BP	Cu, Ba, F-S
106C/13	10a	134 56 54	64 59 30	3400	d	a, b	a, c	a	BP	Cu
"	10b	133 57 42	64 59 30	4300	c	a	a	a	BP	--
106C/14	11a	133 17 57	64 57 15	5750	d	c, d, e	a, b	c, e	VP	Cu, F-S, Co
"	11b	133 17 36	64 57 15	5950	a	e	a, b	e	VP	F-S
"	11c	133 16 48	64 57 57	6000	c	c	a	a	VP	--
"	11d	133 18 06	64 56 43	5600	c	c, e, h	a	a	VP	--
106E/1	12	134 14 50	65 04 20	2600	c	f	b	b	BP	Cu, F-S
"	13	134 14 40	65 04 30	2900	a	a, b, e	a	a	BP	Cu, Co
"	14	134 23 20	65 06 05	3200	c	a	a	a	BP	--
"	15	134 23 30	65 06 20	3650	a	a	a, c	b	BP	Ba, F-O
"	16	134 23 20	65 07 55	4000	a	b, e	a	a	BP	Cu
"	17	134 23 20	65 08 00	3700	a	a	a	a	BP	--
"	18	134 24 00	65 08 25	3900	d	a, b, e	a, b, c,	b, c, d?, e	BP	Cu, Ba, Au
106E/2	19	134 32 30	65 05 15	3600	d	f, h	a	a	BP?	Cu, F-O
"	20a	134 38 20	65 02 30	4000	a, d	c, d?	a, b, c	c, e	VP	Cu, Ba, F-O, F-S
"	20b	134 39 00	65 02 35	3600	f	d	a	a	VP	Cu, Ba, F-O, F-S
"	21	134 39 30	65 02 15	3800	d	c, d	a	a	VP	F-S
106E/1	22	134 26 30	65 10 35	4250	a, c	a	a, b	a	BP	--

Column 6: Nature of site: (a) in outcrop; (b) in deeply weathered outcrop; (c) 1 to 3 samples in scree, rock glacier, slump blocks; (d) several samples in rock glacier, slump blocks; (e) 1 to 3 samples as cobbles in stream bed; (f) several samples as cobbles in stream bed.

Column 7: Nature of host: (a) intrusive, exotic breccia; (b) altered host of breccia; (c) stratiform homolithic volcanic and slide breccias; (d) stratiform, heterolithic, volcanic and slide breccias; (e) stratiform siliceous and carbonate rocks; (f) sandstone, siltstone, shale; (g) dyke; (h) other.

Column 8: Nature of mineralization: (a) finely disseminated; (b) vein; (c) coarsely disseminated.

Column 9: Mineralogy: (a) unknown; (b) brannerite; (c) pitchblende; (d) uranothorite; (e) secondary minerals; (?) questionable.

Column 11: Associated elements or minerals: Cu: copper minerals, mainly chalcopyrite and secondary; Ba: barite; F-O: iron oxides; F-S: iron sulphides; Co: cobaltite; Au: gold.

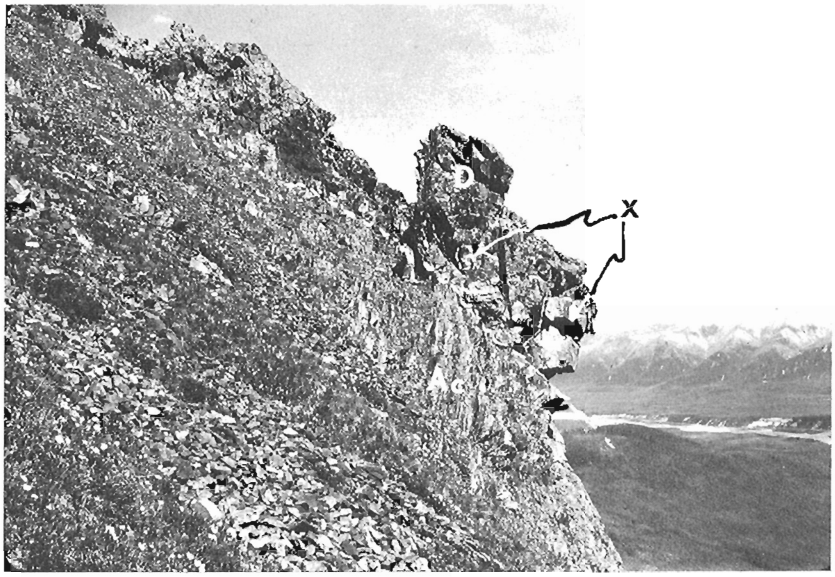


Figure 8.5

Footwall (Ac) of inclined breccia dyke (D) near Quartet Lakes; examples of larger blocks shown at X; the thin bedded dolomitic siltstone dips 70°SE into the hill (GSC 203103-I).

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9. PRELIMINARY EVALUATION OF STRUCTURAL AND GROUNDWATER CONDITIONS IN UNDERGROUND MINES AND EXCAVATIONS

Project 740057

K.G. Raven
Terrain Sciences Division

Introduction

A preliminary review of the structural and groundwater conditions in operating mines and large underground civil engineering projects within the Canadian Shield has recently been completed. This review is a supporting study to a joint project of Atomic Energy of Canada Limited and Geological Survey of Canada (Gale *et al.*, 1976). The primary objective of the project, in conjunction with other branches of the Department of Energy, Mines and Resources, is to determine if igneous or high-grade metamorphic crystalline rocks are suitable host rocks for the storage of solidified radioactive wastes. The main objectives of this study of underground mines and excavations are:

1) to determine the current use and effectiveness of aerial surveys, surface mapping, and diamond-drilling in delineating and predicting subsurface structural and groundwater conditions;

2) to assess the variation of groundwater conditions with depth; and

3) to investigate the relationship (if any) between structural and groundwater conditions and thus to determine if it is possible to correlate the surface and subsurface characteristics of these features and to predict, before excavation, subsurface conditions from surface measurements.

The results of these studies should improve our understanding of how structural and groundwater conditions vary with depth and provide fundamental criteria, which will aid in the evaluation of sites mapped during the 1976 field program and aid in the selection and evaluation of sites for future mapping programs.

Scope

Work to date has involved two closely related studies: (1) a study of the structural and groundwater conditions of large underground excavations and (2) a similar study at a number of mines in Ontario and Quebec.

Four hydroelectric projects (Bersimis No. 2, Chutes-des-Passes, Churchill Falls, and Gull Island) located within crystalline rocks in Labrador and Quebec have been evaluated. The location of these sites is shown in Figure 9.1. A major dewatering project of an existing development (Nilo Peçanha) situated in the Brazilian Shield is the fifth site studied.

The review of structural and groundwater conditions in underground mines involved three components: (1) reconnaissance visits to nine mines located within the Abitibi Greenstone Belt; (2) the distribution of an extensive questionnaire to all operating mines in Manitoba, Ontario, and Quebec and the assessment of the returns; and (3) a detailed investigation of the surface and subsurface structural and groundwater conditions

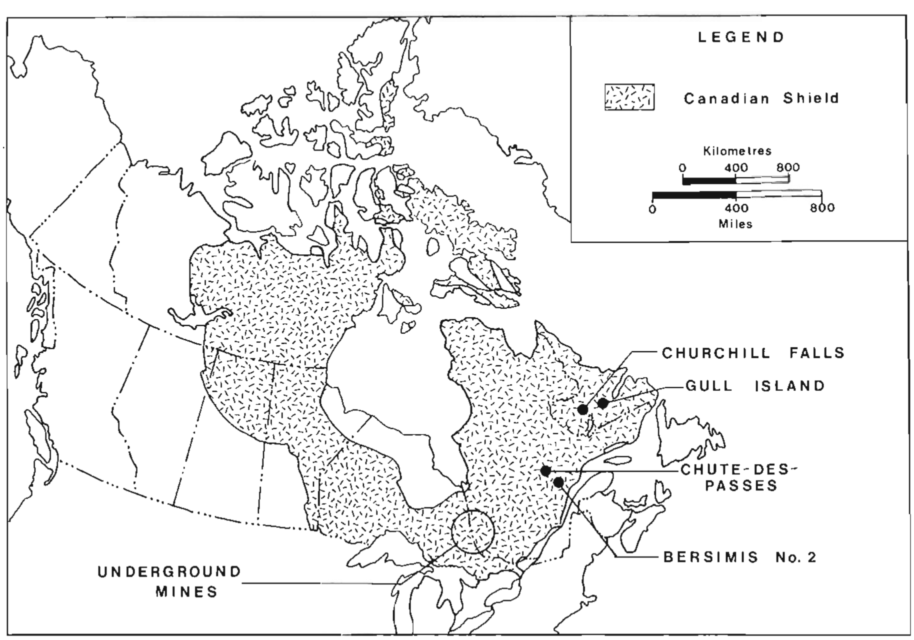

Figure 9.1.

Location of underground mines and excavations studied.

From: Report of Activities, Part A;
Geol. Surv. Can., Paper 77-1A (1977).

at three mining sites (Geco mine, Manitouwadge, Ontario; Millenbach mine, Lac Duffault, Quebec; and Fecunis, North, Levack Strathcona, and Longvack mines, Levack, Ontario) during the 1976 field season.

This report summarizes the preliminary findings of the study of underground excavations and the initial mine reconnaissance.

Underground excavations

Geology

In general all five sites are situated within bedrock composed of Precambrian gneiss having faint to distinct foliation. All four Canadian projects are located in the southeastern exposed part of the Canadian Shield in Grenville Province. This area consists largely of quartzofeldspathic basement rocks which have been exposed through the erosion of cover rocks. The uplift and erosion that occurred both during and after the Precambrian have produced a peneplain which today is transected by numerous deep river valleys.

Bedrock exposures at four of the five sites are minimal, representing possibly 10 to 15 per cent of the surface area. The Bersimis No. 2 site has about 30 per cent bedrock exposure. Both major (faults, shear zones) and minor (foliation, jointing) structures were encountered at the sites studied.

Summary of Findings

Evaluation of the case histories has demonstrated that a marked improvement in the delineation and prediction of subsurface features has occurred with time (1954 to 1975) and that much of this improvement is related to an improvement in the quality of geotechnical investigations. Increased use of aerial photograph analysis, water pressure testing, borehole photography, and geophysics, as well as more detailed core analysis and surface mapping, collectively have contributed to this improvement.

The use of aerial photolineament analysis has proven to be extremely helpful in detecting faults and shear zones which, in some places, were not recognized during surface mapping. Both local and regional scales of aerial imagery should be used in such an analysis.

Unpublished results of water pressure testing by Acres Consulting Service Limited and Lower Churchill Consultants indicated a general decrease of rock mass hydraulic conductivity or permeability with depth, although localized zones of high permeability also existed. These zones of high permeability commonly were associated with open joints, faults, breccia zones, and highly fractured and sheared dykes and sills. This water pressure testing also indicated that the major zones of leakage from surface to depths of 100 to 250 feet were flat-lying sheet joints. These joints, which occurred at most of the sites paralleling the rock surface, were commonly open, possibly as a result of thermal effects, creep, or stress relief due to erosional unloading.

All sites studied are located within river valleys and eroded slopes, regions where the effect of creep and stress relief on the opening of joints is assumed to be considerable. Also, because faults and shear zones control much of the drainage in the Canadian Shield, sites of hydroelectric potential may reflect a relatively high occurrence of these major structures. Thus when attempting to relate the subsurface conditions observed at these underground excavations to other areas of the Canadian Shield, it is important to consider that the excavation sites may represent a worse case in terms of groundwater inflow at depth.

This study has shown that in areas of the Canadian Shield with higher percentages of bedrock exposure, most of the major and minor structural features can be recognized at the surface using aerial photolineament analysis and detailed surface mapping and their existence can be projected to depths of 500 to 1000 feet (152 to 305 m). Since major zones of seepage were associated with structural features such as dykes, faults, and some shear zones, relative subsurface groundwater conditions may be inferred from the presence or absence of surface expression of these structural features. At the sites investigated surface expression of major fractures, including faults and shear zones, consists of linear depressions and/or zones of intensely sheared hematized rock. Where gneissic rocks exist, a plane of weakness occurs parallel to the foliation and consequently foliation shear zones should be anticipated (Benson *et al.*, 1974). These features were encountered at all sites.

Underground Mines

Geology

All the mines visited (Fig. 9.1) are located within the Abitibi Greenstone Belt — a linear zone of volcanic and sedimentary rocks 100 miles (160 km) in width extending from Timmins, Ontario to Val d'Or, Quebec. These rocks form the largest and probably the thickest Archean complex in the Superior Province. During the Kenoran orogeny these rocks were folded, faulted, metamorphozed, and intruded by granitic rocks. Major east-west trending fault zones and some minor late cross-faulting mark this belt.

For the purposes of this report, faults are defined as distinctive breaks characterized by a mud or clay seam, whereas shear zones are marked by wider zones of multiple shearing without a mud or clay seam.

Summary of Findings

A summary of the major seepage zones and groundwater conditions of each mine visited is presented in Table 9.1. This information was obtained by direct inspection of mine workings and discussions with mine personnel. The maximum depth of continuous seepage is the average depth below which there was no visible evidence of sustained groundwater inflow.

From a reconnaissance of structural and groundwater conditions it has become apparent that aside from geological controls, the degree of visible groundwater seepage at depth is a function of several other variables. Groundwater conditions appear to be influenced by:

1) ventilation (evaporation and removal of water directly from drift walls, etc. by circulating air);
2) duration of mine operation (local lowering of groundwater table and drainage of rock mass with time);
3) use of hydraulically emplaced fill (man-made seepage from filled stopes); and
4) nature and extent of surface workings (creation of downcast system for surface water).

In assessing the groundwater conditions observed on different mine levels it is necessary to consider these factors.

Table 9.1

Mine Study Summary

MINE AND DATE OF SHAFT SINKING	DEPTHS OF LEVELS feet (metres)	MAJOR SEEPAGE ZONES: DEPTH INTERSECTED feet (metres)	MAJOR SEEPAGE ZONES: ROCK TYPE	MAJOR SEEPAGE ZONES: NATURE OF ZONE	MAXIMUM DEPTH OF CONTINUOUS SEEPAGE feet (metres)
Camflo (1963)	450-3150 (137-960)	450-1300 (137-396)	Conglomerate	Fractures	2000 (610)
East Malartic (1937)	150-5145 (46-1569)	1000 (305)	Syenite Porphyry	Seam	1500 (457)
Sigma (1935)	100-5860 (30-1787)	225-725 (69-221)	Greenstone, Feldspar porphyry	Faults, Shears	1500 (457)
Chadbourne (1974)	100-950 (30-290)	100 (30)	Andesite, Breccia	Faults	>950 (290)
		950 (290)	Andesite	Faults	
Millenbach (1969)	2350-4018 (716-1225)	1100 (335)	Diorite	Fault	1800 (549)
Kerr Addison (1920)	0-6022 (0-1836)	300 (91)	Andesite	Fault	1500 (457)
		500-1150 (152-351)	Andesite	Joints	
Macassa (1917)	300-6876 (91-2096)	100-500 (30-152)	Conglomerate	Fractures	>3000 (915)
		2475, 3000 (755, 915)	Augite Syenite	Faults, Vugs	
Dome (1910)	0-5200 (0-1585)	0-1000 (0-305)	Conglomerate Quartz feldspar porphyry	Fractures	1000 (305)
Pamour Schumacher (1912)	100-7875	700-1625 (213-495)	Quartz feldspar porphyry	Faults	1800 (549)

In general it has been possible to relate, on an empirical basis, groundwater inflow to depth and geologic structures. From surface to about 1000 feet (305 m) below ground level (B. G. L.) seepage was associated with both major structures (faults and to a lesser extent shear zones) and minor structures (jointing and foliation). Flow within this zone along joint planes is probably a function of weathering and overburden stresses.

From 1000 to 2000 feet (305 to 610 m) B. G. L. seepage was reduced substantially and where present commonly was associated with major structures. From 2000 to 3500 feet (610 to 1067 m) B. G. L. there was limited evidence of continuous seepage. Where present this seepage was associated with faults and not shear zones. From 3500 to 6000 feet (1067 to 1828 m) B. G. L. there was no evidence of sustained groundwater seepage.

In addition to these generalizations, a number of geologic parameters also affect the groundwater characteristics of major structural features. It was observed that in the range of 0 to 2000 feet (610 m) B. G. L. only about 25 per cent of the faults and shear zones seeped continuously. The greatest quantity of seepage issued from faults with distinctive breaks characterized by a mud or clay seam. Shear zones that were marked by wider zones of multiple shearing without a mud or clay seam were usually "dry".

The mineralogy of fault gouge or any fracture infilling also may influence the ultimate permeability of a structural break. For example, the calcite filling a fault zone or joint opening may be dissolved in one area and precipitated in some other area depending on the groundwater chemistry. Hence over long periods of time, dissolution and precipitation of suspended ions could effectively open and close potential groundwater conduits at depth.

It also became apparent from observations of faults and shear zones at depth that although the main zone in some cases may act as a barrier to groundwater inflow, secondary fracturing associated with the main break was usually a source of sustained groundwater seepage. Seepage from this fracturing was found to have the same depth-inflow relationships as major structural features.

Seepage from major structures was observed on two occasions to be recharged directly from surface. These faults were observed at 500 to 700 feet (152 to 213 m) B. G. L.

In addition, it is obvious that the virgin rock mass is saturated at depth and that a lack of groundwater seepage in the deeper mines reflects a low hydraulic conductivity or permeability rather than an absence of water at depth. The presence of free water on advancing faces and the intersection of numerous water-filled features, which drained quickly, tend to substantiate this conclusion.

Both major and minor structures measured at or

Both major and minor structures measured at or near the surface were persistent to the deepest mine workings. Surface expression of persistent faults varied from 1 foot (0. 3 m) wide fracture zones to 100 to 200 foot (30 to 61 m) wide erosional gullies. The thickness of most of these features at depth was generally similar to thickness measured near or at the surface.

Again, as with the case histories, it is important to consider the structural setting of the mines studied in applying the tentative conclusions reached to other areas of the Canadian Shield. Ore deposits in many cases are associated with regions of considerable geologic structure. When combined with the disturbance and fracturing induced by mine blasting, it may be reasonable to assume that subsurface groundwater conditions may be more favourable in massive, undisturbed plutonic rocks.

Acknowledgments

The writer is grateful to the personnel of the nine mines visited, who provided information and allowed the author to inspect and obtain, on site, data on subsurface geologic structures and seepage zones.

The writer specifically wishes to thank Camflo Mines Ltd. ; East Malartic Mines Ltd. ; Sigma-Mines Ltd. ; Noranda Mines Ltd. , Chadbourne Mine; Falconbridge Copper Ltd. , Lac Duffault Division, Millenbach Mine; Kerr Addison Mines Ltd. ; Willroy Mines Ltd. , Macassa Division; Dome Mines Ltd. ; and Noranda Mines Ltd. , Pamour Schumacher Division. The assistance of the Gull Island Power Co. Ltd. , Lower Churchill Consultants Ltd. , and Acres Consulting Services Ltd. , in providing the author with the results of geotechnical investigations and summary geotechnical reports, is greatly appreciated.

The supervision and assistance provided by J. E. Gale and the active participation of J. Plavka are gratefully acknowledged.

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Project 680047

J. Ross Mackay[1]
Terrain Sciences Division

Large ice-wedge polygons are a familiar site in many polar areas. The polygons are bounded by troughs below which lie ice wedges with the apexes tapering downwards. Ice wedges are most commonly seen in fresh exposures along coastal bluffs or in artificial excavations, such as a trench or road-cut. Because ice wedges form a three-dimensional network, the width of an ice wedge seen in a vertical section will depend upon the orientation of the section with respect to the trend (strike) of the ice wedge. The purpose of this note is to comment upon the widths and spacing of ice wedges as seen in natural and artificial sections.

Ice-Wedge Width

Let us assume that an ice wedge is exposed in vertical section along a randomly oriented line making an angle α with the strike of the ice wedge (Fig. 10.1a). If the apparent width of the ice wedge is a, the true width (w) is then:

$$w = a \sin \alpha \qquad (1)$$

Since the angle of intersection (α) can have any value from 0 to $\frac{\pi}{2}$, the mean value of sin α is (cf. Wentworth, 1930):

$$\text{Mean of } \sin \alpha = \frac{2}{\pi} \int_0^{\frac{\pi}{2}} \sin \alpha d\alpha = 0.64 \qquad (2)$$

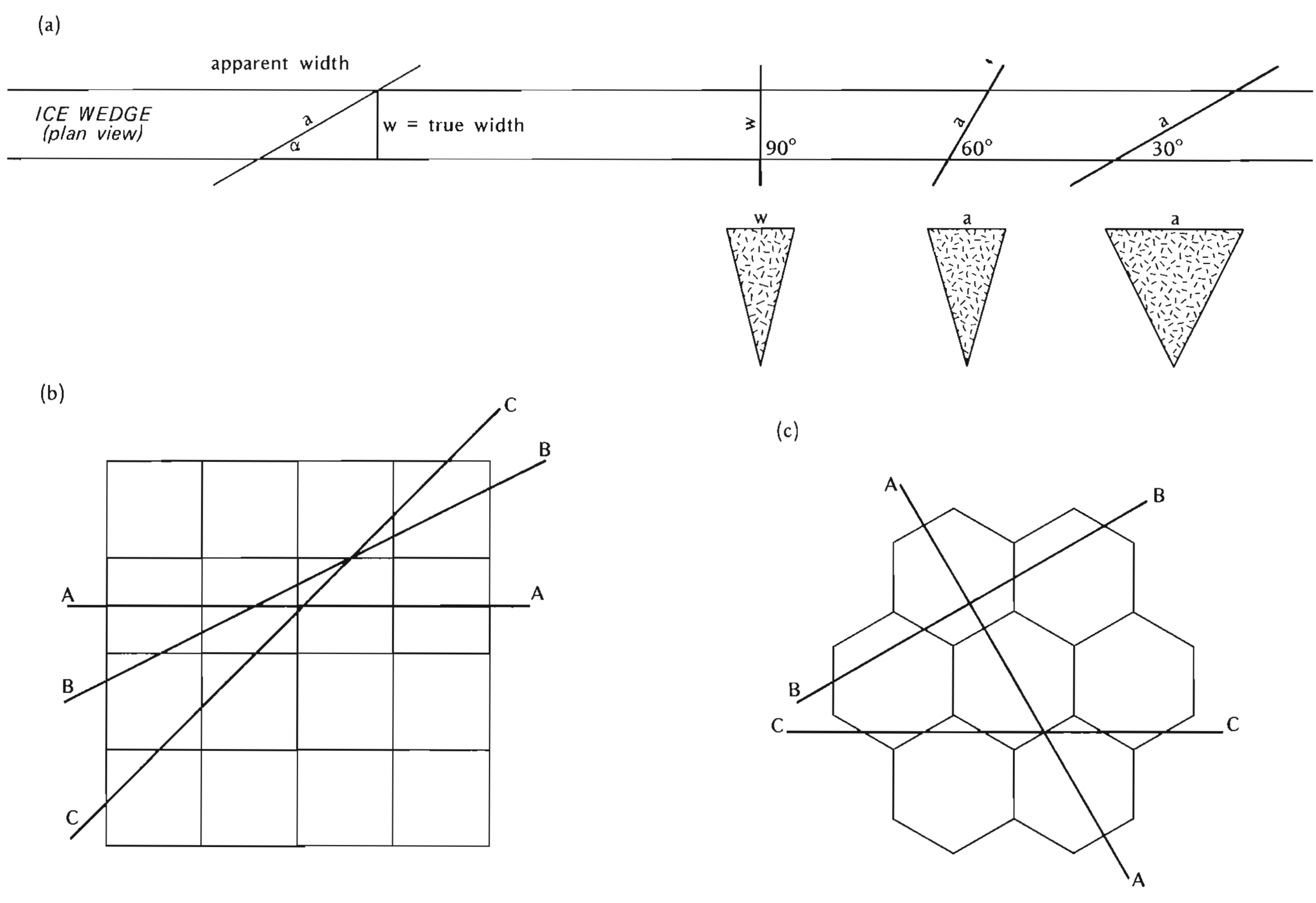

Figure 10.1. (a) Plan and vertical sections of ice wedges to illustrate true and apparent widths.
(b) Rectangular ice-wedge pattern with lines of intersection AA, BB, and CC.
(c) Hexagonal ice-wedge pattern with lines of intersection AA, BB, and CC.

[1] Department of Geography, University of British Columbia, Vancouver, B.C., V6T 1W5

From: Report of Activities, Part A;
Geol. Surv. Can., Paper 77-1A (1977).

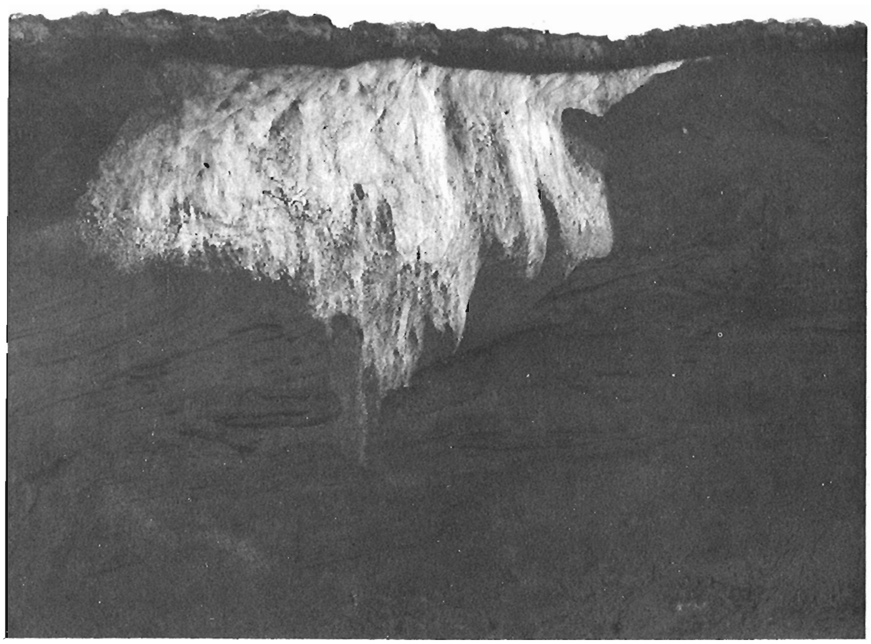

Figure 10.2.

Ice wedge exposed along a bluff at Hooper Island, Northwest Territories. The apparent width of the ice wedge, at the top, is about 6 m, whereas the true width is about 2.5 m. Note the absence of a trough above the ice wedge.

Therefore, if numerous ice wedges of identical size were exposed along a section, the true width would be approximately:

$$w \sim 0.64a \qquad (3)$$

In many exposures the trend of the ice wedge can be measured on the ground surface and the true width then calculated. However, commonly the strike or trend is unknown for a variety of reasons. For example, there may be no trough above an ice wedge on a hill slope; the trough may be snow covered; the ice wedge may be buried, inactive, and have no trough; and so forth (Fig. 10.2). Similar problems arise in the study of fossil ice-wedge casts. The true width of an ice wedge, of course, cannot exceed the apparent width; but if the apparent widths are taken to be the true widths, overestimation, on the average, will be of the ratio of 3/2 or 50 per cent. In general, a correction of 2/3 to the apparent width will give a good approximation to the true width. This also applies to the zone of disturbed structures which commonly parallels an ice wedge. The apparent height of an ice wedge is also the true height, except for the rare case where the line of section is along the trend of the ice wedge, but off centre. Anomalies occur at ice-wedge intersections or with asymmetric wedges.

Ice-wedge Spacings

A substantial proportion of ice-wedge polygons are rectangular or hexagonal in shape. Figures 10.1b and 10.1c show square and hexagonal polygons with lines of sections AA, BB, and CC. Lines AA (Fig. 10.1b and 10.1c) cut opposite square and polygon sides to give the true "widths" of the squares and polygons; lines BB cut opposite and adjacent sides of the squares in Figure 1b, but BB cuts polygon sides obliquely in Figure 1c; and lines CC cut adjacent sides in Figures 1b and 1c. Thus, in general, the spacings of ice wedges will be appreciably smaller than the "widths" of the squares and polygons.

Conclusion

The apparent width of an ice wedge, seen in a randomly oriented vertical section, is, on the average, about 50 per cent greater than the true width or, expressed another way, the true width is about 2/3 of the apparent width. The apparent heights are usually the true heights. The apparent ice-wedge spacing is generally much closer than that of opposite sides of the polygons. When ice wedges are measured in natural and artificial exposures, where the ice-wedge strikes are unknown, apparent widths will be greater than true widths and apparent spacings less than true spacings. Such factors should be considered in the estimation of the volume of ice-wedge ice in artificial excavations.

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11. TERRAIN RECONNAISSANCE, LAKE AGASSIZ MARGIN, NORTHERN MANITOBA

Project 750072

Lynda A. Dredge and Marie-France Dufour
Terrain Sciences Division

Introduction

This paper presents the results of a preliminary survey of types of landforms and materials encountered in northern Manitoba at the edge of the Agassiz lake basin (Fig. 11.1). Mapping was restricted to a strip 6 km wide along Highway 391, beginning at Lynn Lake and ending at longitude 100°W, about 30 km from Leaf Rapids. Roadlines east of this area have been mapped previously by Klassen and Netterville (1973) and Klassen and Veillette (1976) and are not included in this report.

The only exposures in the area were created during the construction of the road. The survey, therefore, was conducted by maintaining a log of materials exposed in ditches along the right of way together with more detailed observations made at borrow pits or extensive cuts.

Classification and Description of Materials

Preliminary Comments

The terrain in this area reflects the variety of materials encountered crossing the limit of a glacial lake basin. The main depositional environments, from west to east, are terrestrial subglacial, subglacial lacustrine (shelving), glaciolacustrine, and deep-water lacustrine. Evidence of littoral environments is not present. Superposed on this main facies sequence are minor amounts of stratified ice-contact and proglacial deposits.

Over much of the area, overburden is less than 5 m thick. As a result, many small unmapped bedrock outcrops are present, particularly in areas mapped as "till". Peat deposits are also widespread, and where these are small or thin (less than 1 m) they have been omitted. Therefore, areas described by a single designator, e.g. "till", are actually composites, consisting primarily of "till" but including minor areas of rock knobs and peat bog.

Materials described below are classed genetically. Because of facies changes, the variation in material within a given unit may be greater than the variation between units. Figure 11.2 has been constructed, therefore, in an attempt to clarify relationships between genetic units, facies changes, and soil texture.

Map Units

1. Bedrock (undifferentiated igneous and metamorphic rocks). Bedrock is a widespread surface feature over the area but only larger outcrops are shown in Figure 11.1. Most outcrops are low and undulatory with polished surfaces. Granitic and gneissic rocks are faceted or grooved; trap rocks have well developed sets of striae. Abrasion marks west of Hughes Lake bear south-southwest. Those east of Hughes Lake tend to bear southeast, although a few are oriented in southwesterly and westerly directions.

2. Till Three facies of till are present, reflecting differences in the environment of deposition.

Type 2a is prevalent in the western part of the area and on high ground in the central region where it flanks bedrock knobs. It has been also observed in exposures beneath lacustrine deposits farther east. This till consists of a sandy matrix with abundant subangular or angular pebbles, cobbles, and boulders of mixed lithology. Fines are present in significant amounts. Figure 11.3, sample 1 (*see* Fig. 11.1 for location of sample sites) shows the textural breakdown of the -4 mm fraction of material typical of unit 2a. The till is interpreted as being a product of terrestrial, subglacial deposition.

Till type 2b differs from unit 2a in that it includes occasional sandy lenses and discontinuous pebble layers, which indicate incipient water sorting. This till is found in the central map zone between Hughes River at Adam Lake and Hughes Lake and probably represents an ice shelf or ice ramp environment, with till from the base of the ice being deposited into standing water. The zone represented by this type of till has been interpreted as the margin of Lake Agassiz.

Till type 2c consists of angular granules and occasional boulders in a poorly sorted, silty matrix. The sand component is notably less than in type 2a and 2b tills (*see* Fig. 11.3, sample 4). This unit probably represents the dropping of debris into the lake basin from calving or rafted ice. Materials of this unit are widespread in the eastern part of the map-area. Minor amounts of type 2c are also present in the Lynn Lake area where small, short-lived glacial lakes existed.

3. Ice-contact stratified deposits Ice-contact stratified drift is restricted to an esker ridge at Esker Lake, 3 km from the eastern edge of the area studied. The ridge is sinuous, well defined, about 10 m in height, and 60 m broad at the base. It is flanked on the south by a narrow kame terrace. Material comprising the esker consists of large, well rounded cobbles and medium to coarse gravelly sand in disrupted cross-beds. This, together with the disposition of the beds, which do not conform to the surface expression of the esker, suggest that deposition occurred in a high energy environment under conditions of closed channel flow. Two gravel pits in the esker, spaced 2.5 km apart, indicate that the esker sediment rapidly grades to sand towards the easterly terminus, perhaps reflecting a rapid loss of competence where debris from the tunnel

From: Report of Activities, Part A;
Geol. Surv. Can., Paper 77-1A (1977)

SURFICIAL DEPOSITS

LYNN LAKE - LEAF RAPIDS, MANITOBA

LEGEND

6 PEAT

5 LACUSTRINE DEPOSITS
- 5a deltaic sand
- 5b silt and clay

4 PROGLACIAL DEPOSITS
outwash sand and gravel

3 ICE-CONTACT STRATIFIED DEPOSITS
esker sand and gravel

2 TILL
- 2a sandy till
- 2b sandy till with bedding
- 2c silty till

1 BEDROCK

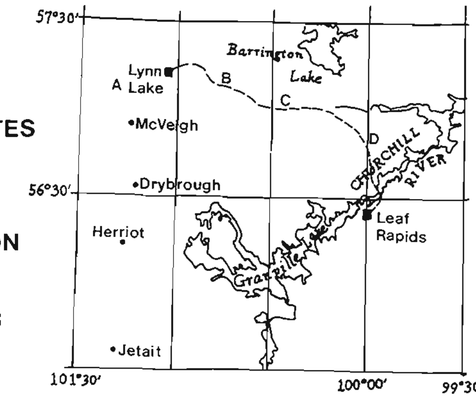
ROAD

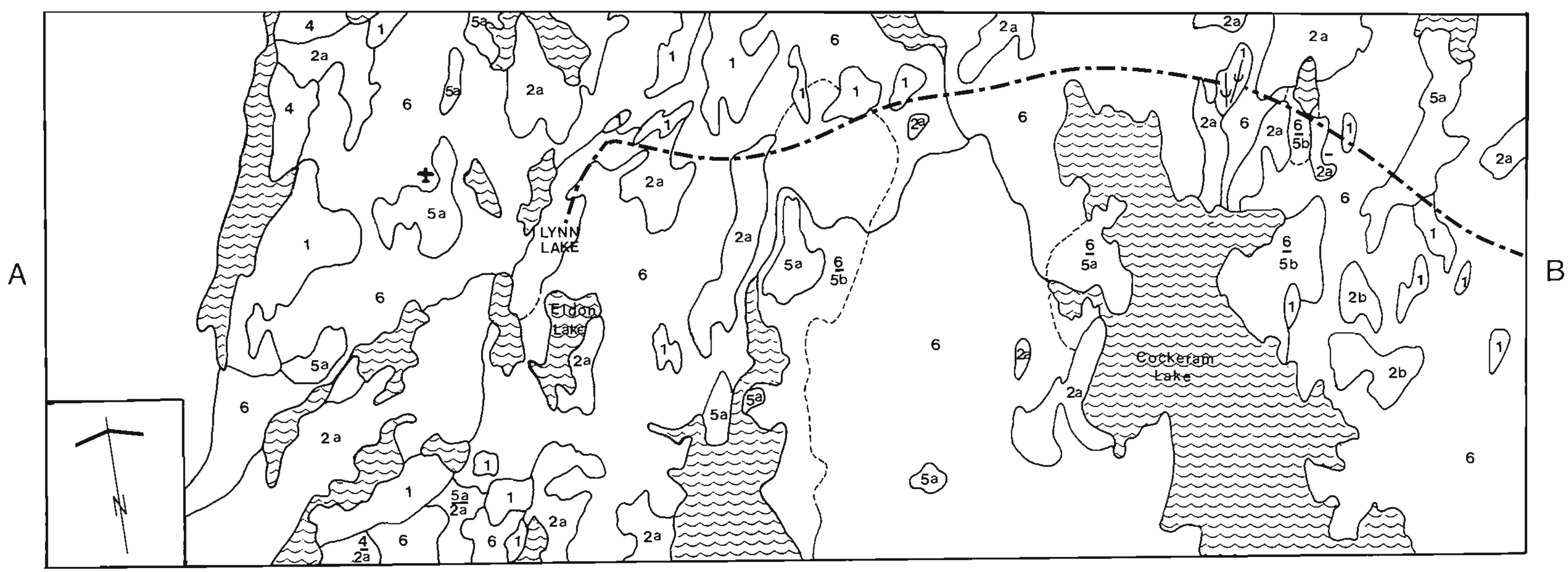
SAMPLE SITES

AIRPORT

STRIAE ORIENTATION

MAP-UNIT BOUNDARY; DEFINED, ASSUMED





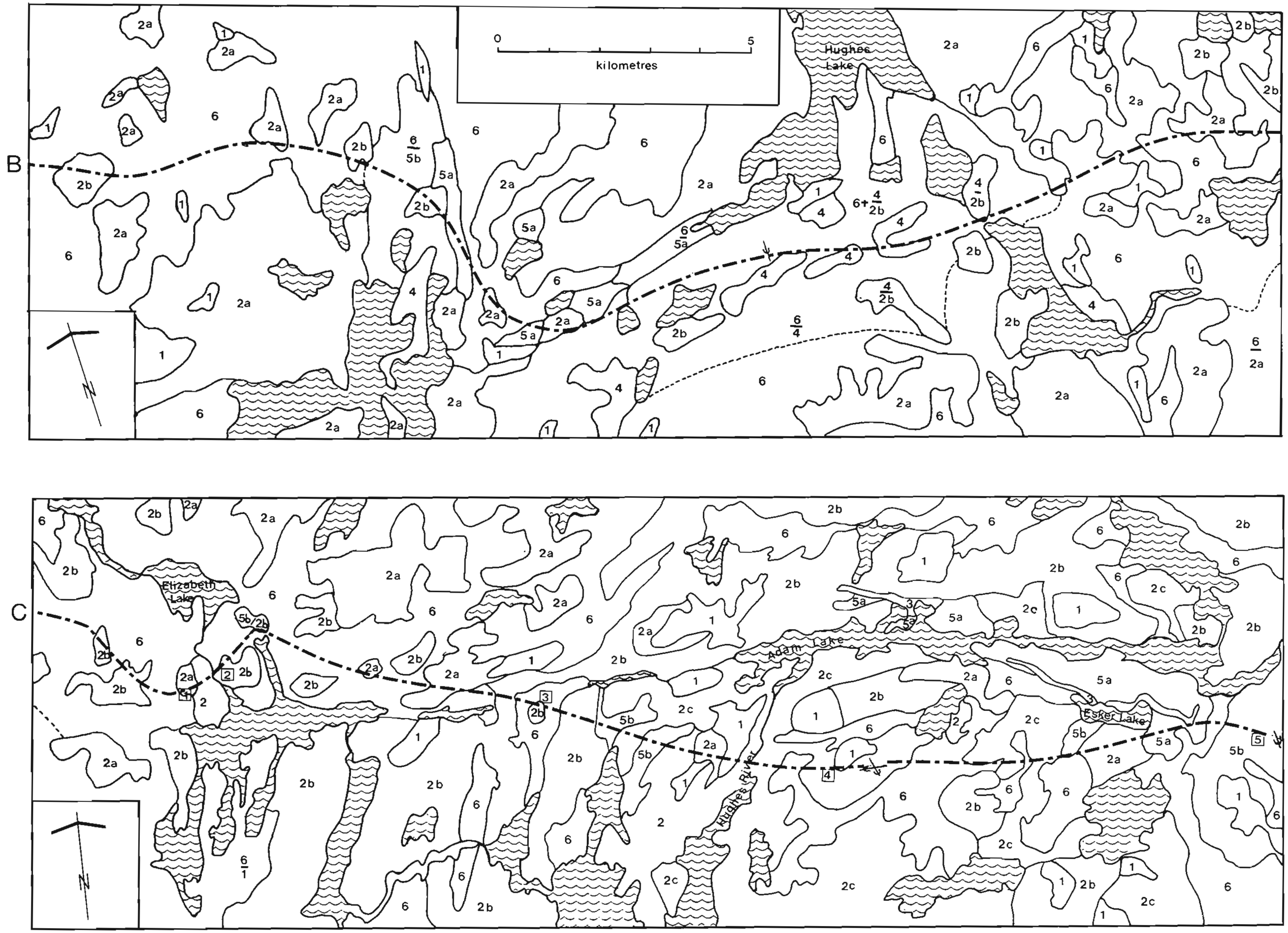

kilometres
0
5
Hughes Lake
Elizabeth Lake
Adam Lake
Esker Lake
Hughes River
B
C
D

Figure 11.1.

GENETIC UNIT	ENVIRONMENT OF DEPOSITION				
	TERRESTRIAL	LITTORAL	NEARSHORE		DEEPWATER
	TEXTURE				
	SAND+GRAVEL	→	SAND	SILT →	CLAY
ORGANIC (6)	—	—	—	—	—
LACUSTRINE (5)		—	5a	→	5b
FLUVIAL-PROGLACIAL (4)	4	→	5a		
FLUVIAL-ICE CONTACT (3)	3	→	5a	→	5b
GLACIAL (2)	2a	→	2b →	2c	—
BEDROCK (1)					

Figure 11.2.

Facies-texture realtionships between map-units.

entered the standing water in the lake basin. The lacustrine extension of the esker, which consists of fine deltaic sand, is included in unit 5a.

4. Proglacial deposits Thin outwash deposits have been found in two locations. They occupy lowland channels on either side of Hughes Lake and overlie till (unit 2a) to a depth of about 2.5 m on the higher ground directly south of the lake. The lithological and textural characteristics of the outwash are similar to type 2b till, but the deposits are more completely stratified and the clasts are subround to round. Glaciofluvial deposits were also identified in channels about 5 km northwest of Lynn Lake; these grade southwards into deltaic sands (unit 5a) near the Lynn Lake airport.

5. Lacustrine sand and clay Unit 5a is the lacustrine facies associated with the glaciofluvial units described above (Fig. 11.2). In pits south of Lynn Lake (directly south of area studied) outwash grades laterally into tabular deltaic foresets consisting of grey, medium grained sand with some fines. A few boulders are included but granules and pebbles are missing. Similar deposits are exposed at the Lynn Lake airport (without outwash). Although there is a slight coarsening upwards, topset beds are absent. The body of standing water into which these sediments were deposited reached to at least 365 m a.s.l. (1200 ft.), 30 m above present day lake levels. Its geographic extent has not been determined, but the extensive distribution of the deltaic material south of Lynn Lake suggests that a large lake may have existed south of the area.

Unit 5a also forms a fan-shaped deposit at the southern end of the esker near the eastern margin of the area studied. Sands here are finer and more micaceous than at Lynn Lake. The deposit is at least 5 m thick; it grades laterally to clay over a distance of about 0.5 km. About 1.5 m of varved clay unconformably overlies the sand. The distribution of the sand and clay suggests that sedimentation occurred subaqueously from the esker tunnel into a moderately deep body of standing water.

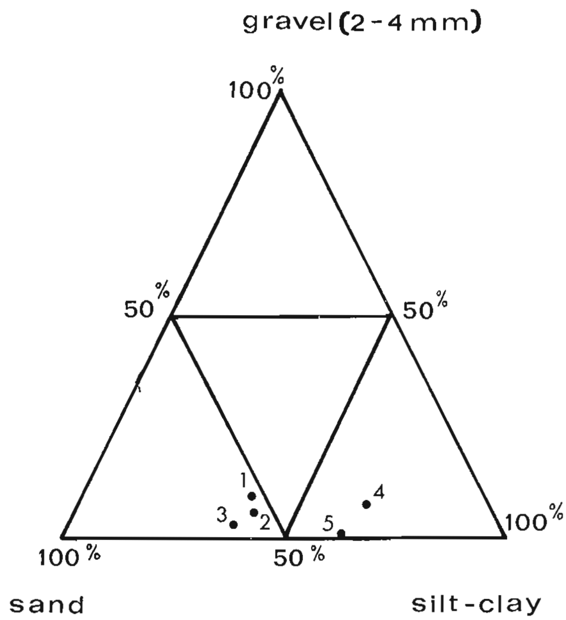
Unit 5b covers the easternmost part of the area and represents a deep-water lacustrine environment. Very local deposits also can be found farther west. The material consists of stiff, plastic, silty clay (clay content = 60%) containing angular ice-rafted boulders. In most exposures the clays are massive but above the deltaic-esker sands, where annually fluctuating currents may have been present, varve couplets 1.5 cm thick have developed. These are composed of thin, light coloured silty bands alternating with thicker dark bands of clay.

6. Peat Peat is extensive in this area and mantles most other deposits. The unit mapped as peat in this report is restricted to deposits more than 2 m deep. Peat is common in the lowlands near drainage outlets where it underlies fens. In these areas it consists of the remains of grassy vegetation, and the entire area is generally waterlogged. The most extensive areas of peat, however, are bogs, which are prevalent on better drained, flat to undulatory sites. Bog peat is woody or fibrous. The bog surface is generally dry but the zone directly above the frost table is wet because of the presence of melting ground ice. In the bogs cored for this report the peat was frozen below a depth of about 75 cm, and crystalline ice was present as thin ice lenses.

Summary of Glacial History

Ice Flow Directions and Ice Fronts

Striae and grooves indicate that ice flowed south, southwest, southeast, and west over the area. The distribution is believed to reflect divergent radial flow from a late-glacial lobe centred near Hughes Lake. The easterly part of this lobe at one time terminated in Lake Agassiz and eventually may have retreated to a quasi-stationary position directly south of Hughes Lake, because an outwash plain is present in that area on relatively high ground. The westerly part of the lobe probably extended to the Lynn Lake townsite.



FACIES

sample 1 – subglacial terrestrial environment

sample 2, 3 – sublacustrine, ice ramp or ice shelf

sample 4 – glaciolacustrine, calving or rafted ice

sample 5 – lacustrine, with minor ice rafting

Figure 11. 3. Textural variation in till associated with changes in depositional environment.

The presence of outwash west of Lynn Lake suggests that a separate, stationary ice front existed about 5 km from the townsite.

Glacial Lakes

The distribution of materials shown on the map suggests that the area is crossed by the northwestern boundary of glacial Lake Agassiz. The clays in the easternmost area are contiguous with more extensive massive and varved clays mapped by Klassen and Netterville (1973) and Klassen and Veillette (1976). Facies changes within till, together with the absence of raised shorelines and other shallow water deposits, suggest that the lake terminated against an ice front within the map-area. The edge of the lake can be taken as the zone between Hughes Lake and Hughes River at Adam Lake, where an ice shelf or ice ramp existed. The presence of type 2a till on rock knobs above an elevation of about 295 m (975 ft.) within this zone indicates that the ice sheet was partially grounded. West of Hughes Lake the ice was completely grounded, and east of this zone ice calved or was rafted into the Agassiz Basin.

The esker at Esker Lake coexisted with Lake Agassiz. The sinuous part represents conditions of closed channel flow beneath the glacier. This part of the esker terminates at the edge of the lake, where sediment debouched into standing water to form a sandy, totally subaqueous, delta. The lake apparently outlasted esker sedimentation by about 100 years since 1. 5 m of varved clay overlies the deltaic sands.

Another glacial lake, extending to at least 365 m (1200 ft.) a. s. l. existed in the Lynn Lake area. Lacustrine deltaic deposits are restricted to the area around the airport at the townsite, but they become much more extensive farther south; therefore, the area investigated may only be the northern extremity of a much larger lake or system of lakes. At Lynn Lake the glacial lake was bounded on the northwest by outwash (and possibly an ice front). The presence of type 2a till lying to the east and south suggests that the lake did not extend beyond the townsite into nearby Eldon Lake basin, which is at a lower elevation. The area was not studied in sufficient detail to determine the cause of blockage, but it is hypothesized that the glacial lake was blocked on the east by the Hughes Lake lobe.

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Project 740063

J.J. Clague
Terrain Sciences Division, Vancouver

Introduction

The Georgia Depression, located between the Vancouver Island Ranges and the Coast Mountains of southwest British Columbia, is a major physiographic element of the Canadian Cordillera (Holland, 1964). That portion of the depression below sea level, and connected to the Pacific Ocean by relatively narrow passages to both the north and south, is termed the Strait of Georgia.

Bathymetric elements in the Strait of Georgia include basins and troughs up to 430 m in depth and banks elongate in a northwest-southeast direction (Fig. 12.1). These features have originated, in large part, through Pleistocene glacial erosion but have been modified by Holocene deposition (Mathews and Murray, 1966; Mathews, 1968; Clague, 1975a, 1976). In addition, tectonic processes probably have contributed to the shaping of the Georgia Depression which has been the site of repeated subsidence and deposition since Late Cretaceous time (Mathews, 1972).

Previous marine geology studies in the Strait of Georgia have been confined primarily to the central

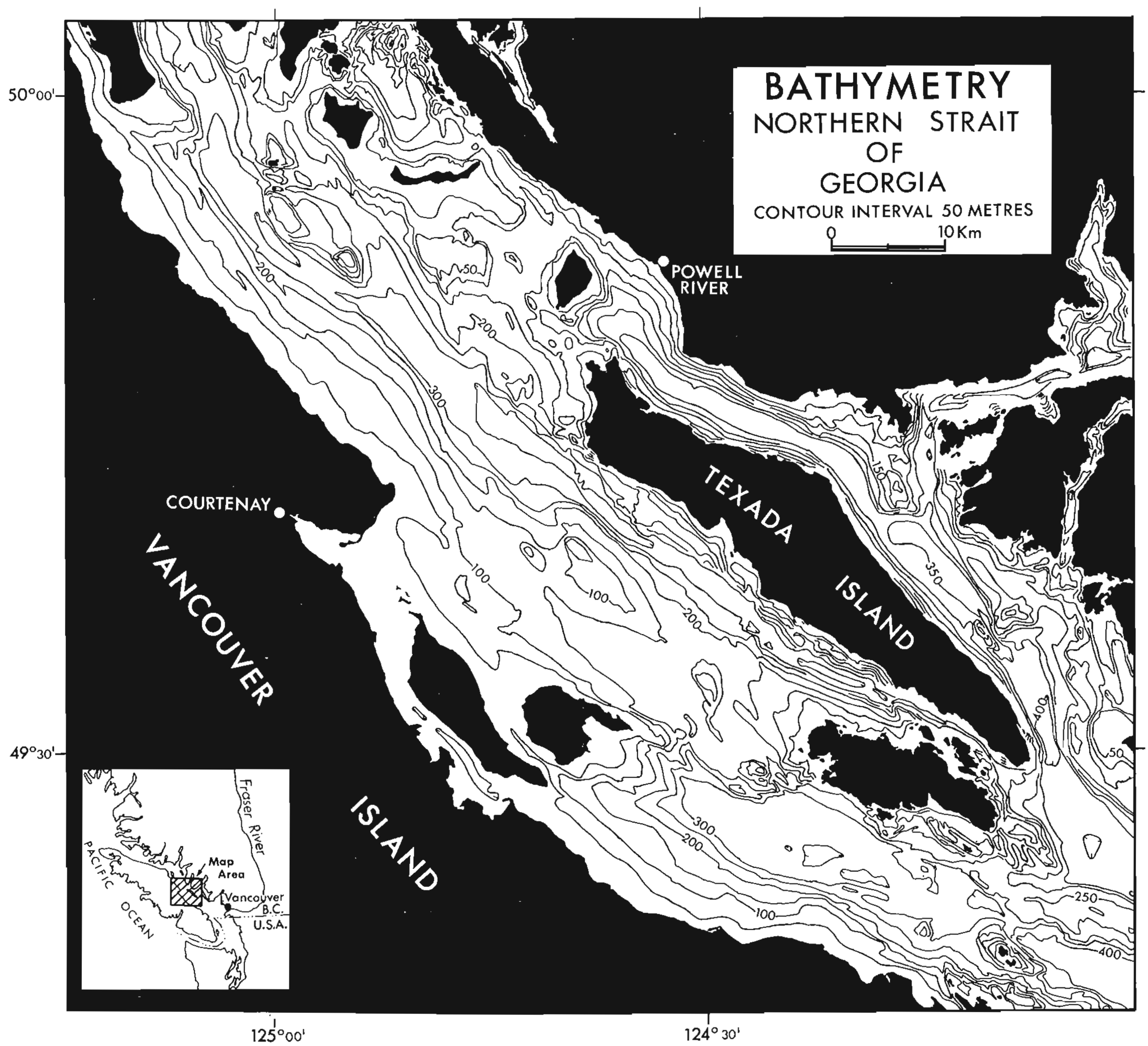

Figure 12.1. Bathymetric map of northern Strait of Georgia.

From: Report of Activities, Part A;
Geol. Surv. Can., Paper 77-1A (1977).

and southern sectors between latitudes 48°50'N and 49°30'N. Cockbain (1963) and Tiffin (1969) described the submarine topography and the subbottom structure, distribution, and thickness of Quaternary sediments in the southern and central strait on the basis of echo-sounder and air-gun seismic reflection profiles. Pharo (1972) and Pharo and Barnes (1976) documented the distribution and genesis of surficial marine sediments in the same area.

Although Waldichuck (1953) briefly described the bottom sediments of the northern Strait of Georgia and generalized their distribution on a small-scale map, there are no comprehensive published accounts of the Quaternary geology or sedimentology of the area. A field program in the northern Strait of Georgia thus was initiated by the Geological Survey of Canada in 1973 to provide additional information on the Quaternary geology of the area. Some results of this on-going project have been reported previously (Clague, 1975a, b, 1976); this report outlines the distribution of Holocene sediments in the northern Strait of Georgia as deduced from seismic surveys conducted during 1973 and 1974. This information complements that obtained in the central and southern Strait of Georgia (Tiffin, 1969). The granulometry of surficial sediments in the northern strait has been described in another paper (Clague, 1975b) and is only summarized in this report.

Methods

The areal distribution and thickness of unconsolidated sediments in the northern Strait of Georgia have been determined from continuous seismic reflection profiles. Fifty profiles, representing about 900 line kilometres, were obtained during 1973 and 1974 with air-gun sources of 16 cm^3 (1 cu. in.) and 82 cm^3 (5 cu. in.).

Holocene* sediments were identified in the profiles using the following criteria: (1) Seismic units considered to represent Holocene sediments are characterized by horizontal and near-horizontal reflectors which, in general, are parallel to the sea floor (Figs. 12.2 and 12.3). In contrast, Pleistocene reflectors commonly are truncated by the sea floor, a result of glacier scour (Fig. 12.2; Clague, 1976). (2) Units thought to be Holocene in age unconformably overlie units continuous with late Pleistocene sediments onshore. (3) Holocene seismic units occur mainly in troughs and basins, which presumably are the loci of present sedimentation, and are thin or absent on slopes adjacent to land areas and on interbasin banks and sills. In contrast, Pleistocene units are distributed with little relationship to present sea floor morphology (Clague, 1976). Corroborating evidence regarding the identity of various seismic units was obtained from grab samples and short cores of bottom sediments and from reconnaissance of the sea floor by submersible.

Some of the sediments identified by the above criteria as Holocene in fact may be glaciomarine sediments or stratified outwash deposited during the transition from glacial to nonglacial conditions at the end of the Pleistocene. However, because these sediments have not been overrun by glacier ice, that is, they are conformable with both overlying strata and the present sea floor, they have been grouped with postglacial sediments.

Distribution and Character of Holocene Sediments

Although Holocene sediments underlie most of the sea floor of the northern Strait of Georgia, in most places they are relatively thin (< 10 m)†. Substantial accumulations of postglacial sediments are limited to troughs and basins (Fig. 12.4), the thickest accumulations (≃ 175 m) occurring in basins at the south end of the study area nearest the large Holocene sediment bodies of the central Strait of Georgia. Although the trough and basin fill, in general, is thinner and more restricted at the northern end of the study area (Fig. 12.4), there is no systematic decrease in either sediment thickness or volume in a northerly direction. Any such trend is masked by the marked variability in sediment thickness within individual depressions. For example, Holocene sediments are thin or absent in some of the deepest parts of the northern Strait of Georgia where troughs are constricted.

Holocene basin materials are divisible in many places into two units on the basis of seismic character and morphology (Fig. 12.5). The lower unit, characterized by numerous, horizontal, internal seismic reflectors, is confined to depressions. This unit overlies bedrock and chaotically bedded or nonstratified Pleistocene sediments (drift?) and is overlain across a sharp, conformable contact by an upper unit of seismically transparent sediments. Unlike those of the lower unit, the upper sediments locally drape over slopes rising from basins and troughs.

Sediments beneath the sea floor adjacent to Vancouver Island probably veneer bedrock. Although the available seismic records in this area are of poor quality, at a few sites dipping reflectors occur immediately beneath the bottom, indicating that at least parts of this slope are underlain by bedrock with only a thin sediment cover.

Bottom sediments in the northern Strait of Georgia become finer with increasing depth and distance from source. The thick Holocene basin accumulations consist mainly of silty clay and clayey silt (Clague, 1975b). The margins of troughs are floored by silty sand and very poorly sorted mixtures of clay, silt, and sand. Holocene sediments on submarine ridges and slopes adjacent to land areas include silty sand, moderately to well sorted sand, and gravel.

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*Holocene is defined as an epoch of the Quaternary Period extending from the end of the Pleistocene to the present. In this report the time interval is considered to be equivalent to that of postglacial time. The boundary between glacial and postglacial time, however, is commonly defined on the basis of time-transgressive lithostratigraphic units and thus is itself diachronous.

†All thicknesses in this report are based upon an assumed seismic velocity of 1600 m/s (Tiffin, 1969).

Discussion

Holocene sediments in the northern Strait of Georgia are derived mainly from local sources including coastal bluffs and shallow submarine banks consisting of unconsolidated sediment (Waldichuck, 1953; Clague, 1975b). Material derived from these local sources is partitioned during transport, the coarsest sediment being deposited as a relatively thin cover in shallow water by tractive currents under a high energy regime and the finest sediment being concentrated in restricted depressions where current energy is low. Anomalies in sediment distribution are probably due to unknown complexities in current patterns.

Most fine detritus is transported in suspension to be deposited as part of the normal sediment rain in deep water; there is no evidence that turbidity currents control the distribution of modern sediment in the northern Strait of Georgia. Rather, uppermost Holocene strata dip over the lower slopes flanking depressions (Fig. 12.5, profiles HH' and II'), suggesting that deposition of fines occurs largely through the settling out of dispersed particles from suspension.

In the central Strait of Georgia sediments are deposited, at least in part, in the same manner; however, Tiffin (1969) suggested that turbidity currents also are important agents of sediment dispersal in this area. Most modern sediments in the central and southern Strait of Georgia are derived from a single source, the Fraser River (Pharo, 1972; Pharo and Barnes, 1976). Tiffin (1969) postulated that much of the detritus deposited off the mouth of this river subsequently is dispersed northward in turbidity currents and deposited over the large basins flooring the central strait.

Depressions in the northern Strait of Georgia, on the other hand, are separated by sills and constrictions from the main basins to the south; therefore turbidity currents originating off Fraser River do not affect the study area. Thus any sediment of Fraser River origin in the northern strait must be transported as dispersed clay particles and mixed in the water column with suspended detritus of local origin.

It is not presently known if detritus of Fraser River provenance is transported the minimum distance of 80 km into the study area. Aerial photograph analysis indicates that suspended sediment is transported with the freshwater surface jet from the mouth of Fraser River into the strait where sediment plumes, in general, veer to the north. This reflects the general northward movement of water through the strait (Waldichuck, 1957).

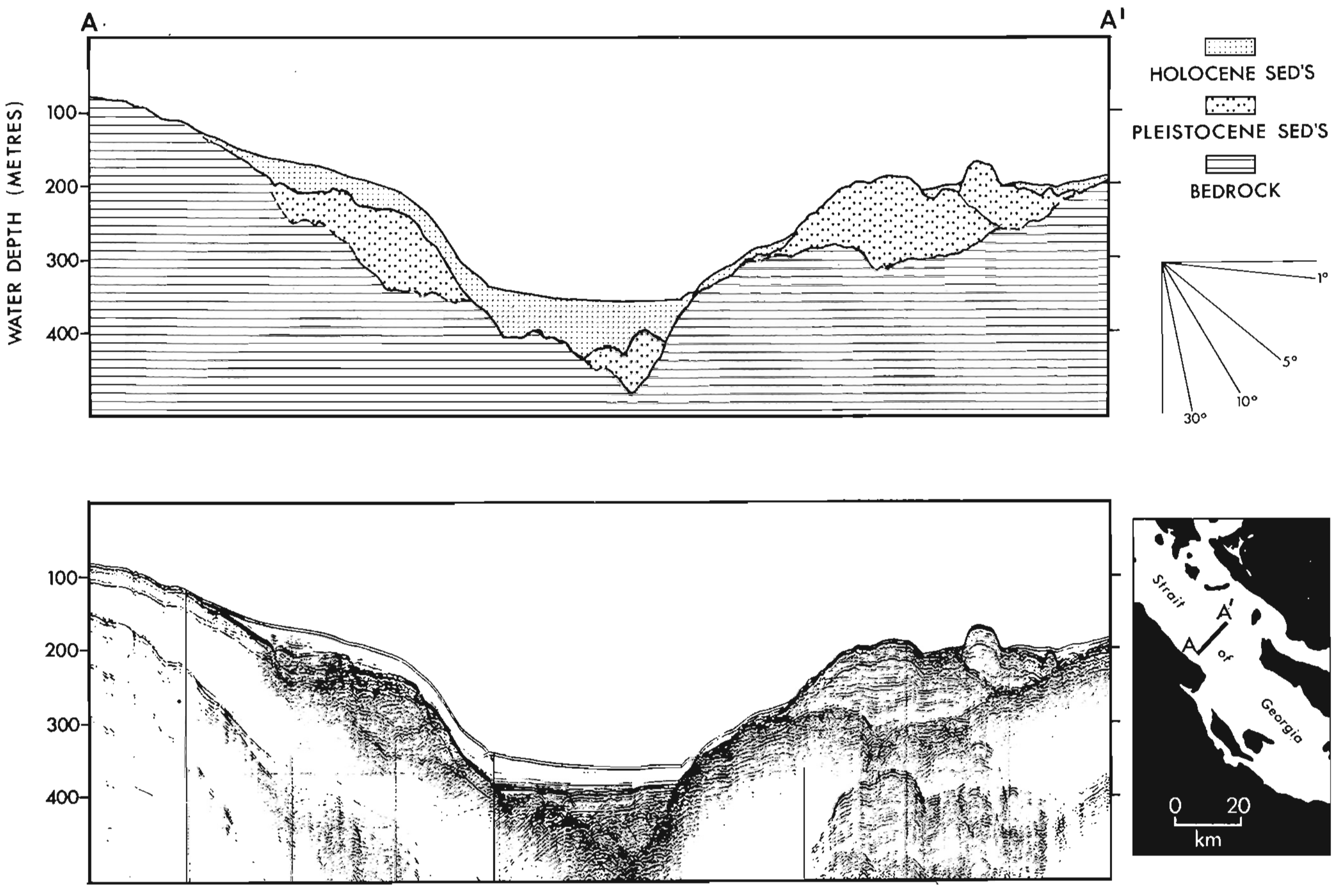

Figure 12.2. Selected seismic profile showing criteria by which Holocene sediments are identified. Horizontal and near-horizontal reflectors are parallel to the sea floor and, in places, unconformably overlie truncated Pleistocene reflectors. Holocene units occur mainly in troughs and basins and thus conform to the present sea floor morphology.

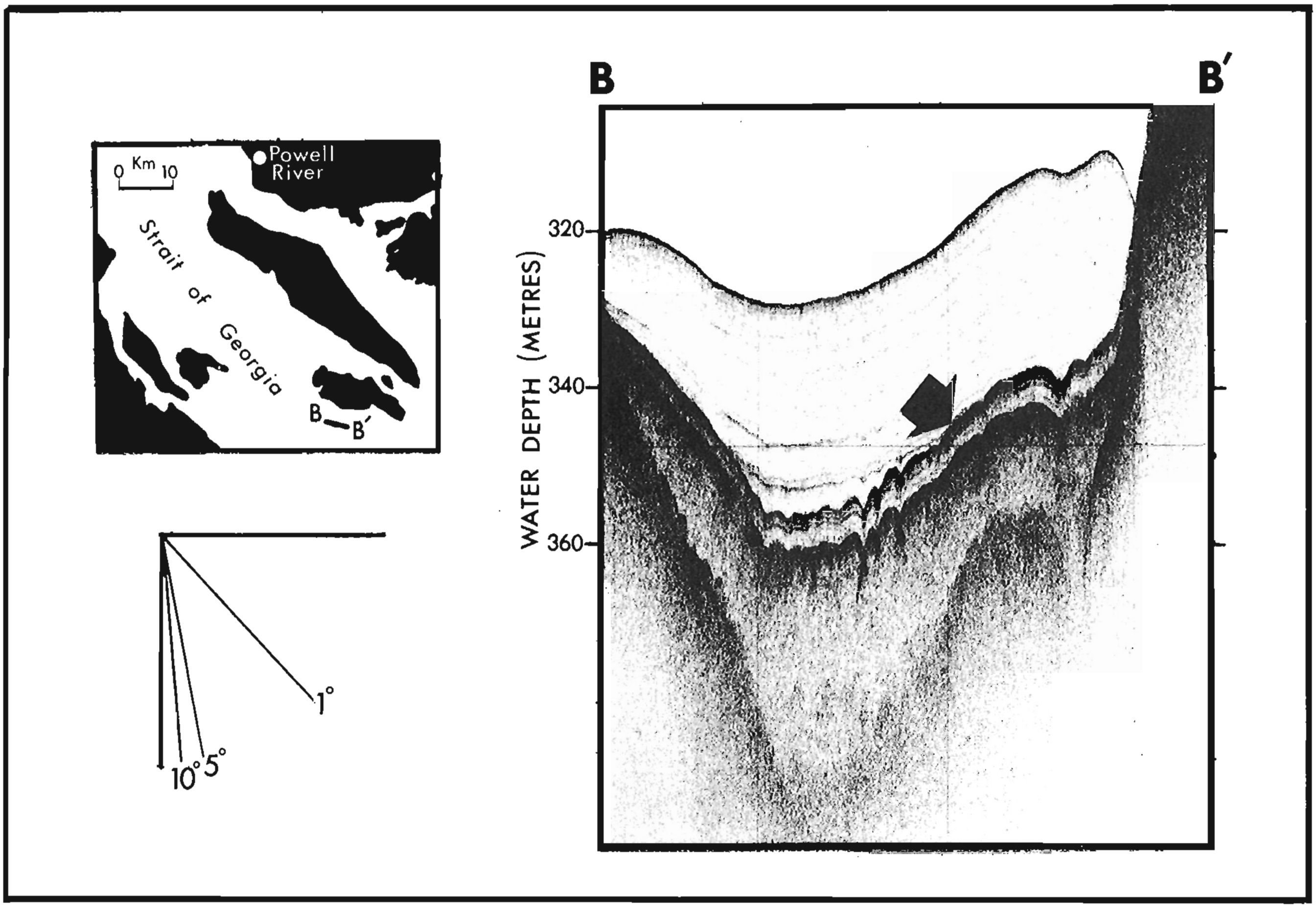

Figure 12. 3. Part of subbottom profile obtained with O. R. E. Model 140 profiling system operated at frequency of 3. 5 KHz. Upper Holocene strata are characterized by weak reflectors or are acoustically transparent. Sediments below arrow have strong seismic reflectors and are thought to be either late Pleistocene or early Holocene in age.

But suspended sediment surveys will be required to determine if detritus of Fraser River provenance is maintained in suspension by surface or deeper currents until reaching the northern Strait of Georgia.

Although present sedimentation in the northern Strait of Georgia likely occurs mainly through the rain of the fine grained silt and clay onto the sea floor and through the tractive transfer of coarser sediment from local sources into deeper water, early Holocene sedimentation patterns were perhaps different. Large amounts of terrestrial drift deposited in the Georgia Depression during the last glaciation became unstable during deglaciation in part because of the removal of supporting glacier ice and in part because relative base level fell (that is, early postglacial isostatic rebound exceeded the concurrent eustatic rise in sea level; Mathews *et al.*, 1970). As a result, much drift covering slopes within and adjacent to the Georgia Depression probably was eroded and mass wasted, and the denudation products were introduced into the Strait of Georgia. It is reasonable to assume that, because of this large influx of material, turbidity currents may have played a larger role in sediment dispersal during early postglacial time in the northern Strait of Georgia than at present. Presumably turbidity currents became less important sediment dispersal agents as the available supply of drift was progressively consumed in gravitational transfer into the strait.

Some support for this sedimentation model is provided by the seismic profiles. The presence of two Holocene units in many Strait of Georgia seismic profiles (Fig. 12. 5) suggests that the pattern of sedimentation has changed during postglacial time. Strong seismic reflectors occur within the lower unit, perhaps indicating that the sediments are strongly layered rather than massive. This stratification may have been produced by the periodic deposition of detritus from density currents. In contrast, the upper unit is almost seismically transparent and is thought to consist of silt and clay carried to the sea floor in a dispersed state. Contacts between the two units are sharp but conformable (Fig. 12. 5).

If the above sedimentation model is correct, it is likely that early Holocene rates of deposition were much greater than contemporary rates. Unfortunately quantitative estimates of these rates in the Strait of

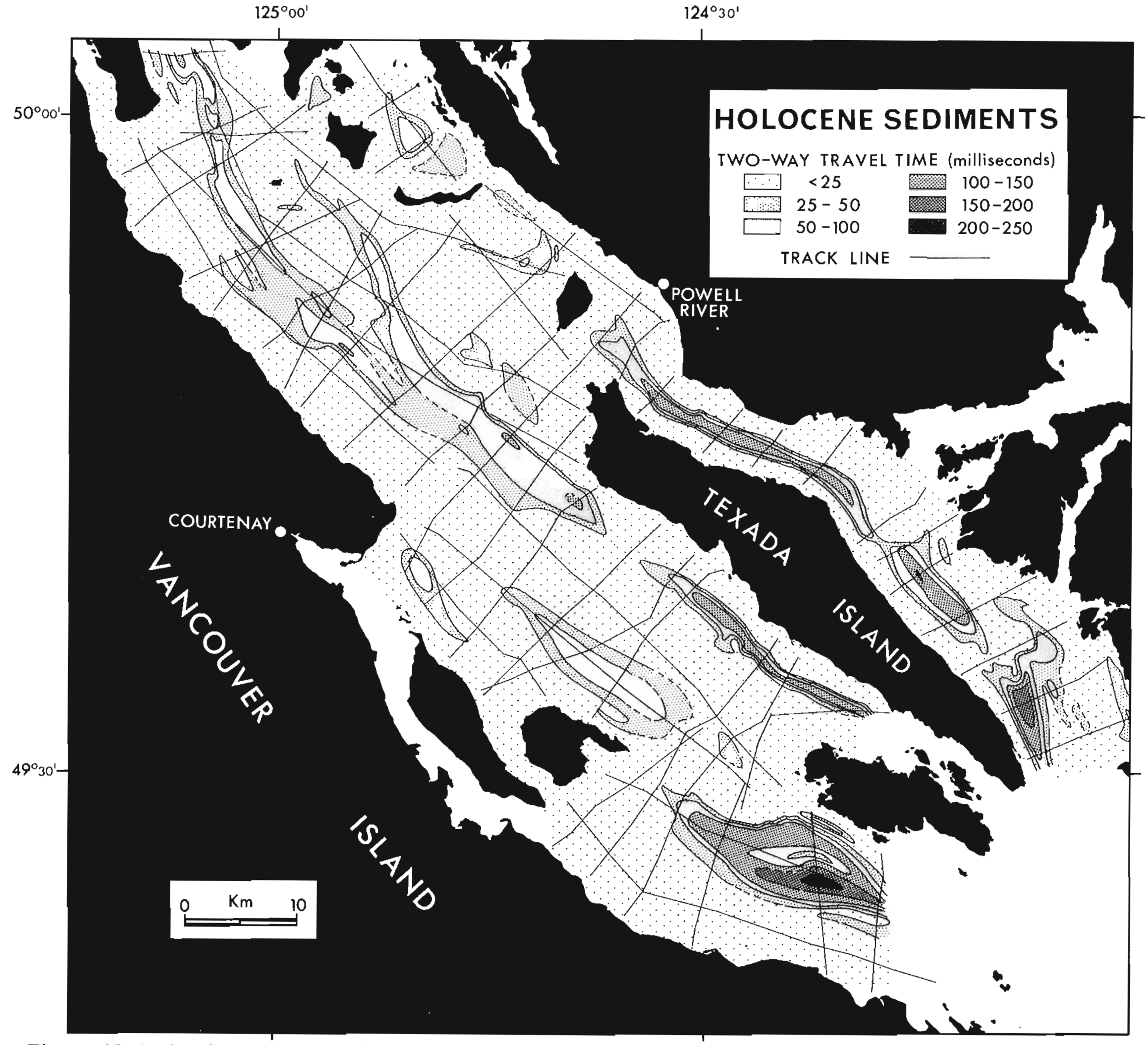

Figure 12.4. Isopleth map showing two-way travel time of sound through Holocene sediments in northern Strait of Georgia. If acoustic velocity through these sediments is assumed to be about 1600 m/s (Tiffin, 1969), isopleths of 25, 50, 100, 150, 200, and 250 ms correspond to thicknesses of 20, 40, 80, 120, 160, and 200 m respectively.

Georgia are limited by the dearth of pertinent radiometric age determinations. Most estimates, except those at the mouth of Fraser River, are averages for the entire postglacial interval and are based on maximum thicknesses of sediments within specific areas (Table 12.1). Present day sedimentation rates are probably smaller than those listed in Table 12.1, whereas early postglacial rates are undoubtedly much larger.

An average sedimentation rate for the entire northern Strait of Georgia for the past 12 500 years* has been determined by dividing the estimated total volume of Holocene sediments, based on the data presented in Figure 12.4 (43 km^3), by the product of 12 500 and the area over which the sediments occur (2670 km^2). The resulting rate, considered to be accurate to within ±20%, is 0.13 cm per year.

Because sedimentation has been concentrated in troughs and basins, rather than being uniform over the entire northern Strait of Georgia, a maximum sedimentation rate was calculated by dividing the maximum thickness of Holocene sediments (175 m) by 12 500. This rate, 1.4 cm per year, is much less than maximum sedimentation rates in Ballenas and Malaspina basins in the central Strait of Georgia nearer the mouth of Fraser River (Table 12.1)†.

*The oldest postglacial material adjacent to the northern Strait of Georgia, from a delta 150 m above present sea level at Courtenay, yielded a radiocarbon age of 12 500 y B.P. (Mathews *et al.*, 1970, p. 695). Sedimentation in the northern strait therefore is assumed to have begun at this time.

†Rates in the northern strait, however, are based on 12 500 years of sedimentation, whereas those from Ballenas and Malaspina basins are based on 10 000 years. If the latter are recalculated using a sedimentation interval of 12 500 years, they become 2.1 and 1.8 cm per year, respectively, still greater than the northern Strait of Georgia rate.

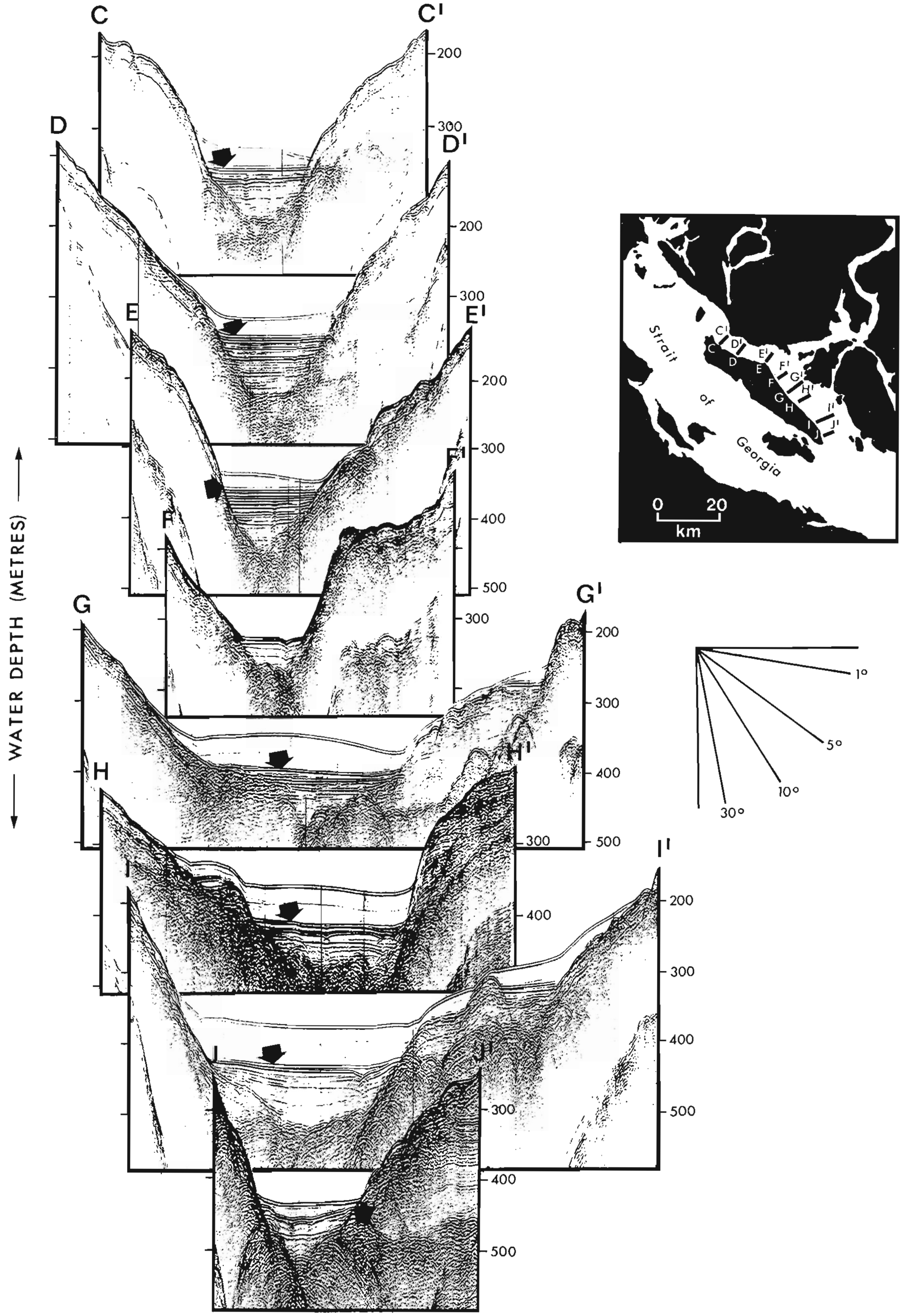

Figure 12.5. Seismic profiles across Malaspina Strait illustrating the Holocene fill underlying the floor of the trough. Postglacial sediments can be divided into two units, an upper, seismically transparent unit and a lower unit characterized by strong reflectors. Contacts between the two, indicated by arrows on the profiles, are sharp but conformable.

Table 12.1

Sedimentation rates in various parts of the Strait of Georgia

Area	Sedimentation rate (cm/y)	Significance of rate	Source
Fraser Delta front at ~ 37 m depth	34*	Average for active delta front	Mathews and Shepard (1962)
Fraser Delta front at ~ 90 m depth	37**	Average for active delta front	Mathews and Shepard (1962)
Strait of Georgia off Fraser Delta	2.7†	Maximum	Mathews and Murray (1966)
Northern central Strait of Georgia	0.6†	Average in basins	Mathews and Murray (1966)
Ballenas Basin, central Strait of Georgia	2.0†	Average	Tiffin (1969)
Ballenas Basin, central Strait of Georgia	2.6†	Maximum	Tiffin (1969)
Malaspina Basin, central Strait of Georgia	2.3†	Maximum	Tiffin (1969)
Northern Strait of Georgia	0.1††	Average for entire area	Clague (this report)
Northern Strait of Georgia	1.4††	Maximum in basins	Clague (this report)

Data in part summarized by Pharo and Barnes (1976, their Table 2).

*This rate of vertical aggradation corresponds to a rate of lateral aggradation of 4.6 m/y.

**This rate of vertical aggradation corresponds to a rate of lateral aggradation of 8.5 m/y.

†Rates are averages for postglacial time which is assumed to be 10 000 years in duration.

††Rates are averages for postglacial time which is assumed to be 12 500 years in duration.

Although no total volumetric estimate of Holocene materials in the southern and central Strait of Georgia has been made, Tiffin (1969) determined the approximate volume of sediments in Ballenas Basin, the largest basin in the central Strait of Georgia. His estimate of 72 km^3 includes sediments below a depth of 183 m west of the mouth of Fraser River and sediments underlying the entire length of Ballenas Basin, an area of approximately 250 km^2 (Tiffin, 1969, p. 69-70). This volume, determined for only part of the southern and central Strait of Georgia (total area ~ 3700 km^2), is much larger than the total volume of Holocene sediments in the northern strait (approximatley 43 km^3 over an area of about 2700 km^2).

This disparity in sediment volumes reflects the extent to which modern sedimentation in the Strait of Georgia is dominated by Fraser River. The rugged sea floor morphology and restricted areas of thick Holocene sediments in the northern sector attest to the relative isolation of the area from the influence of Fraser River. In contrast, the more subdued bottom morphology and thick sediment accumulations in the central and southern sectors are direct results of Fraser River sedimentation.

Acknowledgments

D.L. Tiffin supplied the technical expertise and equipment with which the seismic reflection data were collected. Also gratefully acknowledged is the ship-board assistance provided by I. Frydecky, J.L. Luternauer, K.E. Ricker, J.W. Scott, D.L. Tiffin, and crew members of the *C.S.S. Parizeau* and *C.S.S. Vector.*

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Project 700049

John E. Harrison
Terrain Sciences Division

"They [the sands]* belong chiefly to the time of emergence of the land and were deposited in the shallow waters near shore as the sea gradually receded, being derived very largely by wave and current erosion acting along the shore and on the shallow bottom". (Johnston, 1917, p. 27.)

The purpose of this note is not to report a new discovery or concept but rather to rediscover an idea which the author believes has not received the attention it deserves.

Following deglaciation, the Ottawa area, with the exception of the tops of the higher hills to the north, was inundated by the Champlain Sea. This large body of water was destined to be but a transient feature, for the land was beginning to rise even as the sea gained access to the region. During the time that elapsed between the first incursion of salt water and its final disappearance, the emerging land was subjected to all those active processes we generally study under coastal geomorphology.

Since the mid 1940's enormous strides have been made in our understanding of the processes active along modern coastlines. Electronic measuring devices, scuba diving gear, and advances in coastal engineering have contributed to an increased understanding of coastal processes. The application of this new knowledge to the many geomorphic features in the Ottawa area may contribute significantly to the understanding of the glacial and postglacial history of the region, as well as providing some aspiring graduate student with an interesting thesis topic.

In their investigation of geomorphic features in the Ottawa area researchers have placed primary emphasis upon glacial and fluvial processes. This paper presents the hypothesis that many geomorphic features in the area were formed by coastal rather than glacial or fluvial processes. Features which the author believes owe their form, structure, and fabric largely to coastal processes include most of the region's sand and gravel deposits, the upper portion of some clay deposits, and the occurrence of a till-like material reputed to be evidence of a glacial readvance.

Much of the sand and gravel in the Ottawa area occurs in elongate ridges variously considered to be an ice-contact sand and gravel moraine, interlobate moraine, or esker. In addition, where there is reasonable agreement that the feature is an ice-contact sand and gravel moraine (Bowesville moraine), disagreement centres on whether the ice that built it stood to the northeast or southwest. Studies attempting to resolve these conflicts rely heavily on measurement of cross-bed dips to establish the direction of flow during deposition. In none of these studies has the author been convinced that the researchers have taken enough care to distinguish between unmodified ice-contact sand and gravel and the spits, bars, beach ridges, and flanking deposits created by wave action. The extent of modification of original glacial features is best illustrated by the complex of landforms which lies at and to the south of Ottawa International Airport (Fig. 13.1).

As the ice retreated from the Ottawa area it left a ridge of material from the present location of the airport south for approximately 6 miles (it may have extended farther north or south but has since been eroded). As the marine waters shallowed due to uplift, this ridge (the Bowesville moraine) became an island in an extensive, shallow sea.

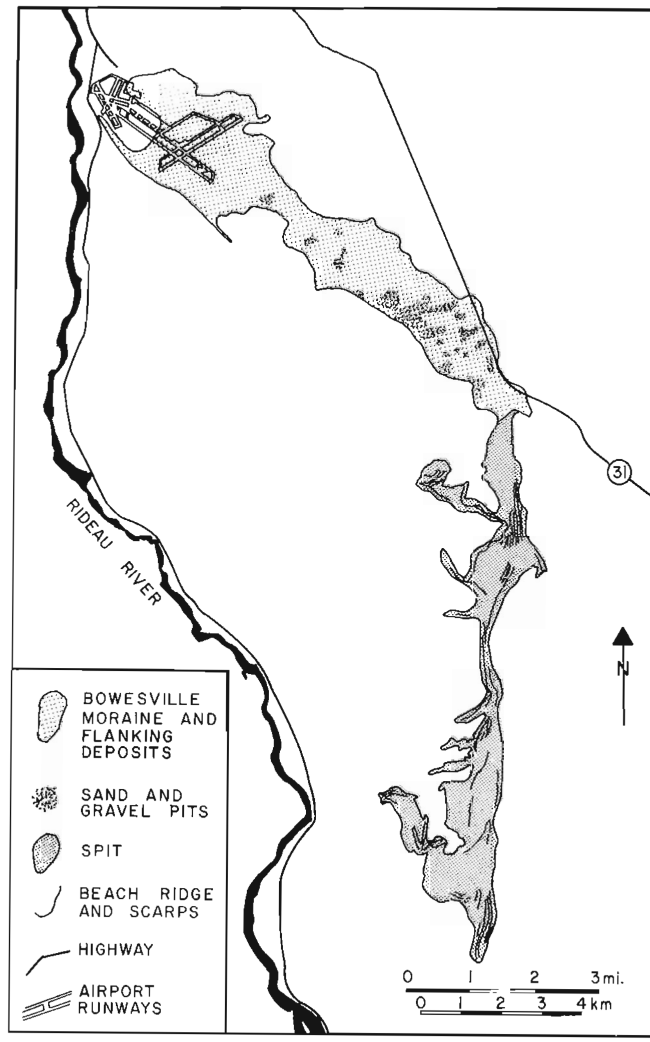

Figure 13.1. Map showing the Bowesville moraine, flanking deposits, and the recurved spit extending from the south end of the moraine (adapted from Richard, 1976a, b).

*Author's insert.

From: Report of Activities, Part A;
Geol. Surv. Can., Paper 77-1A (1977)

Waves eroding this island spread fine sand far out into the surrounding sea, and some sand and gravel was deposited down the flanks of the island as the material above sea level was eroded. Eventually the moraine was planed almost flat. Much of the eroded material was redeposited along the flanks of the original ridge or as a recurved spit extending from the south end of the deposit. The waves which generated the currents to form this spit probably originated to the northeast. The moraine at this time fronted on some 100 miles of open water. As the spit was built to the south, waves refracting off its southern end built segments which trended west. Renewed spit building left these segments sticking out of the growing spit as barbs (recurved spit). The continuous supply of sediment necessary for the growth of the spit came from the moraine which was being cut away as it was exposed by the slow drop in relative sea level. As the spit grew (and sea level fell), portions of it, too, were subjected to erosion, adding to the material supporting the southward growth of this feature. Today this spit is marked by numerous bogs on the east, the side that fronted the oncoming waves. In the lee of the spit, the west side, sand accumulated in flat plains or lobes.

Figure 13.1 shows the present extent of the modified moraine, flanking deposits, and spit. The enormous amount of material eroded, transported, and deposited in the building of the recurved spit associated with the Bowesville moraine provides ample evidence of the tremendous modification possible in the coastal environment. It is this demonstration of the ability of waves to rework large amounts of material to surprising depths which creates suspicion of the interpretation of the dip of sand and gravel beds exposed in gravel pits. Neither the presence of large boulders nor the absence of fossils guarantee that the deposit has not been reworked in the coastal environment, and the presence of cross-beds may reflect the slope of a beach or deposition at the end of a spit rather than glacially related depositional processes.

Large sand-covered areas lie both in the lee and on the open water side of wave-modified sand and gravel bodies. In the lee of these ridges and spits the sand probably represents washover lobes. The presence of seaweed (^{14}C dated at 10 800 years B.P., GSC-570), buried under sand in the Twin Elms area southwest of Ottawa, suggests that the areas behind bars and spits may represent good areas to look for material which could date the falling water level of the Champlain Sea. On the open water side of sand and gravel bodies thin sand commonly overlies thick clay deposits which lack the uppermost (reworked?) unit. The sands in the area northeast of the Bowesville moraine are marked by elongate, low, discontinuous ridges. These sands might have been deposited as ridges and runnels offshore of a sand beach. As sea level dropped these features migrated downslope in the direction of incoming waves (northeast), leaving behind a blanket of sand in the form of abandoned beach ridges and nearshore bars.

Some researchers have divided the marine clay in the area into an upper and lower unit based upon colour, grain size, strength, and structure. Perhaps the upper clay represents a reworking of the original marine clay by waves feeling the bottom as they crossed areas of broad, shallow, foreshore. If, in addition, the area was subjected to tides, surges, or seiches, reworking of the clay on "tidal flats" must certainly be considered as a possible origin of this upper clay. A detailed examination of air photographs for the shallow dendritic pattern associated with tidal flats, and a mathematical modelling of lower Ottawa Valley under shallow marine conditions, may help to prove whether tides could have played a part in deposition of this unit of the marine clay.

Finally, the theory that till-like material lying over fossiliferous strata may represent a glacial readvance into the Champlain Sea should be examined with respect to what is known about the effects of sea ice on the shore. There is reason to believe that sea ice formed in the Champlain Sea, and large ice floes driven onshore by wind can certainly rearrange the stratigraphy nearshore.

It is hoped that this note stimulates interest in the many fascinating coastal features in the Ottawa area. Where else is it possible to find 100-foot-deep gravel pits cut into beaches, bars, and spits. The opportunity to study the structure of these features (without getting wet) should not be missed.

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SOILS OF NORTH-CENTRAL KEEWATIN

C. Tarnocai[1]
Terrain Sciences Division

Introduction

Soils of north-central Keewatin, including NTS map-areas 56 E, K, L, M, N and the northwest quarter of 56 F, an area of approximately 30 000 km^2, were studied during the 1976 field season. This investigation was one component of a terrain study which also included surficial geology, geomorphology, and vegetation studies.

During the survey, data were obtained from 175 sites along helicopter and ground traverses. Information obtained from these sites included a general description of the terrain, a description of the soil based on soil pits dug to the frost table, and a description of vegetation around the soil pits. Usually either a soil parent material sample was collected or all horizons of the soil profile were sampled. In addition, detailed studies also were carried out at five sites (approximately one day per site). At all of these sites detailed soil studies were conducted using an electric hammer to dig into the perennially frozen portion of the soil to a depth of 1 to 1.5 m. The field activities resulted in the collection of 220 soil samples and 91 soil moisture and ice content samples.

Soil temperature and moisture studies also were carried out on ten sites which varied in soil parent material, drainage, genetic soil type, and exposure. Readings were taken frequently during the field season using a portable digital thermocouple-type instrument ("Digimite").

This paper presents a summary of findings, along with some of the data.

Ecoregion

The ecoregion represents similarities of climate as determined by vegetation, soils, to some degree, the permafrost condition, which then produce specific ecosystems on material having similar properties.

The entire study area is part of the "Low Arctic" ecoregion (Tarnocai and Boydell, 1975; Tarnocai *et al.*, 1976; Tarnocai and Netterville, in press). This ecoregion is characterized by continuous vegetation cover (except on bedrock and eroding surfaces). Soils are Cryosols with moderately well developed B horizons.

Soils

Soils in the study area are characterized by a shallow active layer, various amounts of ice in the perennially frozen horizons, and a cold soil climate. Cryogenic processes, especially cryoturbation, play a major role in the formation of these soils. Soils with these characteristics are classified as Cryosols.

Soils developed on till and fine textured marine silt materials are associated with strong cryoturbation (Turbic Cryosols) and patterned ground (e.g. circles and earth hummocks). Most of these Turbic Cryosols developed on till have a vesicular pore structure. The pores have a diameter of 3 mm or less and are especially common in the upper 20 cm of the soil. The development of pores is associated with microscale ice lens formation in the soil. The melting of these ice lenses in the spring creates an oversaturated condition (Washburn, 1969), and the slightest disturbance (e.g. vibration) will cause liquification of the soil material. This phenomenon is known as rheotropy (Yong and Warkentin, 1966).

Soils developed on ice-contact, or marine, sand and gravel and lacustrine sand are not affected by cryoturbation (Static Cryosols), but they are associated with ice-wedge polygons and frost cracks. Some of the coarse textured Brunisolic Static Cryosols, especially in the southern part of the study area, show weak leaching and, consequently, weak Ae horizon development.

Poorly drained soils (Gleysolic Turbic Cryosols) are water saturated throughout the summer and are associated with a shallow surface peat cover. Because of this peat cover the permafrost table is close to the surface (30 to 40 cm). Most of these soils, especially those developed on marine silt materials, show evidence of cryoturbation, and their perennially frozen horizons are associated with high ice content or pure ice layers.

Soil Temperature and Soil Moisture

Soil temperature measurements were taken at depths of 2.5, 5, 10, 20, and 50 cm and, at some sites, at the 100-cm depth between July 9 and August 15. The surface (2.5 cm) was the warmest with the temperature rising as high as 22°C (July 16) when the air temperature was 20°C. Sites with a southern exposure were the warmest throughout the summer (as was also found by French, 1970), especially site K7 on a mid-slope position (*see* Fig. 14.1 for site locations). Site K7, also with a southern exposure, warmed up considerably in spite of the fact that it is a poorly drained soil receiving a great deal of seepage water from a melting snowbank above. The coolest sites were those Gleysolic Turbic Cryosols which were associated with some surface peat cover (sites K2 and K18) and site K19 which is associated with a thick growth of lichens, mainly *Alectoria* spp.

Soil temperatures increased gradually during the month of July, but after July 28 the trend was reversed. During the stormy, snowy, and cool first week of August, the frost table rose markedly at several sites, especially those associated with a surface peat cover (Gleysolic Turbic Cryosol, Peaty Phase) or with a continuous thick growth of vegetation as on site K19 (*see* Figs. 14.1 and 14.2). On site K2 the frost table

[1]Canada Soil Survey, University of Manitoba, Winnipeg, Manitoba.

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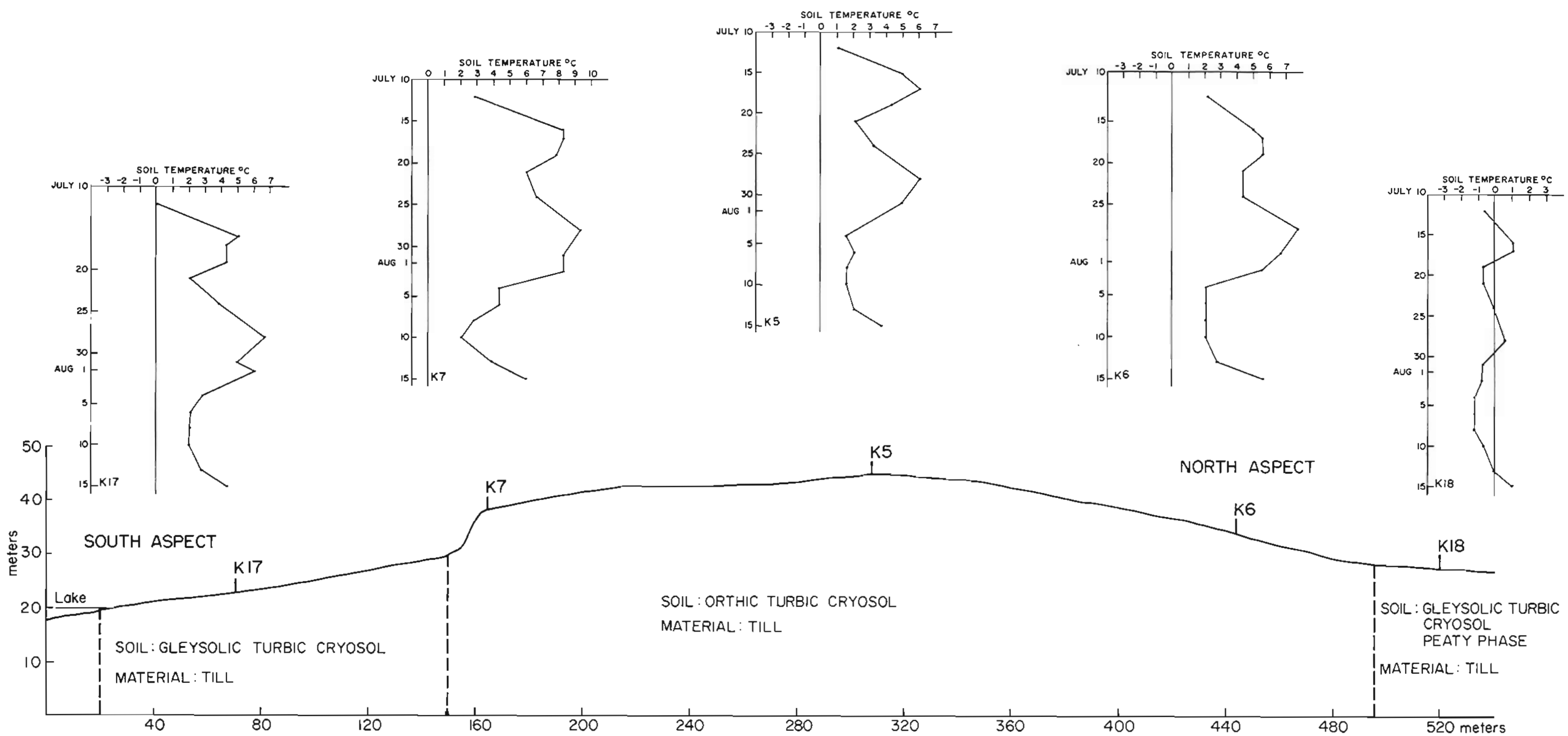

Figure 14. 1. Cross-section showing the location of soil temperature sites (66°06'N, 94°21'W) on southern and northern exposures of glacial till. All soil temperatures are given for a depth of 50 cm.

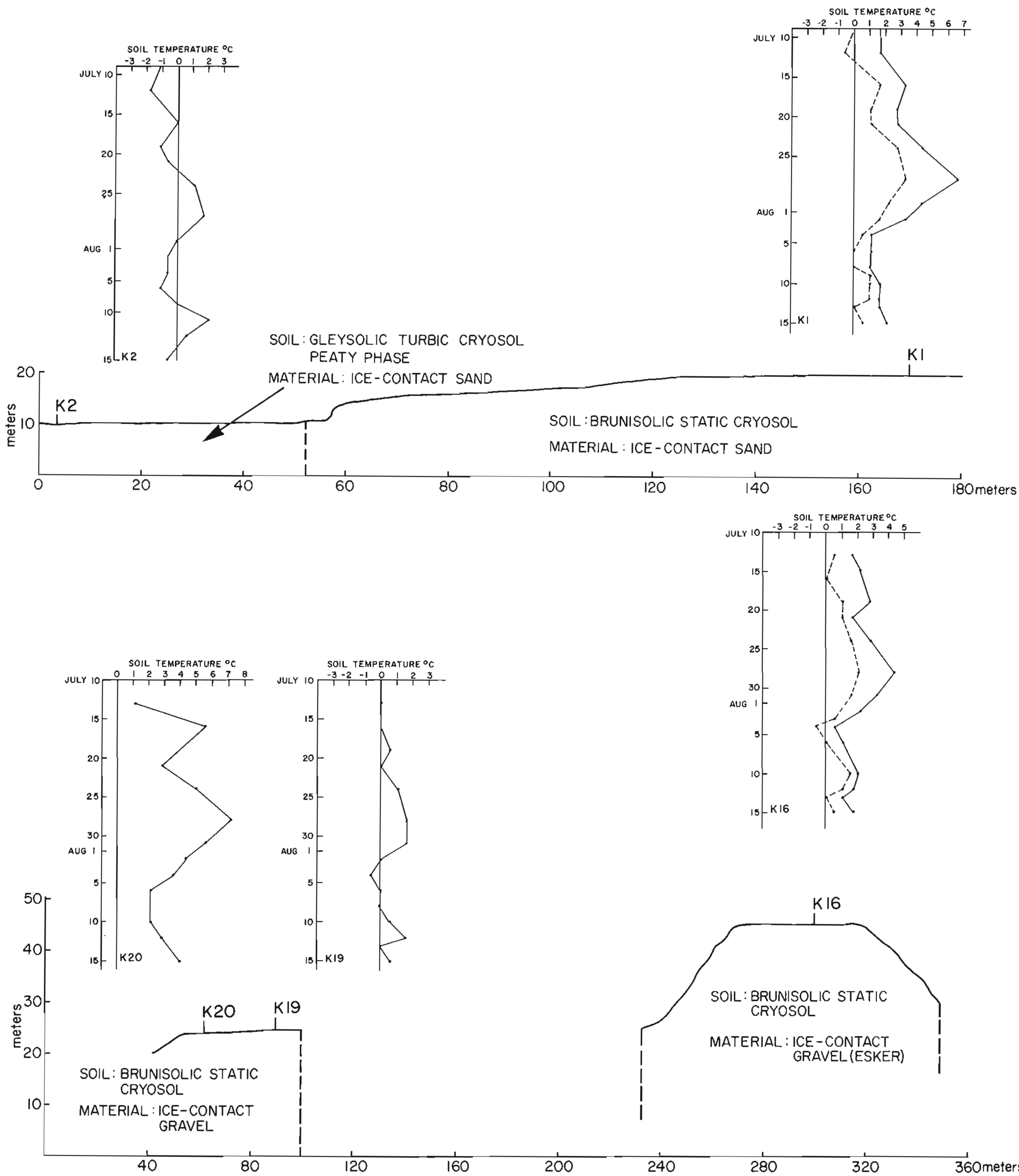

Figure 14. 2. Cross-sections showing the location of soil temperature sites K16 (66°04'N, 94°22'W), K19, and K20 (66°05'N, 94°22'W) on ice-contact gravel, and sites K1 and K2 (66°04'N, 94°21'W) on ice-contact sand. Note the soil temperature difference between K20 (disturbed) and K19 (undisturbed). Soil temperatures are given at 50 cm (——) and 100 cm (- - - -) depths.

rose at least 20 cm in the first week of August. In the second week of August the weather, and the soils, became warmer, and this warming trend lowered the frost table (Figs. 14.1 and 14.2).

Site K20 was located on a disturbed site; otherwise, the soil material was similar to that at site K19. The vegetation cover on site K20 was removed, and some of the surface soil was disturbed due to continuous travel over it by a Honda ATC motor tricycle. Site K20 showed much higher soil temperatures than site K19 (Fig. 14.2). For example, site K19 had temperatures, at the 50-cm depth, above 1°C on only 10 days and reached a maximum temperature of 1.7°C (July 28-31). On the other hand, the disturbed site continuously had temperatures above 1°C during the study period (34 days) with 33 days above 2°C, 15 days above 4°C, 4 days above 6°C, and a maximum temperature of 7.2°C on July 28.

One of the thermal characteristics of Cryosolic soils is that they have very little buffering capacity and respond quickly to a change in air temperature. This is due mainly to the shallow active layer, which provides only a small amount of stored energy. The underlying permafrost acts as a heat sink and continuously removes energy from the thawed layer of the soil. Hence, the slightest decrease in air temperature (due to storms or shortening of the hours of daylight) quickly lowers the soil temperature at all depths. This cooling causes the permafrost table to rise long before the surface freezes.

The moisture content of these sites was monitored by collecting core samples. The results of the analysis of these samples are not yet available and thus only general observations can be made. The moisture content of Gleysolic Turbic Cryosols was high throughout the summer, and they were saturated to the surface in the early part of July. The water table receded somewhat during the later part of July, but after rains (31 mm during the study period) they again became saturated to the surface.

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15. FRASER DELTA SEDIMENTATION, VANCOUVER, BRITISH COLUMBIA

Project 740062

John L. Luternauer
Terrain Sciences Division, Vancouver

Detailed grain size analyses have now been completed on 25 short (<1 m) cores obtained from the Fraser Delta tidal flats (Fig. 15.1) during the summer of 1975. The piston device used for recovering these samples was developed by the Department of Geological Sciences at the University of British Columbia during an investigation of the local sedimentary geochemistry (carried out under contract to this office). Figure 15.3 illustrates representative examples of these cores.

Core 24 (69 cm) was collected at the edge of the salt marsh lying between the Superport and Ferry causeways. The decrease in mean grain size from bottom to top probably reflects a gradual decrease in local wave energy and/or relative supply of sand which may be linked to the erection of the causeways. Fragments of wood (such as those at 20 and 50 cm) commonly are found in the sediments from this area. Throughout the core, clay constitutes no more than 14 to 20 per cent of the mud fraction.

Core 25a (72 cm) from a site to the north of the end of the sewage outfall channel causeway consists mainly of fine sand at its top and bottom. Silty layers and lumps at the midsection of the core reflect a radical, although temporary, change in the character of sedimentation at this location. The proportion of clay in the mud fraction in this part of the core ranges from 22 to 33 per cent. Throughout the core the grain size of the primary sand mode is 2.0 to 2.5Φ. Core 25b (76 cm) was collected 10 m to the north of where core 25a was obtained. A massive clayey silt lies at the bottom of the core. This unit is overlain by sands similar to those at the bottom of core 25a (although somewhat coarser). The midsection of the core consists of interbedded sand and silty mud layers much like those at an equivalent depth in core 25a. The core is capped by sands which appear to be identical to those at the top of core 25a. The proportion of clay in the mud fractions ranges from 13 to 24 per cent. The occurrence of silt layers at the bottom and midsection of this core indicates that in this area of the flats there may be rather dramatic regular cyclic variations in the sedimentary regime.

Core 33, obtained from the edge of a sedge marsh just north of the mouth of Main Channel, is strikingly different from those previously discussed. A progressive decrease in the sediment grain size from bottom to top is evident. Rootlets and other vegetative matter become more abundant near the top of the core. The proportion of clay in the mud fraction ranges from 16 to 21 percent. This core illustrates the sequence of changes which may accompany the establishment of a vegetated marsh upon the sandy outer tidal flats.

Core 35 (83 cm) was collected at the mouth of Middle Arm near patches of marsh vegetation. Near the top of the core fine rootlets are evident. Sediment throughout the core is generally sandy although silt laminations are clearly visible at various levels. The proportion of clay in the mud fraction ranges from 10 to 23 percent. Throughout the core the grain size of the primary sand mode does not vary. At the top of the core there are clear indications that the marsh vegetation is getting a foothold in this area. The sequences of silt laminations at several places in the core may indicate that marsh environments may have been established here earlier only partially, or were eroded away, or they simply may reflect "flood-plain" deposition at the channel margin.

Core 36 (96 cm) was collected near the edge of the sedge marsh approximately 1 km south of core 35. Numerous fine rootlets are present in the top 19 cm of the core. There is only very sparse vegetative debris scattered throughout the remainder of the core although shell fragments are common. Sediments within this core have grain size characteristics similar to those of core 35, i.e. they consist of fine silt laminations within a dominantly sandy sequence. Clay constitutes from 16 to 33 percent of the mud fraction. The grain size of the primary sand mode ranges from 1.5 to 2.0Φ to 3.0 to 3.5Φ. Thus conditions governing the deposition of sand at this site seem to be more variable than those at site 35 which actually is nearer the major local source of sediment (Middle Arm).

Core 37 (72 cm) was collected just north of the mouth of Canoe Pass inside a clump of marsh grass. The base of this core from 28 to 72 cm consists of alternating laminae and beds of sand and mud. There are only traces of vegetative debris. A section of massive silty sand extends from 20 to 28 cm. Above this to the 10-cm level, the sedimentary sequence consists of interlaminated mud and sand. The core is capped by a relatively massive clayey silt. Results of the grain size analyses of 3-cm sections of the core reveal that the proportion of sand gradually increases towards the midsection of the core then decreases fairly abruptly. The proportion of clay within the mud fraction ranges from 14 to 27 percent. At the site where this core was collected there appears to have been a gradual transgression of the sand flat environment which climaxed with the deposition of the massive sand at the midsection of the core. Following the deposition of the sand, energy levels at this site dropped considerably and, once again, silt and clay deposition resumed.

Core 38 (70 cm) was collected approximately 1 km south of core 37. The lower half of the core consists mainly of fine sand; the upper half consists of interlaminated silts and sands such as those in core 37. The proportion of clay in the mud fraction ranges from 15 to 18 per cent. Within the sandy segments of the core the grain size of the primary sand mode ranges from 2.0 to 2.5Φ to 2.5 to 3.0Φ. This core graphically

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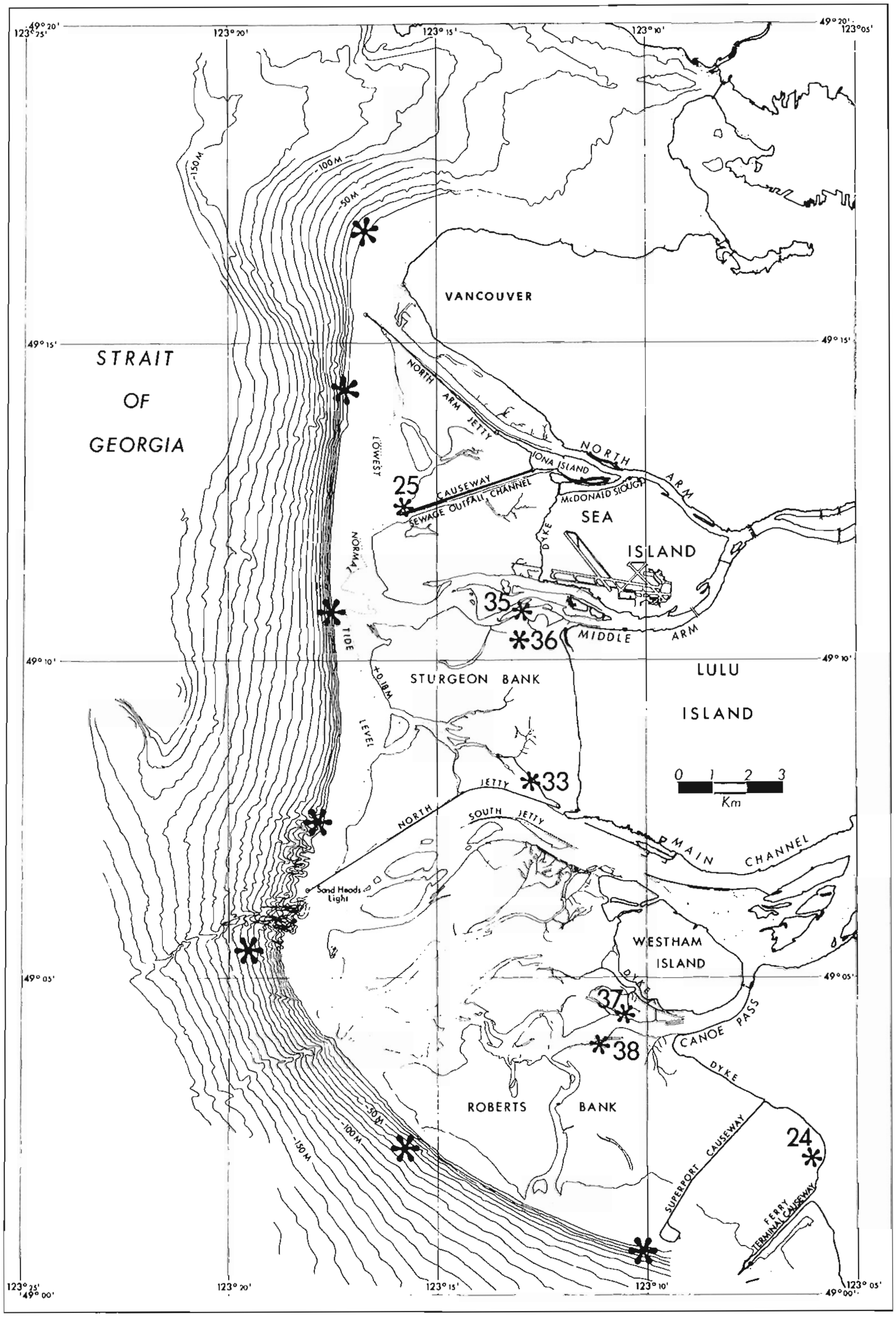


Figure 15. 1. Locations where short (<1 m) piston cores described in Figure 15. 3 were obtained (summer 1975) from the Fraser River Delta tidal flats (numbered asterisks) and where vibro-cores (approx. 18 m) were obtained (July 19-22, 1976) from the head of the Fraser River Delta slope (larger, unnumbered asterisks). The vibro-cores are being analyzed under contract to Dr. J. W. Murray, Department of Geological Sciences, University of British Columbia.

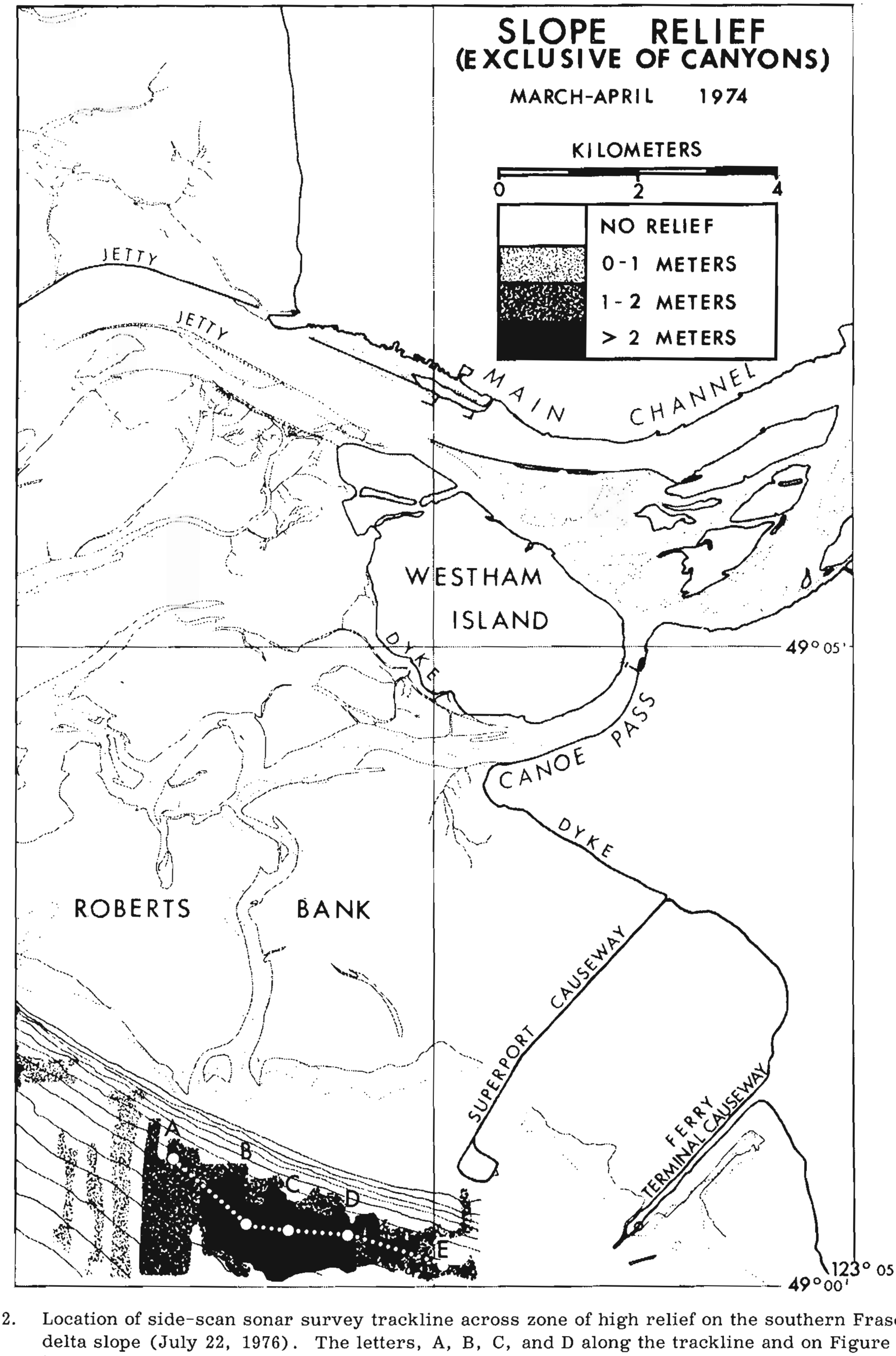


Figure 15.2. Location of side-scan sonar survey trackline across zone of high relief on the southern Fraser River delta slope (July 22, 1976). The letters, A, B, C, and D along the trackline and on Figure 15.4 indicate corresponding segments on echo-sounding and side-scan records. Parallel contours on slope start at 10 m and continue at 10 m intervals.

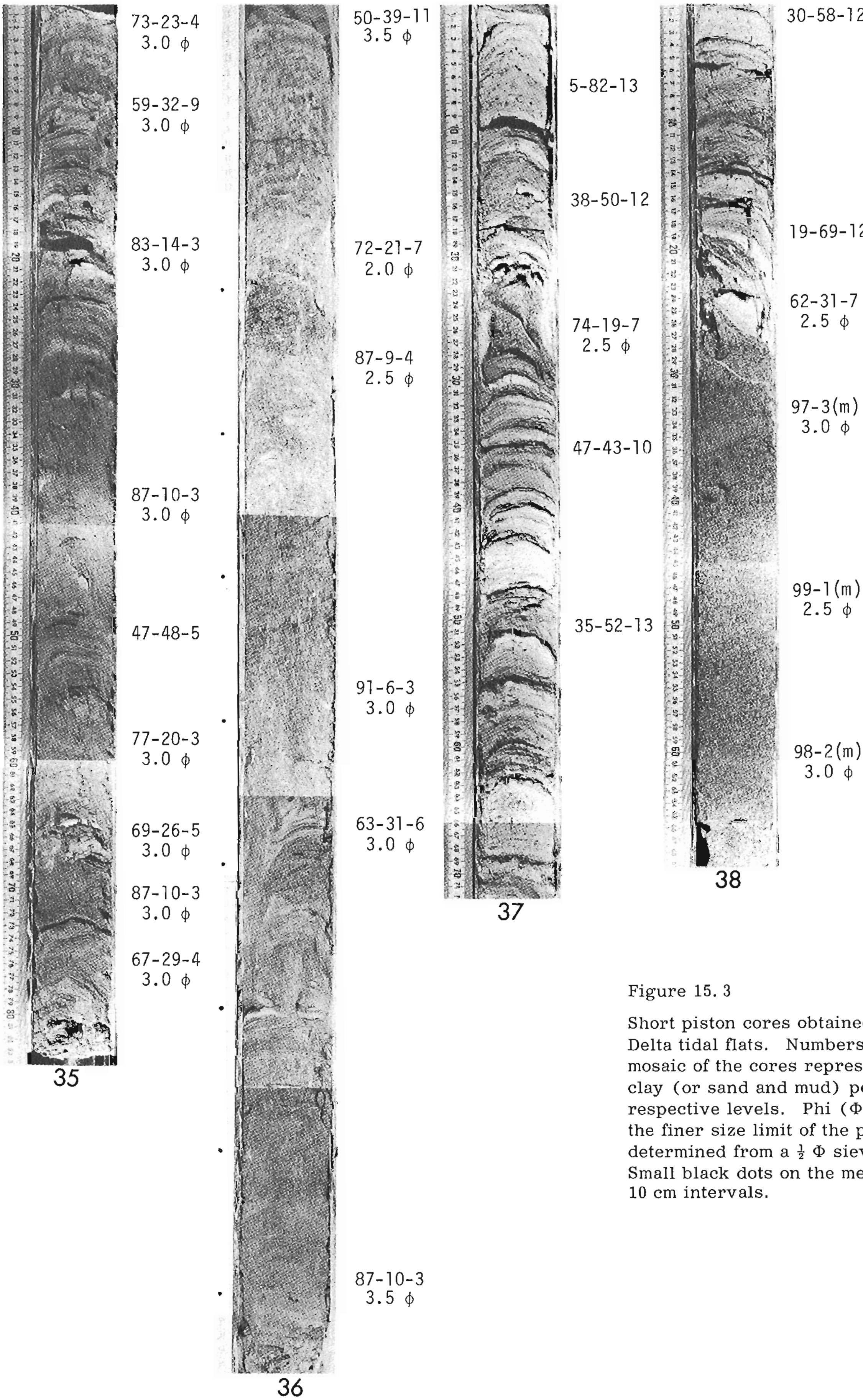


Figure 15. 3

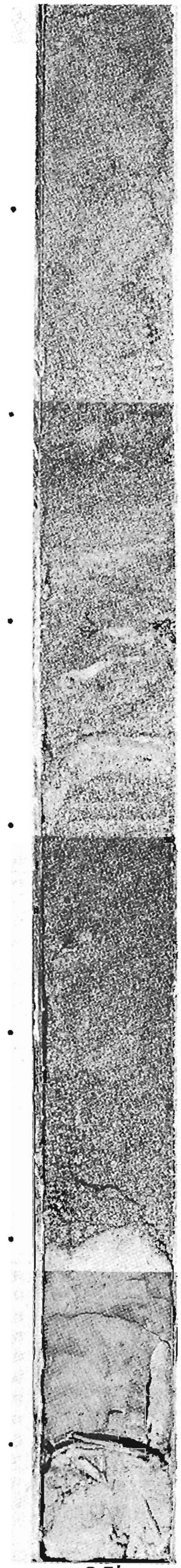
Short piston cores obtained from the Fraser River Delta tidal flats. Numbers to right of the photo mosaic of the cores represent sand, silt, and clay (or sand and mud) percentages at the respective levels. Phi (Φ) values indicate the finer size limit of the primary sand mode determined from a ½ Φ sieve-interval analysis. Small black dots on the metre stick indicate 10 cm intervals.


31-59-10
39-52-9
22-63-15
47-44-9
49-41-10

24


97-3(m)
2.5 ϕ
97-3(m)
2.5 ϕ
79-15-6
2.5 ϕ
58-28-14
2.5 ϕ
51-38-11
2.5 ϕ
96-4(m)
2.5 ϕ
98-2(m)
2.5 ϕ
98-2(m)
2.5 ϕ

25a


97-3(m)
2.5 ϕ
96-4(m)
2.5 ϕ
76-21-3
2.5 ϕ
71-24-5
2.0 ϕ
97-3(m)
2.0 ϕ
2-74-24

25b

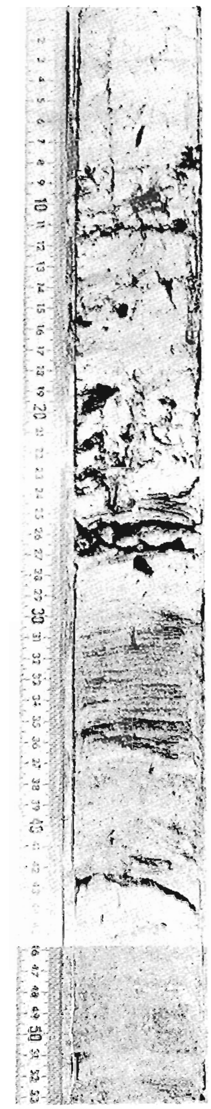
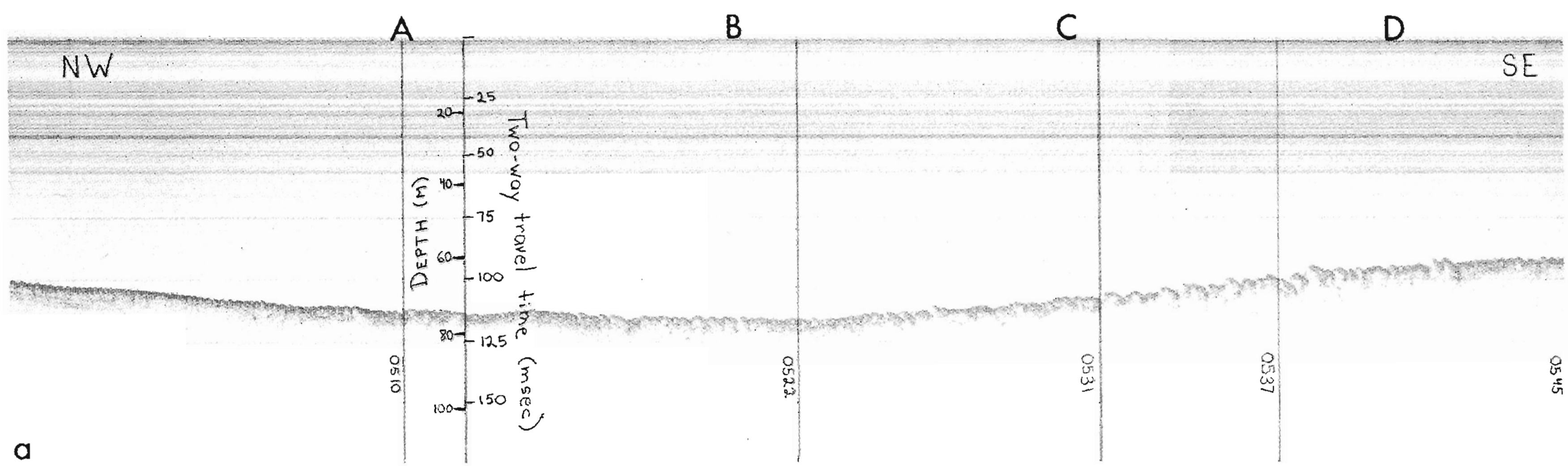
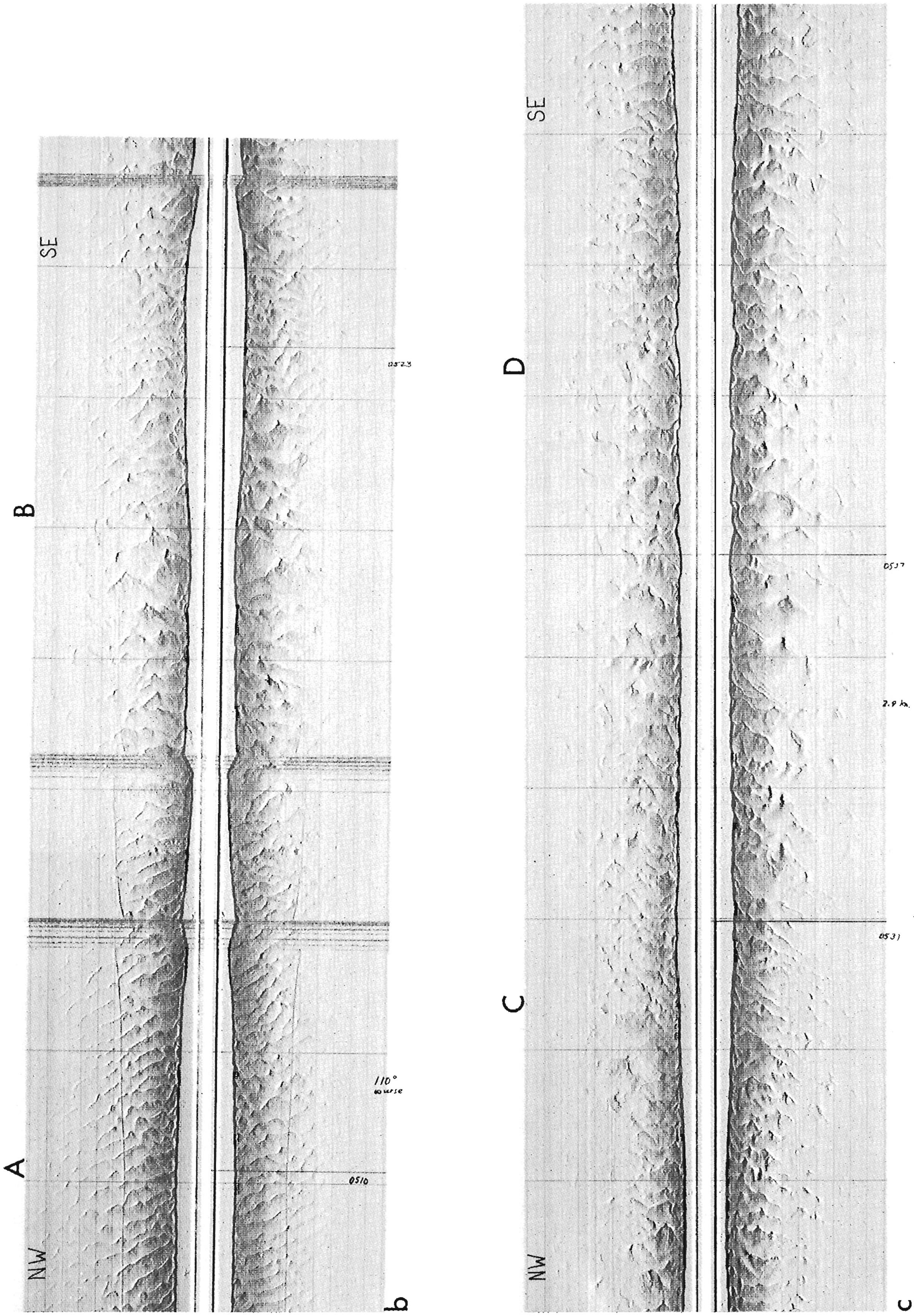

7-74-19
10-71-19
45-46-9
29-59-12
46-46-8
54-38-8
3.5 ϕ

Figure 15. 4

a. High resolution echo-sounding record of the sea floor obtained along the trackline shown in Figure 15. 2 (July 22, 1976).

b, c. Side-scan sonar record obtained along the trackline shown in Figure 15. 2. Record spans 150 m to each side of the median strip. Upper case letters A, B, C, and D in this figure and in Figure 15. 2 indicate corresponding sections.




NW
A
B
SE
110° course
b
NW
C
D
SE
c

illustrates how abrupt the transition from sand to mud flat at any one site may be.

It is apparent that sedimentary processes on the Fraser Delta tidal flats can be exceedingly variable both in time and space. It should be emphasized, however, that variations in one aspect, such as mud concentration, may not be accompanied by significant changes in another variable, such as grain size of the primary sand mode (cores 25a and 35).

It is also important to note that clay is the minor component of the mud fraction. This may be significant because it is possible that development-induced alterations to river flow may more substantially affect the silt than the clay supply to biologically vital marshes.

Deeper coring of the tidal flats should reveal what normal, longer term fluctuations in sedimentologic regimes are and allow one to better distinguish development-induced changes from expectable natural variations. As yet no determinations have been made of sedimentation rates on the flats.

Vibro-cores were obtained at the edge of the tidal flats and/or top of slope (Fig. 15.1) from the *C.F.A.V. Endeavour* during the period July 19-22, 1976. When analyzed, these should reveal what has been the recent depositional history of the most dynamic zone on the delta.

A preliminary reconnaissance side-scan survey of one of the zones of high relief on the delta slope (Luternauer, 1976) (Fig. 15.3 and 15.4) was performed on July 22, 1976. It is apparent from the records obtained that the relief can be attributed to the presence of mega-ripples on the sandy sea floor. It is suspected that tidal currents are largely responsible for these features evident at approximately 70 to 90 m depth. It is not clear why the features exhibit such a range in size or why they tend to develop only on specific sections of the slope. A further, more detailed side-scan survey coupled with an investigation of the oceanographic regime over the area is planned.

Acknowledgments

Ms. Lesley Simpson of the Terrain Sciences laboratory in Vancouver performed all grain-size analyses of the cores. Side-scan sonar equipment was kindly supplied by the Canadian Hydrographic Service.

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Project 740046

P. J. Kurfurst
Terrain Sciences Division

A continuing research program consisting of laboratory and field measurements of certain physical properties of frozen soils and rocks is being carried out by Terrain Sciences Division, Geological Survey of Canada in order to acquire geological and geomorphological information concerning northern terrain and its behaviour. As a part of this program a detailed study of the acoustic properties of frozen soil samples which differ in type of material, ice content, and geographical location was originated in spring of 1976.

A special apparatus has been installed in the temperature-controlled "cold" room located in the Geological Survey building, 601 Booth Street, Ottawa to measure ultrasonic compressional and shear wave velocities of frozen specimens. A block diagram of the apparatus is shown in Figure 16.1. The apparatus consists of a loading frame, fluid-pressure control panel, and electronic equipment. The loading frame is connected to the fluid-pressure control panel, and a pressure intensifier is used to achieve the required pressure. A set of specially designed transducer holders is placed in the loading frame, and a specimen is placed between transducer holders which are connected to the electronic equipment. A detailed description of the compressional and shear wave transducer holders was reported by King (1970).

For the ultrasonic measurement tests it is necessary to prepare 5 cm-diameter specimens, at least 10 cm long, from larger core samples in their frozen state. A multipurpose grinder or a lathe, installed in a temperature-controlled room, is used to grind the cylindrical and flat surfaces of the specimens to the required dimensions. Details of the sample preparation are similar to those described by Kurfurst and King (1972).

The electronic equipment consists of a pulsed oscillator, time mark generator, amplifier, and oscilloscope.

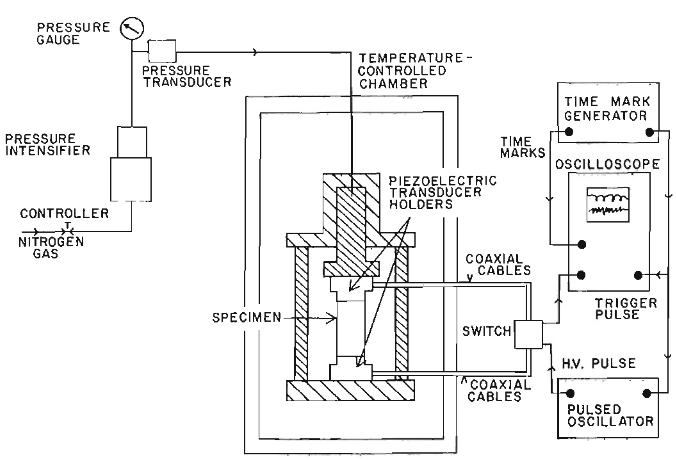

Figure 16.1. Block diagram of acoustic velocity apparatus.

The ultrasonic wave velocities are obtained by measuring the time taken for a pulse to traverse the length of the specimen. A pulsed oscillator is used to apply a pulse to the transducers, which in turn transmit this pulse through the sample. The electrical signal produced is amplified and displayed on a dual-beam oscilloscope where the trace of the first arrival of the signal is either measured or photographed. From the values obtained compressional and shear wave velocities are calculated.

Frozen soil samples, which differ in material and ice content, are being tested at temperatures ranging from -1 to -7°C; the samples are subjected to a uniaxial state of stress. An increase in both compressional and shear wave velocities occurs as the temperature is reduced below 0°C due to formation of ice in large pore spaces. As the ultrasonic wave velocity is much higher in ice than in water (compressional wave velocity is about 3.0 km/s in polycrystalline ice as compared to 1.49 km/s in water), the observed velocity for a frozen soil is greater than that for the same sample thawed. The rate of velocity increase is proportional to the increase in the amount of ice within one soil type, so the velocities measured are indicative of the amount of ice in the specimen.

Using a strain indicator and/or the deflection dials, the testing apparatus can be used to measure stress-strain relationships, porosity-ultrasonic velocity-shear strength relationships, and the effects of stress cycles on soil samples at various temperatures.

The results obtained indicate that the laboratory ultrasonic wave measurements can:

(a) provide a measure of control in the interpretation of field records in permafrost areas;
(b) help to delineate the extent of permafrost in conjunction with the results of field shallow seismic and DC resistivity surveys; and
(c) help to estimate, using the existing results of shallow seismic surveys, the ground-ice content in frozen soils and rocks and thus reduce the amount of drilling necessary for large construction projects such as highways, pipelines, etc.

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RECONNAISSANCE STUDIES OF BIG RIVER, BANKS ISLAND, DISTRICT OF FRANKLIN

Project 750079

T.J. Day and C.P. Lewis
Terrain Sciences Division

Introduction

This report presents both an approach to the reconnaissance study of large river systems and some preliminary results from the application of that approach to the Big River system on Banks Island in the Arctic Archipelago (Fig. 17.1). The division of Big River into major reaches based on morphology, the identification of typical subreaches, the types of field data obtained from the subreaches, and the relationship between the morphologic divisions and statistically identifiable segments of the river-long profile are discussed.

Big River has its headwaters in eastern Banks Island. From its origin in small lakes and hummocky terrain of the silty till deposits associated with the western maximum of the Wisconsin Laurentide Ice Sheet, the river flows northwest for 120 km then, turning in a more westerly direction, it continues for another 120 km until reaching the Beaufort Sea at approximately 72°30'N, 125°20'W.

Downstream of the Wisconsin ice limit, Big River flows in a broad shallow valley typical of western Banks Island. At least four discontinuous terrace surfaces can be identified. The lowest alluvial surface bordering the present channel exhibits a relict braided channel pattern and may represent sedimentation during the Wisconsin. Within this surface a further paleochannel of meandering form, which postdates the braided pattern, can be seen. Underlying present alluvial deposits (and possibly Quaternary deposits as well) are sands and gravels of the Beaufort Formation (Tertiary).

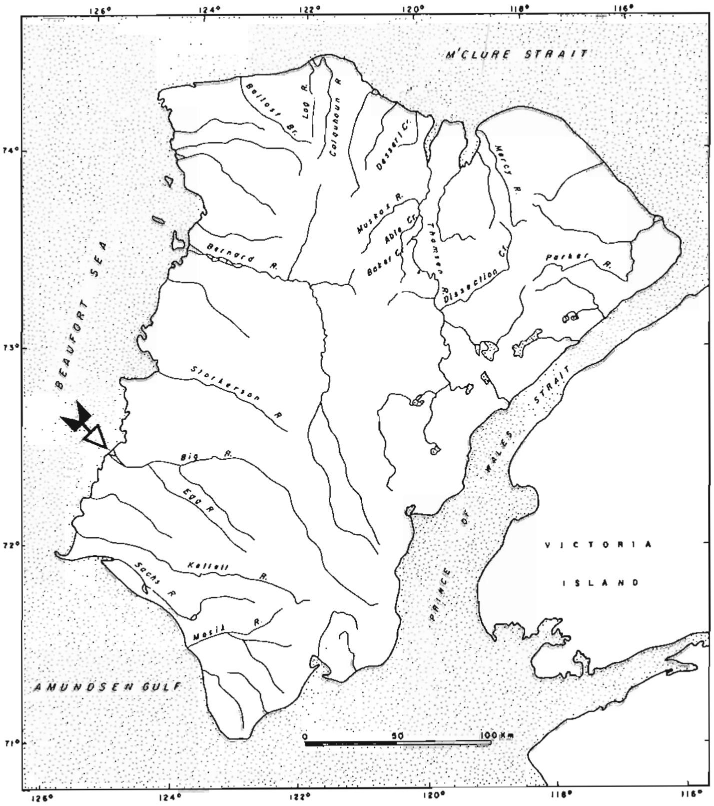


Figure 17.1. Map of Banks Island showing location of Big River (arrow).

Reaches Identified from High Altitude Aerial Photographs

Initial analysis of Big River was based on high altitude aerial photographs at a scale of 1:100 000. A mosaic of the river is shown in Figures 17.2A and 17.2B. Each figure covers a length of about 120 km, with the division in approximately the position of the major change in flow direction. On the basis of variations in channel pattern and lateral channel activity, as seen in these photographs, the river was subdivided into six reaches. The downstream end of each reach is indicated by an open arrow and number on the left bank (looking upstream) in Figures 17.2A and 17.2B. Descriptions of channel pattern, lateral activity, along with reach length, mean slope, and the relative width of the active channel compared to the low flow channel visible in the photographs, are listed in Table 17.1. Definition of channel pattern and lateral activity are those used by Bray and Kellerhals (1972) (or more recently, Kellerhals *et al.*, 1976). Channel pattern refers to the entire active channel, not just the low flow channel shown in the photographs.

Within each reach a short, typical, and geomorphologically homogeneous subreach was chosen for field study. Data collected within each subreach include: surveyed cross-sections (usually 4 to 6); a surveyed long profile along the thalweg; water width, depth, and surface slope at the time of survey; active layer depths under at least one surveyed cross-section; grain size data on bed, bank, and surface channel armour material; and qualitative observations on bedforms, terrain features, nature of the valley flat, etc. The locations of the subreaches are shown by diamond markers in Figures 17.2A and 17.2B. Photographic enlargements of each subreach are shown in Figures 17.3 and 17.8 for subreaches 1 to 6 respectively (the paleochannel patterns can be seen in Figures 17.3 to 17.6). A cross-section for each subreach is shown in Figure 17.9, and a summary of data is given in Table 17.2.

From: Report of Activities, Part A;
Geol. Surv. Can., Paper 77-1A (1977).

Figure 17. 2A. Aerial photograph mosaic showing Big River from its mouth to approximately 120 km inland. Open arrows and numbers on the left bank (looking upstream) indicate the downstream ends of the channel reaches distinguished on the basis of channel morphology and lateral activity. Solid arrows and numbers on the right bank indicate the downstream portions of the channel reaches distinguished on the basis of the segments of the channel long profile. The diamond markers locate the position of the field surveys of subreaches. Channel elevations at 50 m intervals also are shown.

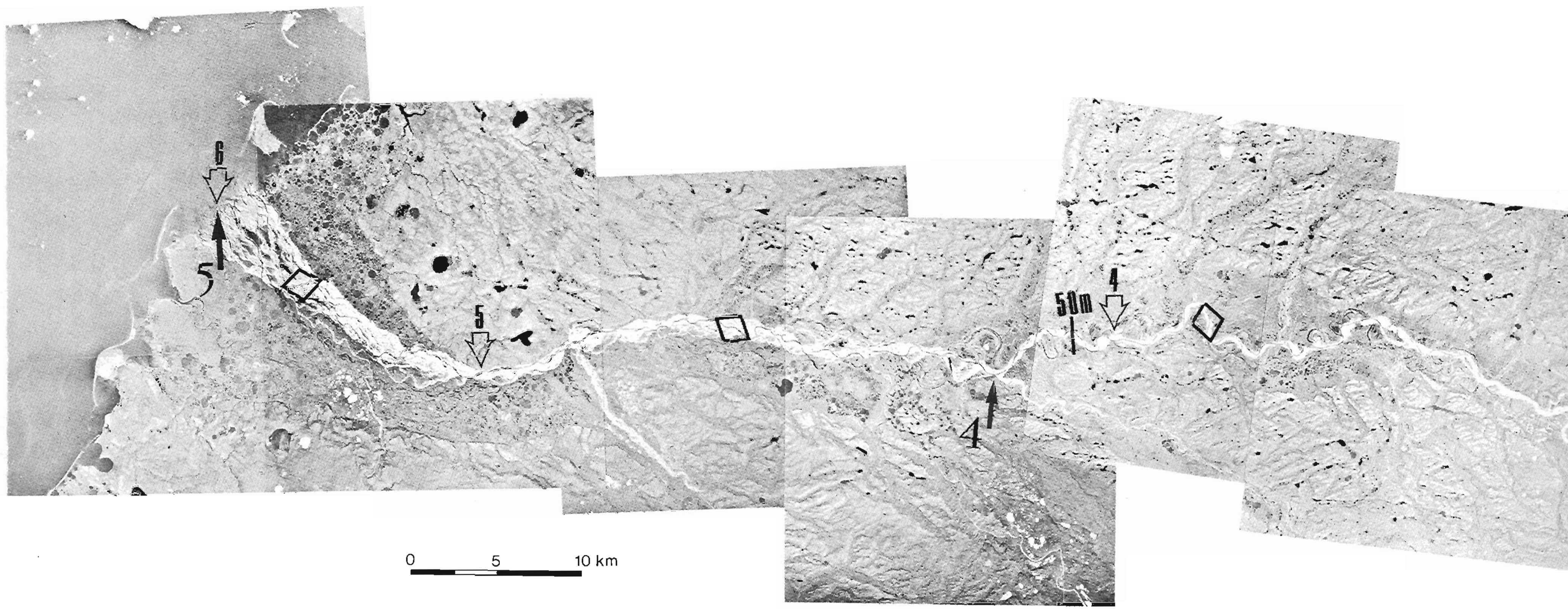
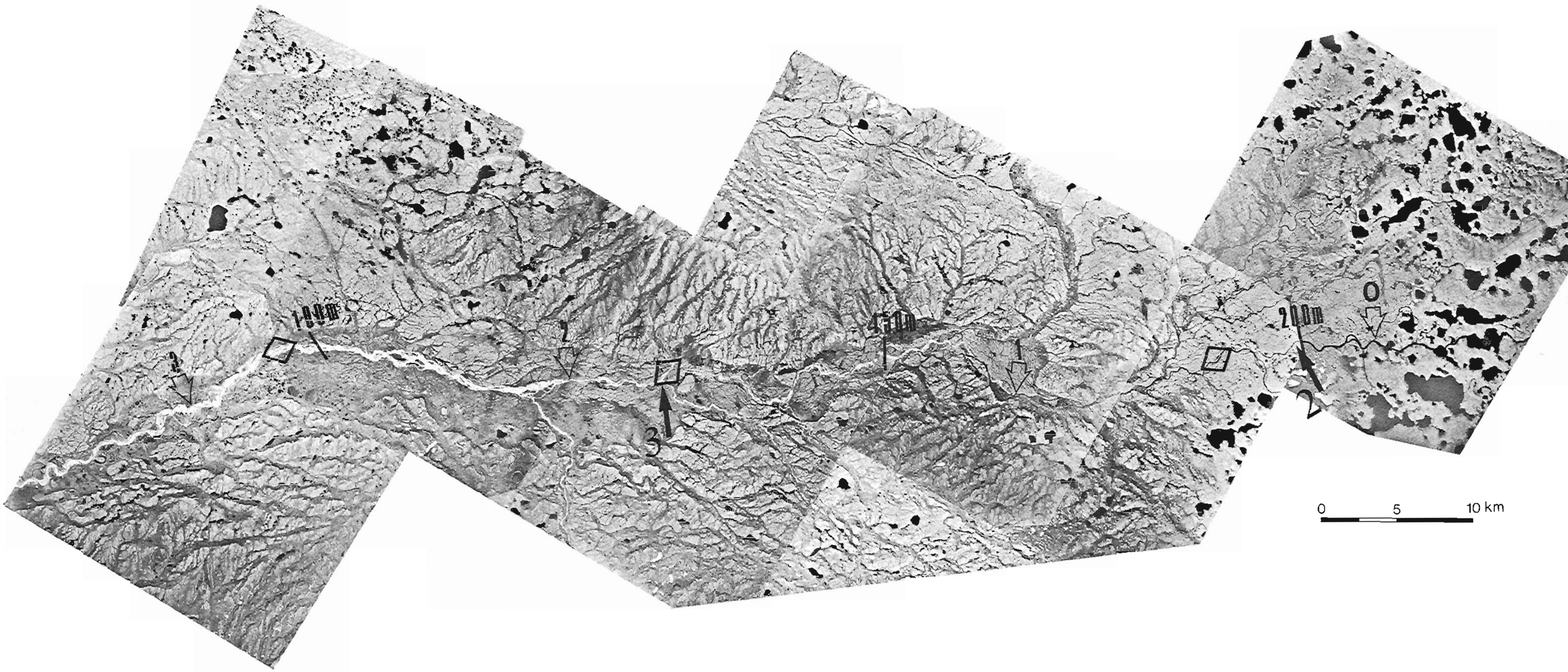


Figure 17. 2B. Aerial photograph mosaic showing the upper 50 km of Big River. Definition of symbols is the same as in Figure 17. 2A.



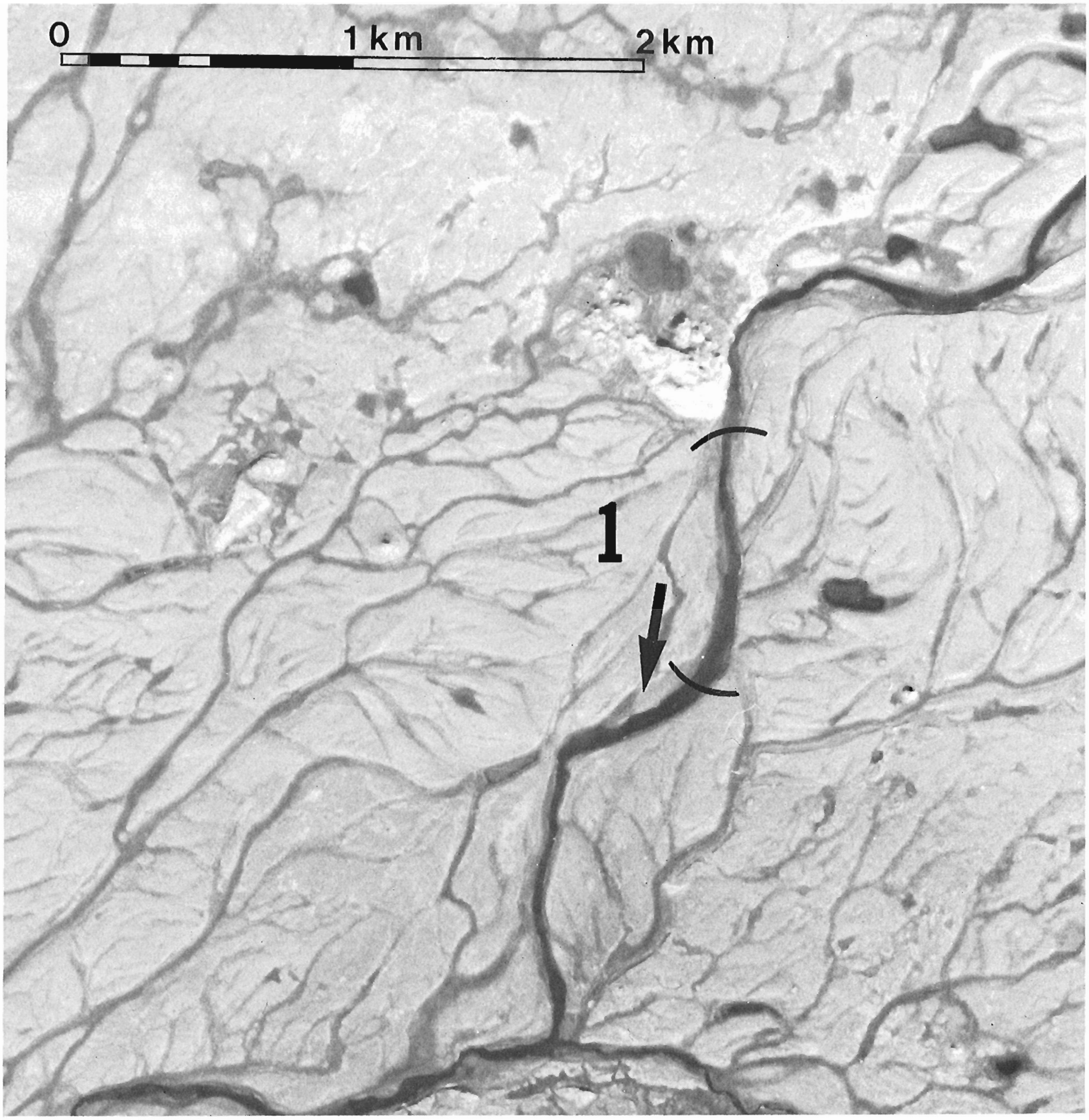


Figure 17. 3. Enlargement of aerial photograph (A-17057-58) showing location of subreach No. 1.

Table 17. 1. Summary of Reach Characteristics

Reach	Length km	Mean Slope	Channel Pattern	Lateral Activity	Ratio of Flood Plain Width to Low Flow Channel Width
0-1*	16. 8	0. 0052	irregular	not detectable	1
1-2	38. 5	0. 0010	irregular	irregular	2
2-3	28. 8	0. 0011	sinuous	irregular	3-5
3-4	56. 3	0. 0007	irregular meanders	downstream progression	2-3
4-5	42. 5	0. 0007	sinuous	irregular	9-13
5-6	22. 5	0. 0009	straight	avulsion	≈25

*Note that this morphologic reach does not extend into the small lakes region (Fig. 17. 2B)

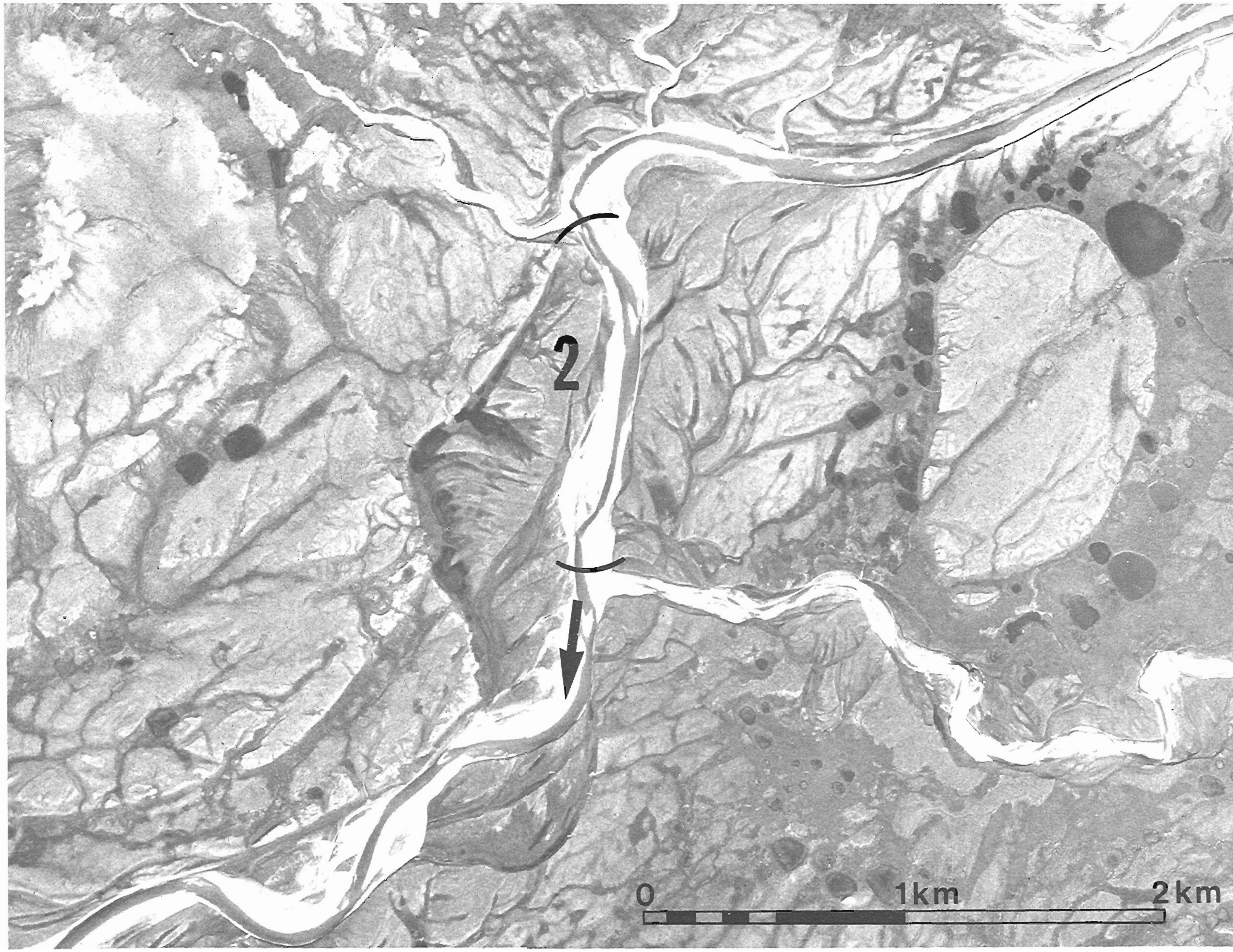


Figure 17. 4. Enlargement of aerial photograph (A-17057-38) showing location of subreach No. 2.

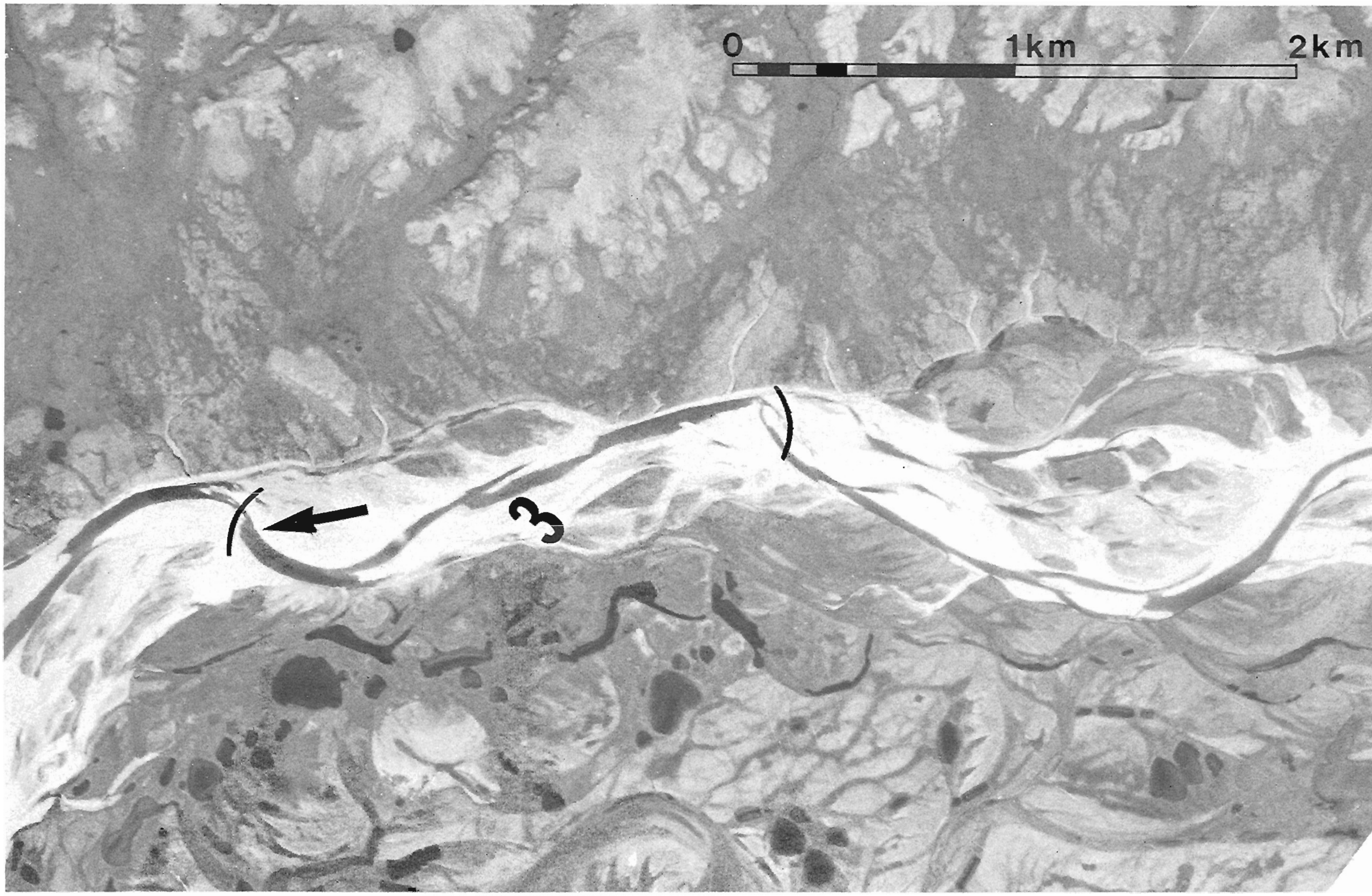


Figure 17.5. Enlargement of aerial photograph (A-17057-16) showing location of subreach No. 3.

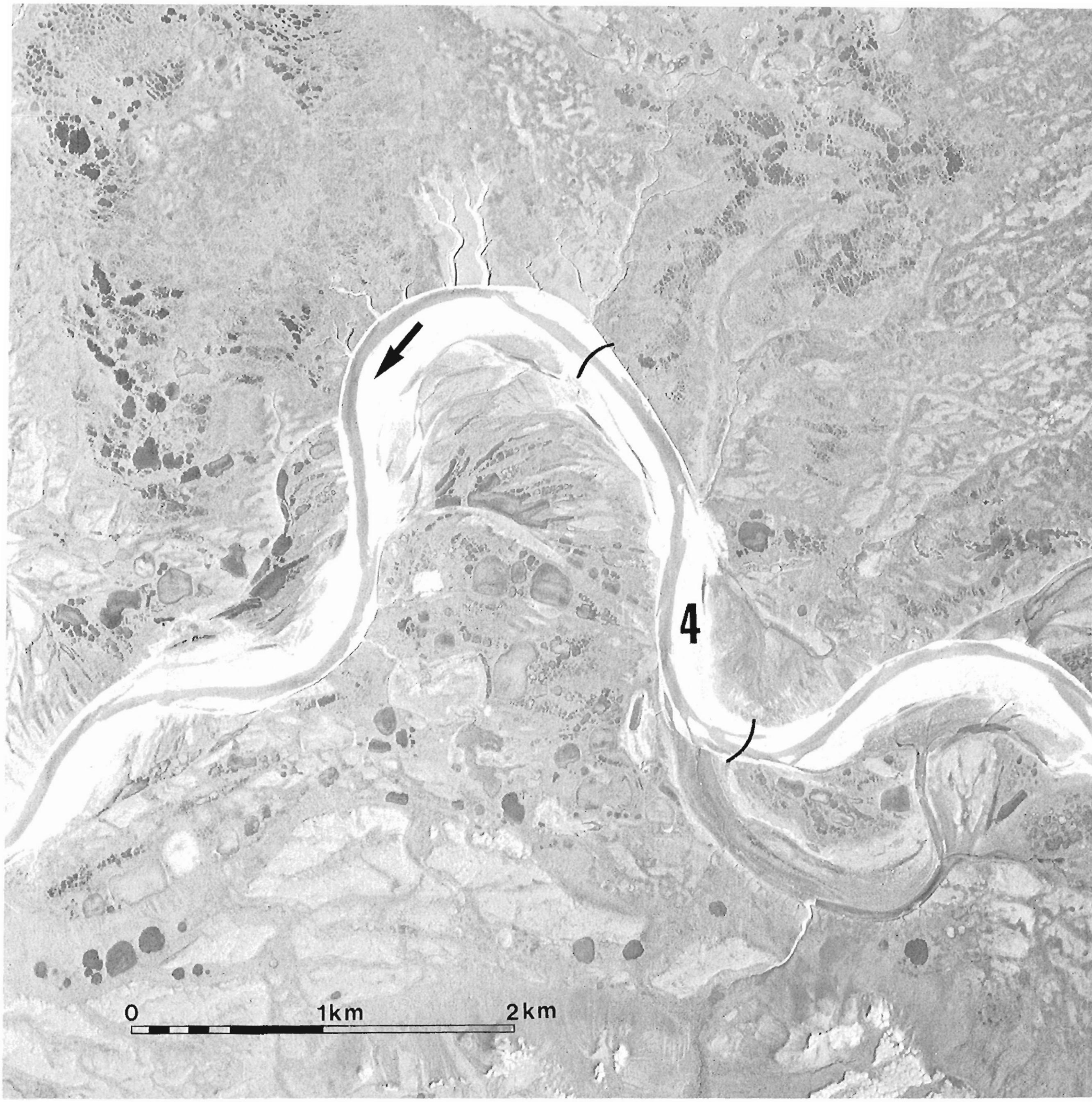


Figure 17.6. Enlargement of aerial photograph (A-17381-95) showing location of subreach No. 4.

Table 17.2. Summary of Subreach Data

Subreach	Length m	Slope*	Mean Active Channel Width* m	Mean size Bed Material m	Mean size Armour Material m
1	1200	0.0003	45	8.6	54
2	650	0.0014	155	12.0	27
3	1100	0.0007	240	9.0	23
4	800	0.0014	180	8.0	30
5	1100	0.0011	170	8.6	17
6	850	0.0011	130**	8.6	35

*Approximate
**Channel survey in an actively braiding reach, width refers to definable channel only.

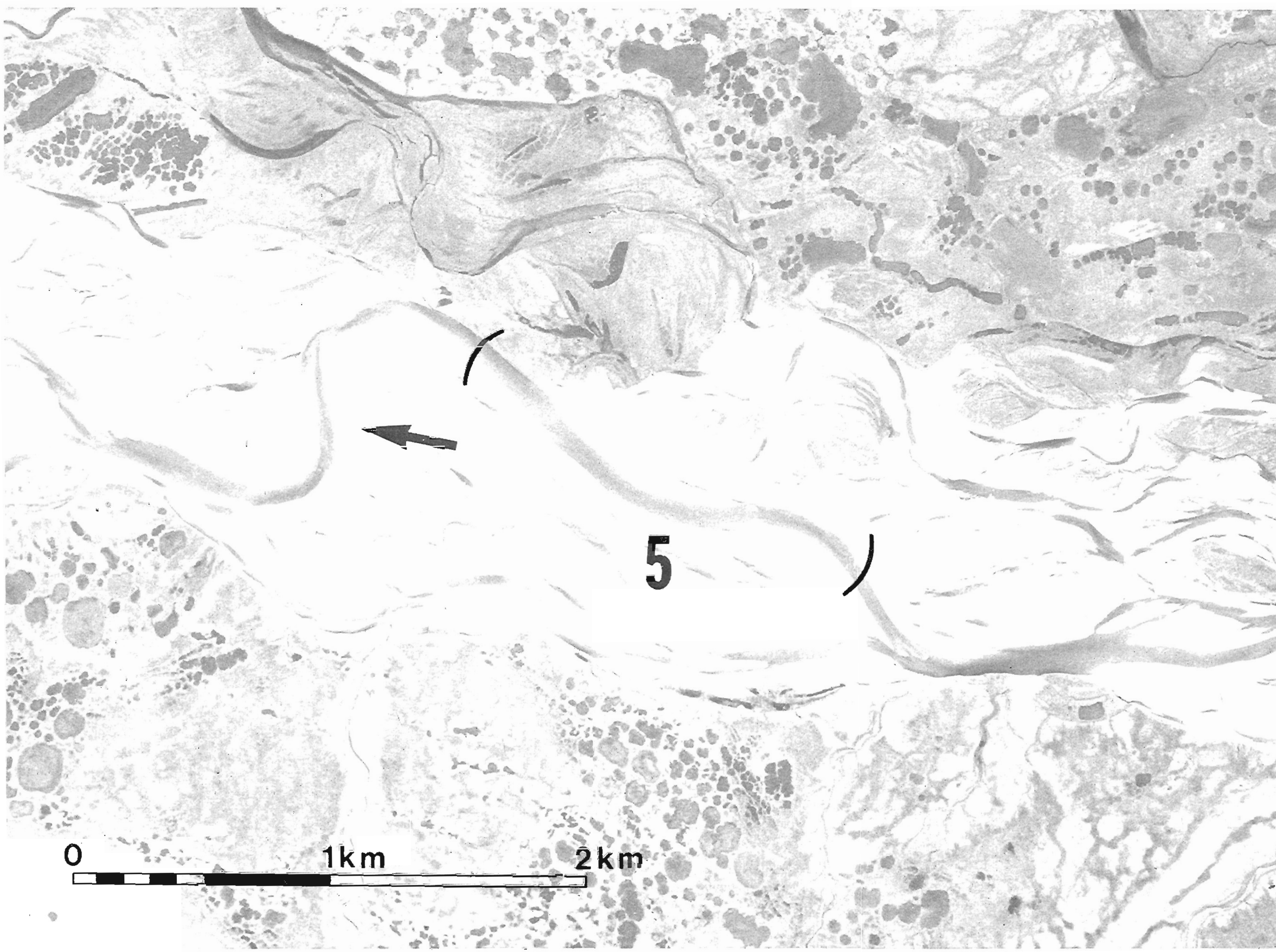


Figure 17.7. Enlargement of aerial photograph (A-17375-24) showing location of subreach No. 5.

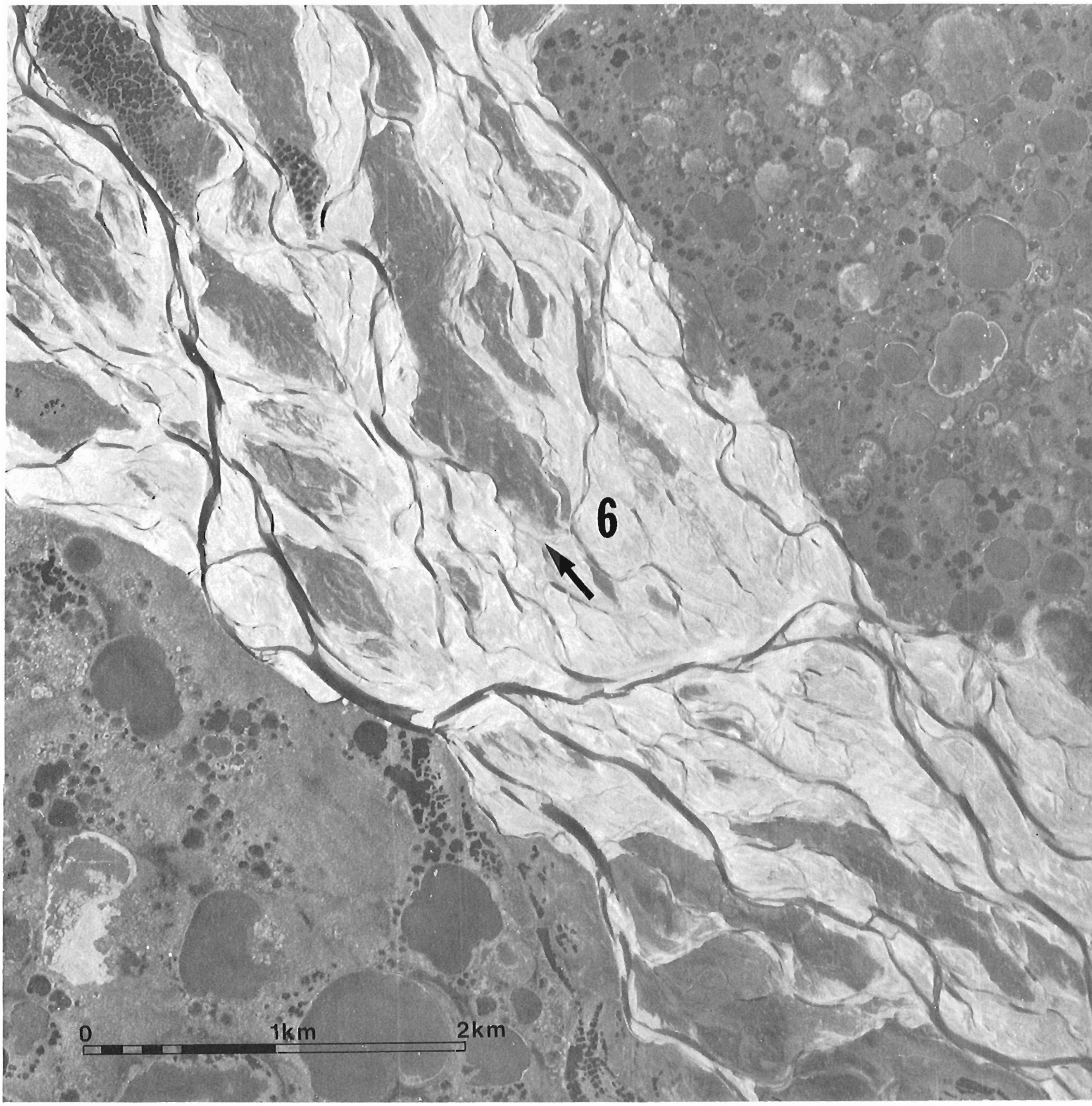


Figure 17. 8. Enlargement of aerial photograph (A-17381-29) showing approximate location of subreach No. 6.

Table 17. 3. Longitudinal Profile Analysis

Reach	Length km	No. of Data Points	Best Fitting Equation	RSQ Value, %
0-1	2.5	7	$y = 358.46 - 19.07x$	97.2
1-2	30.5	15	$y = 326.08 - 6.61x + 0.09x^2$	97.2
2-3	63.3	11	$y = 246.83 - 2.47x + 0.01x^2$	93.9
3-4	91.2	9	$y = 298.10 - 2.51x + 0.01x^2$	99.8
4-5	54.9	6	$y = 179.71 - 0.73x$	96.9

Long Profile Analysis

The general structure of the long profile of Big River was determined by photogrammetric height and distance measurements in order to examine the relationship between profile characteristics and surface morphology. The smallest possible height interval which could be resolved was 7.7 m (25 feet), with an accuracy of ± one interval. The resulting long profile is shown in Figure 17.10A. The locations of morphologic units and survey sites are indicated.

On first inspection the long profile shows a concave shape, typical of rivers possessing either a downstream decrease in mean size of bed material or an aggrading behaviour in the lower reaches. The former explanation is not possible, however, as mean size of the bed material does not show an exponential decrease downstream (Table 17.2). The river does show some characteristics of aggrading behaviour, with degradation in the upper reaches (incision in the paleo-outwash deposits) and braiding channel patterns in the lower reaches. The presence of vegetative islands within

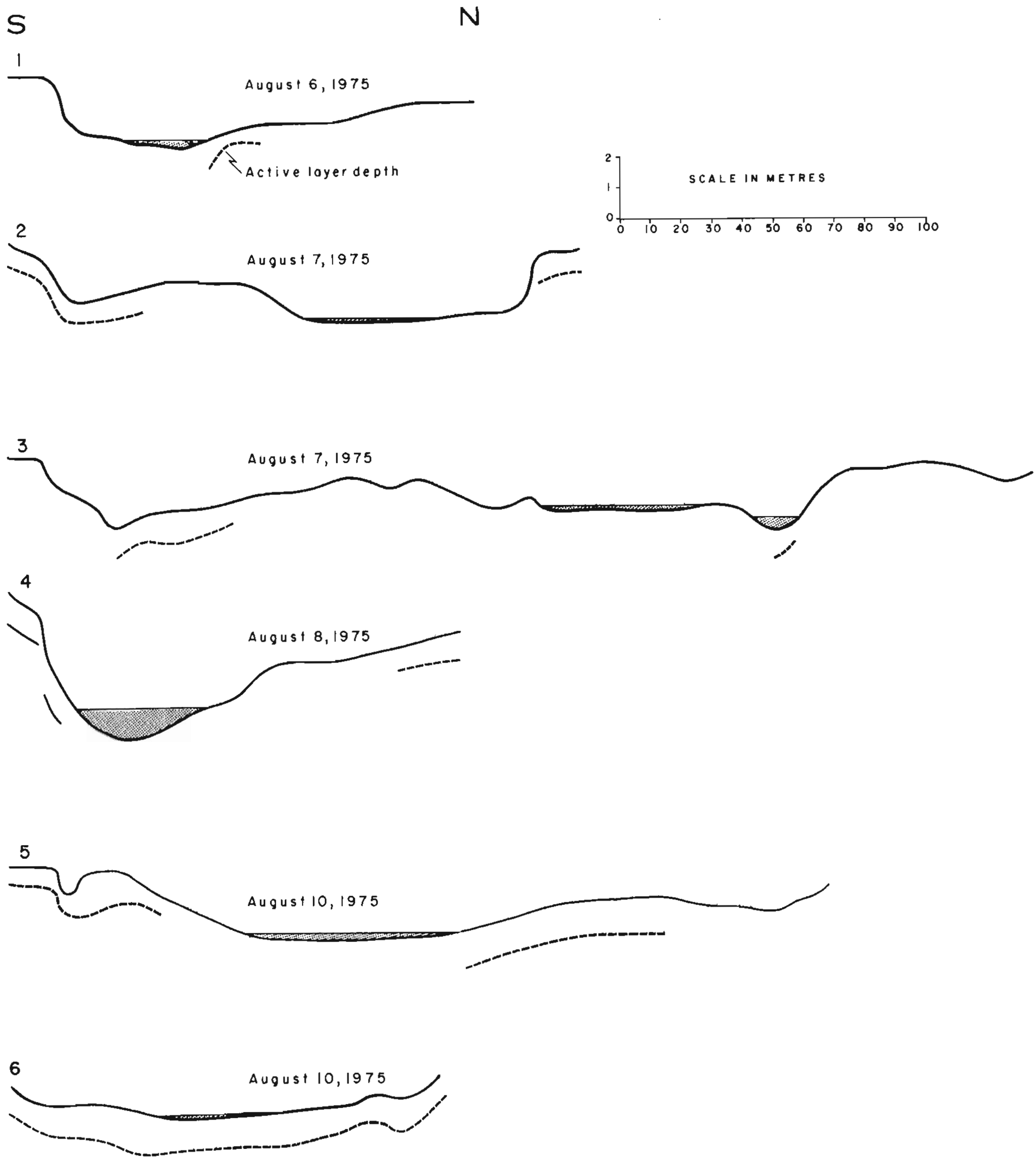


Figure 17.9. Typical channel cross-sections for each surveyed subreach. Active layer depths are shown by dashed lines.

the flood plain near the river mouth (Fig. 17.2A) and the convexity of the profile through the lowermost kilometres (Fig. 17.10A), however, are indicative of lateral channel stability and degradation respectively. This suggests that the entire river cannot be looked at as a unit; rather, the long profile may exhibit differing properties along its length and must be viewed in this context.

In order to identify homogeneous profile segments, a second order parabola of the form

$$y = 305.50 - 2.71 + 0.01\, x^{2} \qquad (1)$$

where y is height above sea level and x is distance from the headwater, was fitted to the complete profile and variations about this mean trend were examined.

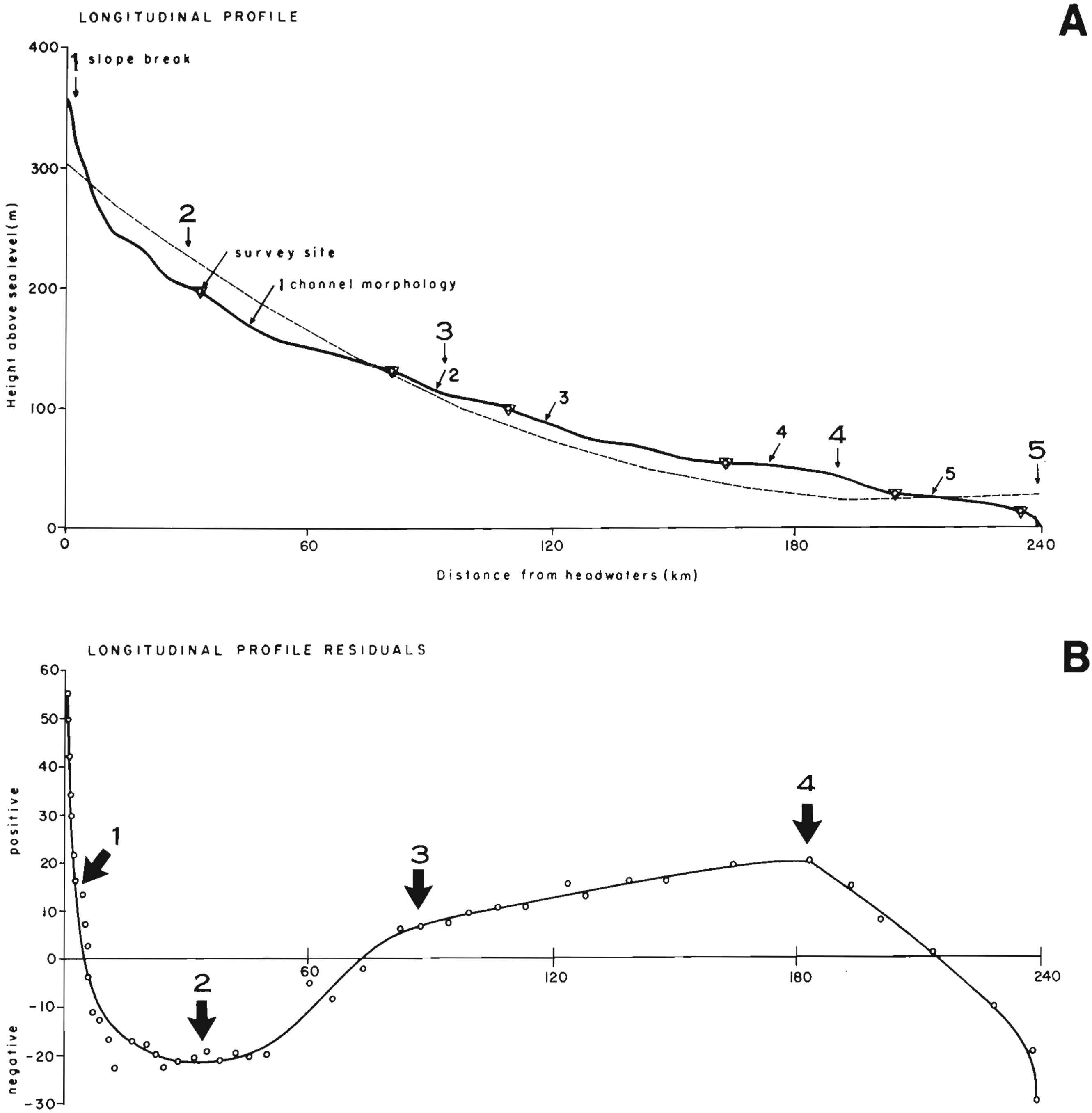


Figure 17.10. Longitudinal profile data for Big River:
A. the long profile of Big River showing the channel reaches distinguished on the basis of long profile segments and channel morphology, the subreach survey sites, and the position of the parabola (equation (1)) fitted through the profile;
B. a plot of residuals from the fitting of a parabola to the long profile data.

This trend is shown as a dashed line in Figure 17.10A. The residuals from this regression, plotted in Figure 17.10B, show strong autocorrelation; a runs test indicates a nonrandom sequence (at the 5% level of significance). Four major changes in direction of the value of the residuals are shown in Figure 17.10B, and these indicate five profile segments. The location of the downstream end of each profile segment is indicated by the solid arrows and numbers on the right bank in Figures 17.2A and 17.2B for comparison with the morphologic reaches previously defined.

The shape of each profile segment was investigated through the fitting of linear or second order parabolic equations; the best fitting equations are listed in Table 17.3. Profile segment 0-1 has a rectilinear shape which signifies a graded reach where sediment input equals sediment through-put. On physical grounds, however, this interpretation must be rejected as the profile segment is composed of short channels flowing between small lakes and is not hydraulically homogeneous. The second profile segment, 1-2, has a concave (parabolic) profile, the physical significance of which has been mentioned in the preceding paragraph. Both explanations are inapplicable as this segment possesses a stable channel (Fig. 17.3) with a heavily armoured bed (cf. columns 5 and 6 of Table 17.2).

Profile segment 2-3 also shows a parabolic form. The upper third of this segment (Fig. 17.2B) has a stable channel, and the lower portion shows an increased active channel width. Segment 3-4 has a similar profile shape. Both segments end near the junction of tributaries, although not every tributary junction results in a profile discontinuity. The last segment, 4-5, has a linear shape. Explanation of this segment as an equilibrium channel is inappropriate as the upper portion is braiding and the last lower portion appears to be degrading (presence of vegetated areas within flood plain, Fig. 17.8).

The juxtaposition of morphologic reaches and profile segments is shown in Figure 17.10A, where it is obvious that profile segments do not exhibit distinctive morphologies. Only the downstream locations of profile segment 2-3 and morphologic reach 1-2 are coincident, although segment and reach 3-4 end in approximately the same location.

Conclusion

The preceding presentation outlines an approach to the reconnaissance of large river systems and an introduction to the type of data that can be extracted from aerial photographs and brief field surveys. The lack of recognizable relationships between surface morphologies and statistically identifiable profile segments indicates the inappropriateness of attempting to use these segments alone as a basis for defining homogeneous river reaches. Thus the long profile approach (at the scale used here) must be used with caution and possibly used only to corroborate river reaches determined on the basis of planform geometry, that is, the profile characteristics of each morphologic unit should be studied individually.

Variations in planform geometry, although distinctive, have not as yet been explained. Such variations could result from either slope changes and/or increases in sediment and water discharge associated with the entrance of tributaries. However, no clear pattern between either or both possibilities and observed changes in morphology appears to exist. This suggests that underlying physical or historical factors may be present. Some information on sediments underlying Big River valley is available. Tucker (1975) briefly described shotpoint data from the flood plain near the confluence of Egg River (Fig. 17.1) just downstream of morphologic reach 3-4 (Fig. 17.2A). The stratigraphic information is too generalized to identify the source of materials near the surface, but Tucker does state that reworked Beaufort sands and gravels exist to about 20 m. These data suggest, at least at this locality, that no structural control on river behaviour exists.

A possible explanation for this complex river morphology is the presence of a sediment wedge composed of Wisconsin age material (M. Church, pers. comm., 1976). The portion of the long profile (Fig. 17.10A) above the mean trend between kilometres 85 and 200 could represent the present position of such a sediment wedge, with the present river degrading, or attempting to degrade, into it. The limits of this wedge appear to be relatively coincident with the relict outwash surface (defined by the paleobraiding pattern).

Comprehensive interpretation of the present behaviour of Big River and other rivers, such as Bernard River, which exhibit similar complexities, requires further investigation into historical aspects. Not only must terrace surfaces be dated and delimited, but the effect of fluctuating sea level and isostatic uplift along the east coast (J-S. Vincent, pers. comm.) also must be defined. Channel pattern variations over time must also be reconciled with changes in climate and sediment supply.

Acknowledgments

The field assistance of R. Gale, S. Miller, and J. Meunier is gratefully acknowledged, as is the photogrammetric work by G. Mizerovsky. Thanks also are extended to M. Church for his suggestion of the Wisconsin sediment wedge, and to P.A. Egginton, C.F.M. Lewis, and W.W. Shilts for their comments.

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A SYSTEM FOR THE OFFLINE PROCESSING OF BOREHOLE GAMMA-RAY SPECTROMETRY DATA ON A NOVA MINICOMPUTER

Project 760048

Q. Bristow
Resource Geophysics and Geochemistry Division

Introduction

This paper is a summary of progress to date on the instrumental and data processing aspects of a project led by P. G. Killeen, concerned with nuclear techniques for borehole logging.

The project has been described in more detail by Killeen and Bristow (1975), Killeen (1976), and by P. G. Killeen and Q. Bristow in a seminar, "A Backpack Portable Borehole Gamma-ray Spectral Logging System with Digital Recording", presented to the 46th annual international meeting of the Society of Economic Geophysicists in Houston, Texas, October 1976.

The hardware and software in this system are designed to accept serial ASCII data from a cassette tape, on which six channels of data are recorded from a portable gamma-ray spectrometry system used for *in situ* borehole measurements.

The six channels of data which are scanned onto the tape in a repetitive cycle at timed intervals are as follows:

CODE	A five digit thumbwheel switch for manual entry of operator codes as required during the log.
Total Count Potassium Count Uranium Count Thorium Count	Four digit numbers corresponding to the count obtained in each channel over the preset time interval.
DEPTH	A six digit number incremented continuously throughout the log (both down and up) by a contact on the winch pulley at 0.1 m intervals.

Hardware

A McPhar Instruments Spectra 44 portable four-channel gamma-ray spectrometer was custom modified to a specification drawn up by Q. Bristow to include a digital recording cassette tape drive which records the data in standard serial ASCII format. A separate playback unit, also supplied by McPhar Instruments, allows the data to be played back via a standard RS232C EIA interface to any suitable device, e.g. a portable data terminal, a telephone modem for data transmission to a remote computer, or directly into a minicomputer via a second standard data terminal EIA interface. This digital recording and playback scheme is now part of the McPhar standard product line. The format of the data is such that 29 digits are scanned onto the tape at the end of each counting time interval followed by linefeed and carriage return characters.

These serve two functions: when data are being played back via an ordinary data terminal (e.g. Texas Instruments Silent 700, DEC Writer, etc.) each data scan is printed on one line with the carriage being returned and the paper advanced automatically before the next; when the playback is into a minicomputer, the linefeed and carriage return characters serve as markers for recognition of the end of scan, over and above the digit counter which is used as well. If for some reason a recording or playback error occurs which causes digits to be added or omitted from the regular cycle of 29, then the proper sequence can be recovered before the next line by using the carriage return character as an overriding condition for ending the current line irrespective of the number of digits which are in it.

The playback unit is AC operated with four front panel selectable baud rates of 110, 150, 300 and 1200.

The minicomputer used in this case is a Data General Corp. NOVA 1220 interfaced to a Tektronix CRT terminal with hard copy unit, and with a five channel digital-to-analogue converter built onto one of the NOVA "general purpose" boards, to allow data presentation on a multichannel strip chart recorder. A second standard TTY board with an EIA RS232C-compatible interface is used to read in the data from the playback unit via the standard 25 pin connector used in most EIA-compatible devices.

Software

A program was designed, written and put into operation which performs the following functions:

- Reads in the data from the cassette tape at 1200 baud into a memory buffer having a capacity of 3000 data records (i.e. scans). The data are sorted into six separate memory blocks as it is read in, corresponding to the six channels of data referred to earlier.
- Allows the data to be smoothed, multiplied by a scaling factor, corrected for Compton scatter and plotted as up to four simultaneous profiles with depth fiducials on a fifth channel.
- Carries out an automatic edit function on the depth data to remove erroneous values due to recording or playback errors, allows manual editing of similar gross errors in the count rate data.

The program was written in assembler language and combined with a sophisticated triple precision fixed

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point math routine which was also written by the author. The combined software occupies 2.5 K of core in the 32 K memory, leaving 29.5 K for the buffer and for four working blocks into which data from any of the blocks in the buffer can be transferred after any one of the processing functions.

Detailed Program Description

Commands are entered by single letters on the keyboard, in response to a "C=" prompt. A summary of the ten command functions with their code letters is given below:

- R READ: Causes cassette data to be read in to the buffer. If the capacity is exceeded or when the End of File mark is seen from the cassette, the number of records, the maximum depth and the depths at which the thumbwheel code was changed are all listed on the terminal.

- L LIST: Causes the contents of the buffer and the four working blocks to be listed along with the record numbers, between depth limits specified by keyboard entry. Used to check data for gross errors which require editing.

- E EDIT: Causes individual data points to be replaced by the average of the adjacent ones by keyboard entry of the block and record number concerned. Used to remove data errors caused by recording or playback problems.

- S SMOOTH: Carries out a three point running average on a block of data specified by keyboard entry and deposits the result in a destination block also specified by keyboard entry. Source and destination can be one and the same block.

- M MULTIPLY: Multiplies all data points in a source data block by a scaling factor and deposits the result in a destination block. The source, destination and scaling factor are all specified by keyboard entry. Used to scale data up or down as required to fit analogue records. Source and destination can be one and the same block.

- C COMPTON CORRECTION: Calls for keyboard entry of the α, β, and γ stripping ratio constants and the destination blocks for the corrected potassium and uranium data. Performs the stripping calculation according to the formulae:

$$U_{CORR} = U - \alpha Th$$

$$K_{CORR} = K - \gamma U - (\beta - \alpha\gamma) TH$$

- F FIND: Searches through block specified between depth limits specified by keyboard entry for highest data point value in the range. Prints this value with the depth of occurrence.

- P PLOT: Plots data from four blocks simultaneously on the strip chart recorder following keyboard entry of the following parameters:

  - plot speed in milliseconds per point
  - the four source blocks for data
  - whether the plot is to be in ascending or descending order of depth data.

  The depth is indicated on a fifth channel of the recorder as a small fiducial at 1 m intervals and a large fiducial at 10 m intervals. The normal or reverse plot capability allows the plot obtained when logging from top to bottom of a hole to be overlain with the plot obtained from bottom to top, for direct comparison of the count rate profiles.

- I INTEGRATE: Sums data values in block specified and between depth limits specified by keyboard entry; prints out total.

- D DIVIDE: Causes point by point ratios to be calculated from data in two blocks specified by keyboard entry with the result being deposited in a destination block also specified by keyboard entry. Due to the wide dynamic range of ratios encountered for low count rate phenomena having a high statistical variance, the ratios are slotted into ten equispaced "channels" between 1 and 225, which correspond to powers of two between 2^{-5} and 2^{5}. The ratio unity is given a value of 127. The values are then suitable for plotting via the five-channel D/A converter which has an 8 bit resolution corresponding to a maximum of 255. The subsequent plot is then a pseudologarithmic one of the point by point ratios in the two blocks specified. The ratio data in the destination block are smoothed automatically following calculation of the ratios as described.

- BLOCK NOMENCLATURE: Single letter keyboard entries address source and destination blocks as follows:

C	CODE contains thumbwheel entries
I	Integral Count
K	Potassium Count
U	Uranium Count
T	Thorium Count
D	Depth
1	Destination block for processed data
2	" " " " "
3	" " " " "
4	" " " " "

```
NUMBER OF RECORDS ON THIS FILE IS:     1283
MAX. DEPTH IS:       62.6
 CODE DEPTH(M)
     6       0.0
  1006       0.2
  2006      31.3
  3006      31.4
  4006      62.6

C=F
 SOURCE BLOCK: I
 BETWEEN & INCLUDING DEPTHS (METERS): 0 31.3
   PEAK     DEPTH(M)
   1500      10.7
C=F
 SOURCE BLOCK: I
 BETWEEN & INCLUDING DEPTHS (METERS): 31.4 62.5
   PEAK     DEPTH(M)
   1806      52.0
C=L
 LIST BUFFER  BETWEEN & INCLUDING DEPTHS (METERS): 10 11.2
   REC  CODE   INT    K    U   TH   ONE  TWO THREE FOUR  DEPTH
   120  1006   309    8    6    4    45    0    80  102   10.0
   121  1006   303    9    6    4    50    0    80  102   10.1
   122  1006   300   10    5    4    65    0    80  102   10.2
   123  1006   351   12    8    4    50    0    80   85   10.3
   124  1006   546   17   11    4    50   23    80  102   10.4
   125  1006   914   30   19    4    45   85   100  119   10.5
   126  1006  1339   50   28    7   110  109   127  144   10.6
   127  1006  1500   63   30    8   170  109   147  136   10.7
   128  1006  1193   52   22    7   170   62   127  111   10.8
   129  1006   730   30   10    4   125   15    80  119   10.9
   130  1006   364   11    4    1    40   15    47  111   11.0
   131  1006   269    7    3    2    40    0    34  111   11.1
   132  1006   228    5    5    2    15    7    40   86   11.2
C=L
 LIST BUFFER  BETWEEN & INCLUDING DEPTHS (METERS): 51.9 52.2
   REC  CODE   INT    K    U   TH   ONE  TWO THREE FOUR  DEPTH
   921  3006   580   20    8    1    50   46    20  186   51.9
   922  3006   718   26   12    1    45   77    34  177   51.9
   923  3006   923   30   14    3    75   62    47  169   51.9
   924  3006  1185   41   17    3   105   85    74  152   51.9
   925  3006  1520   53   23    5   140  101    80  160   52.0
   926  3006  1734   66   29    4   140  163   107  160   52.0
   927  3006  1806   70   32    7   175  140   120  160   52.0
   928  3006  1732   67   36    7   125  171   147  152   52.1
   929  3006  1585   62   33    8   135  132   140  152   52.1
   930  3006  1469   55   31    6    95  148   127  152   52.1
   931  3006  1330   49   27    5    90  132   100  160   52.2
   932  3006  1194   39   28    4    15  155    87  169   52.2
   933  3006   995   30   25    4     0  132    74  177   52.2
   934  3006   812   28   20    3    15  109    67  169   52.2
C=
```

Figure 18.1. A sample of part of the data terminal record.

A typical sequence of operations involving use of the commands, to produce for example smoothed potassium, uranium and thorium profiles corrected for Compton scatter, together with uranium/thorium ratios would be as follows:

- Read in the data cassette via the playback unit.
- Find the peaks in each of the I, K, U, T blocks.
- If any peak data values are suspect, list the buffer between depths bracketing the suspect values. If from this it is clear that a peak value is in fact a gross error, then note the record number and edit it using the "EDIT" command.
- When the "FIND" command turns up peaks which are genuine, smooth the K, U and T blocks depositing the results back into the same blocks.
- Apply the Compton scatter correction factors to the data and deposit the corrected potassium and uranium data into blocks 1 and 2 respectively.
- Take ratios of uranium (corrected) and thorium data (blocks 2 and T) and deposit the results into block 4.
- Find peak values of smoothed and corrected data now in blocks T, 1 and 2.
- Multiply each of the blocks 1, 2 and T by a factor approximately equal to 255/peak value in order to obtain full scale profile plots; deposit the results into blocks 1, 2 and 3 respectively.
- Plot contents of blocks 1, 2, 3 and 4 between the depth limits corresponding to the down-going log as a "normal" plot.
- Plot contents of blocks 1, 2, 3 and 4 between the depth limits corresponding to the up-going log as a "reverse" plot.

A sample of part of the data terminal record obtained by using the program to perform operations on the data is shown in Figure 18.1.

Conclusion

The relatively simple step of adding a cassette recording capability to an instrument which was already an essentially digital device, has opened up a new dimension in terms of the subsequent off-line data processing possibilities which are now available with the aid of a suitable minicomputer program such as the one described above. The use of an ASCII format and the commonly used EIA RS232C interface for the cassette playback, allows use of such a system by virtually any make of minicomputer or large computer centre with conventional telephone modem access. In the case of the system described here, the 32 K memory combined with a very compact software package written entirely in assembler language and tailored to the requirements, allowed a large amount of data to be stored without the necessity of external disc or mag tape storage. Further work in this area will be reported in due course.

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19. COMPUTATION OF *IN SITU* POTASSIUM, URANIUM, AND THORIUM CONCENTRATIONS FROM PORTABLE GAMMA-RAY SPECTROMETER DATA

Project 760045

P. G. Killeen and G. W. Cameron
Resource Geophysics and Geochemistry Division

Quantitative *in situ* determinations of potassium (K), equivalent uranium (eU) * and equivalent thorium (eTh) * concentrations in outcrop can be computed from data obtained with a portable gamma-ray spectrometer when it is properly calibrated. Calibration of a spectrometer includes not only the energy calibration (setting the windows over the proper peaks in the gamma-ray energy spectrum) but also determination of nine factors (stripping ratios, sensitivities and backgrounds) for the instrument. These calibration factors, unique to each portable gamma-ray spectrometer, are computed by the Geological Survey of Canada for users who carry out the calibration procedure on the five concrete calibration pads at Uplands Airport in Ottawa. In order to calculate K, eU, and eTh concentrations from *in situ* measurements, the calibration factors must be utilized in equations of the same form as those from which they were derived. This report shows how the calibration factors are used in determining the K, eU, and eTh concentrations from field data and presents a numerical example of the calculation.

Typical Calibration Factors

A set of nine typical calibration factors for a portable gamma-ray spectrometer with a 76 x 76 mm (3 x 3 in.) NaI(Tl) detector are:

Stripping Ratios: can be defined as number of counts from element A in window B per count from element A in window A

α = 0.62 (counts from Th series in U window per count in Th window)

β = 0.74 (counts from Th series in K window per count in Th window)

γ = 1.12 (counts from U series in K window per count in U window)

Sensitivities: ratio of corrected counts per minute (cpm) to known calibration pad concentration

Ks = 209 cpm/%K
Us = 18.6 cpm/ppmU
Ths = 8.3 cpm/ppmTh

Background: the count rates which would be obtained if the pads had zero K, eU, eTh concentrations

Kb = 53.4 cpm
Ub = 19.9 cpm
Thb = 7.9 cpm

The preceding factors are typical of a portable spectrometer with three windows set to the following energy ranges:

	K	U	Th
Energy window width:	0.200 MeV	0.200 MeV	0.400 MeV
Window centred at:	1.46 MeV	1.76 MeV	2.62 MeV

Calculating an *In Situ* Analysis

Assume the counts measured (in cpm) in the three spectrometer channels (windows) are Kc (potassium), Uc (uranium) and Thc (thorium).

A. To calculate eTh in ppm (parts per million)
   1. To obtain the net thorium count subtract thorium background from the measured thorium count,
      i.e. Th net = Thc − Thb
   2. To obtain eTh ppm divide the net thorium count by the thorium sensitivity (Ths)
      i.e. eTh ppm = (Th net)/Ths

B. To calculate eU in ppm:
   1. Determine the thorium contribution to the uranium count. This is found by multiplying the net thorium count by the factor α.
   2. To obtain the net uranium count subtract the uranium background and the thorium contribution from the measured uranium count
      i.e. U net = Uc − Ub − α x Th net

* Gamma-ray spectrometric determinations of uranium and thorium are based on measurement of gamma-radiation from the decay of Bi^{214} in the $uranium^{238}$ decay series and from Tl^{208} in the $thorium^{232}$ series. The determination is based on the assumption that the daughter nuclide (e.g. Bi^{214}) is in equilibrium with the parent nuclide (U^{238}), that is none of the intermediate steps in the decay series has been disrupted (for example, loss of radon or removal of radium). Consequently what is determined is the amount of uranium or thorium equivalent to what would be in equilibrium with the measured radioactivity of the bismuth or thallium isotope, and the results of these gamma-ray analyses are referred to as "equivalent uranium (eU)" and "equivalent thorium (eTh)". In the case of potassium, the radioactivity of a potassium isotope (K^{40}) is measured directly.

From: Report of Activities, Part A;
Geol. Surv. Can., Paper 77-1A (1977)

3. To obtain eU ppm divide by the uranium sensitivity (Us)

$$eU\ ppm = (U\ net)/Us$$

C. To calculate potassium in per cent:

1. Determine the thorium contribution to the potassium count, i.e. $\beta \times Th\ net$
2. Determine the uranium contribution to the potassium count, i.e. $\gamma \times U\ net$
3. To obtain the net potassium count, subtract the potassium background and the thorium and uranium contributions from the measured potassium count,
$$K\ net = Kc - Kb - \beta \times Th\ net - \gamma \times U\ net,$$
4. To obtain the per cent potassium divide the net potassium count by the potassium sensitivity Ks
$$\%K = (K\ net)/Ks$$

Numerical Example

For a numerical example illustrating the use of the above equations, suppose the count rates reduced to cpm (counts per minute) obtained in the three channels at some location in the field were:

$$Kc = 1195.8\ cpm$$
$$Uc = 463.2\ cpm$$
$$Thc = 422.9\ cpm$$

Then using the previous typical calibration factor the calculations A, B and C below give the *in situ* analyses* for thorium, uranium and potassium respectively:

A. $$\begin{aligned} eTh\ ppm &= Th\ net/Ths \\ eTh\ ppm &= (Thc-Thb)/Ths \\ &= (422.9-7.9)/8.3 \\ &= 415/8.3 \\ &= \underline{50\ ppm} \end{aligned}$$

B. $$\begin{aligned} eU\ ppm &= U\ net/Us \\ eU\ ppm &= (Uc - Ub - \alpha \times Th\ net)/Us \\ &= (463.2 - 19.9 - 0.62 \times 415)/18.6 \\ &= (463.2 - 19.9 - 257.3)/18.6 \\ &= 186/18.6 \\ &= \underline{10\ ppm} \end{aligned}$$

C. $$\begin{aligned} \%K &= K\ net/Ks \\ &= (Kc - Kb - \beta \times Th\ net - \gamma \times U\ net)/Ks \\ &= (1195.8 - 53.4 - 0.74 \times 415 - 1.12 \times 186)/209 \\ &= (1195.8 - 53.4 - 307.1 - 208.3)/209 \\ &= 627/209 \\ &= \underline{3.0\%} \end{aligned}$$

*It must always be remembered that the *in situ* analysis is only valid if the geometry of the field location is similar to that of the calibration test pads, i.e. a relatively flat surface.

METEOROLOGICAL SATELLITES AID AIRBORNE OPERATIONS IN REMOTE REGIONS

S. Washkurak
Resource Geophysics and Geochemistry Division

The Geological Survey in co-operation with PCSP (Polar Continental Shelf Project) have developed a portable meteorological satellite receiver. During the 1976 field season receiver stations were operated at Lynn Lake, Manitoba, and at Resolute Bay, N. W. T.

The station at Lynn Lake provided satellite imagery in support of the airborne gamma-ray spectrometry survey in northern Manitoba, carried out using the GSC Skyvan system. The stretched infrared VISSR (Visible/Infrared Spin Scan Radiometer) data from the geostationary satellite provided a coarse but timely (every half-hour) view of weather conditions affecting the survey area. This data complemented the more detailed imagery received from the Polar orbiting satellite (NOAA). Weather charts of atmospheric pressure patterns, fronts and wind conditions transmitted on the weather facsimile (WEFAX) station from Washington, D. C. were also received using high frequency radio.

At Resolute Bay, only data from the NOAA satellite was received. This data was used by the program manager of PCSP to co-ordinate airborne support of scientific parties working in the Arctic Islands.

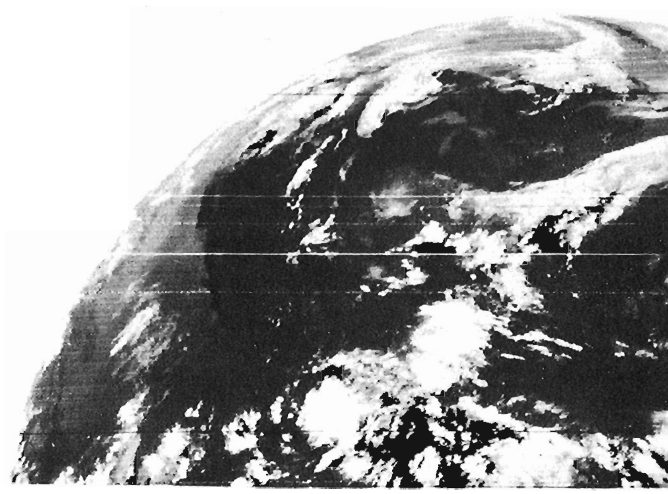

Figure 20. 2. Geostationary satellite image.

The accompanying photographs show the equipment used and illustrate the type of information presently obtainable in the field. Figure 20. 1 depicts the electronic consoles used at Lynn Lake to obtain WEFAX data, and imagery from the geostationary and polar orbiting satellites. Figure 20. 3 shows the receiving antennae at Lynn Lake. Figure 20. 2 is an image received at Ottawa from the geostationary satellite. Note that the image offers good general coverage of most of North America south of latitude 60°N. North of latitude 60°N distortion renders the data less useful. The increase of distortion with latitude means that much of Canada's weather is not observed from the geostationary satellite. Figure 20. 4 obtained at Resolute Bay from one NOAA pass shows the type of coverage available daily over all of Canada. For reference north is as indicated, Greenland is to the right and the Arctic Ocean north of Alaska is on the left. Note the cyclonic disturbance centred at Thule Greenland. Devon Island is clearly visible near the right edge of the centre frame. Much of central Ellesmere Island is cloud covered. To the west of Baffin Island, Victoria Island is visible but Banks Island is cloud covered. From this type of information supply aircraft leaving Resolute Bay had an indication of what type of weather to expect en route to remote survey camps.

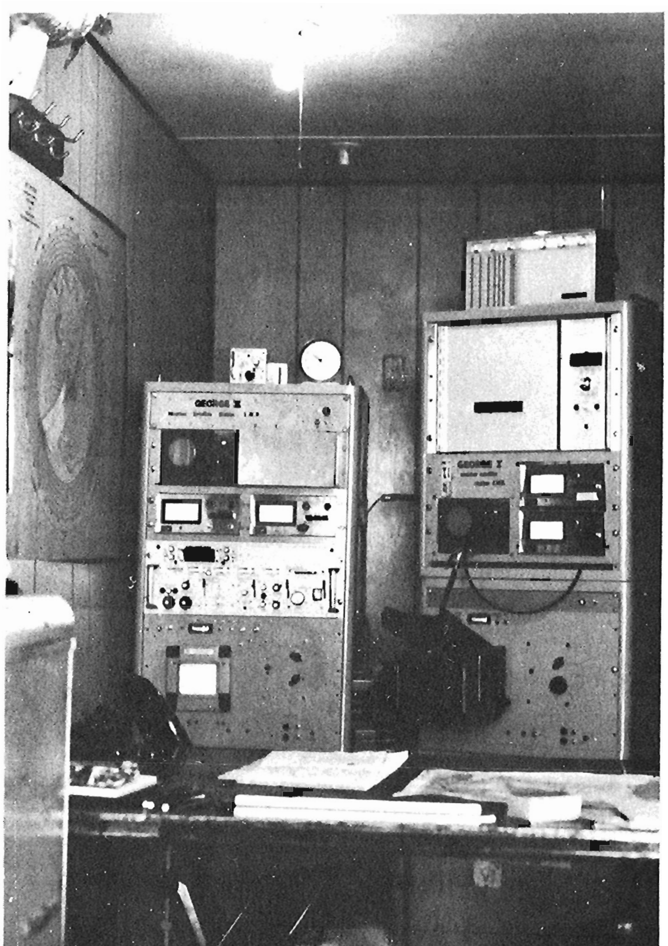

Figure 20. 1. Receiving stations.

From: Report of Activities, Part A;
Geol. Surv. Can., Paper 77-1A (1977)

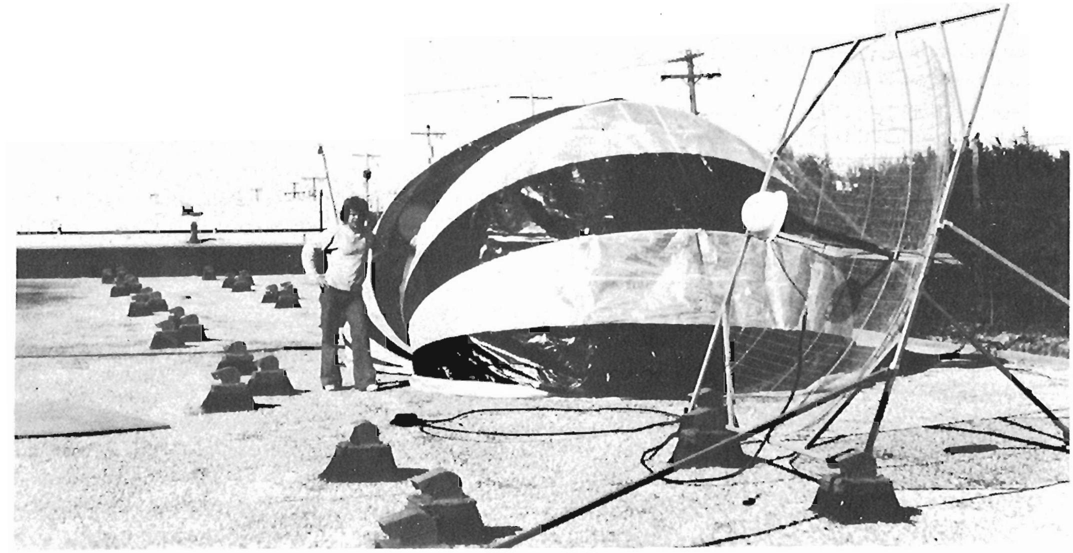

Figure 20. 3. Receiving antennae.

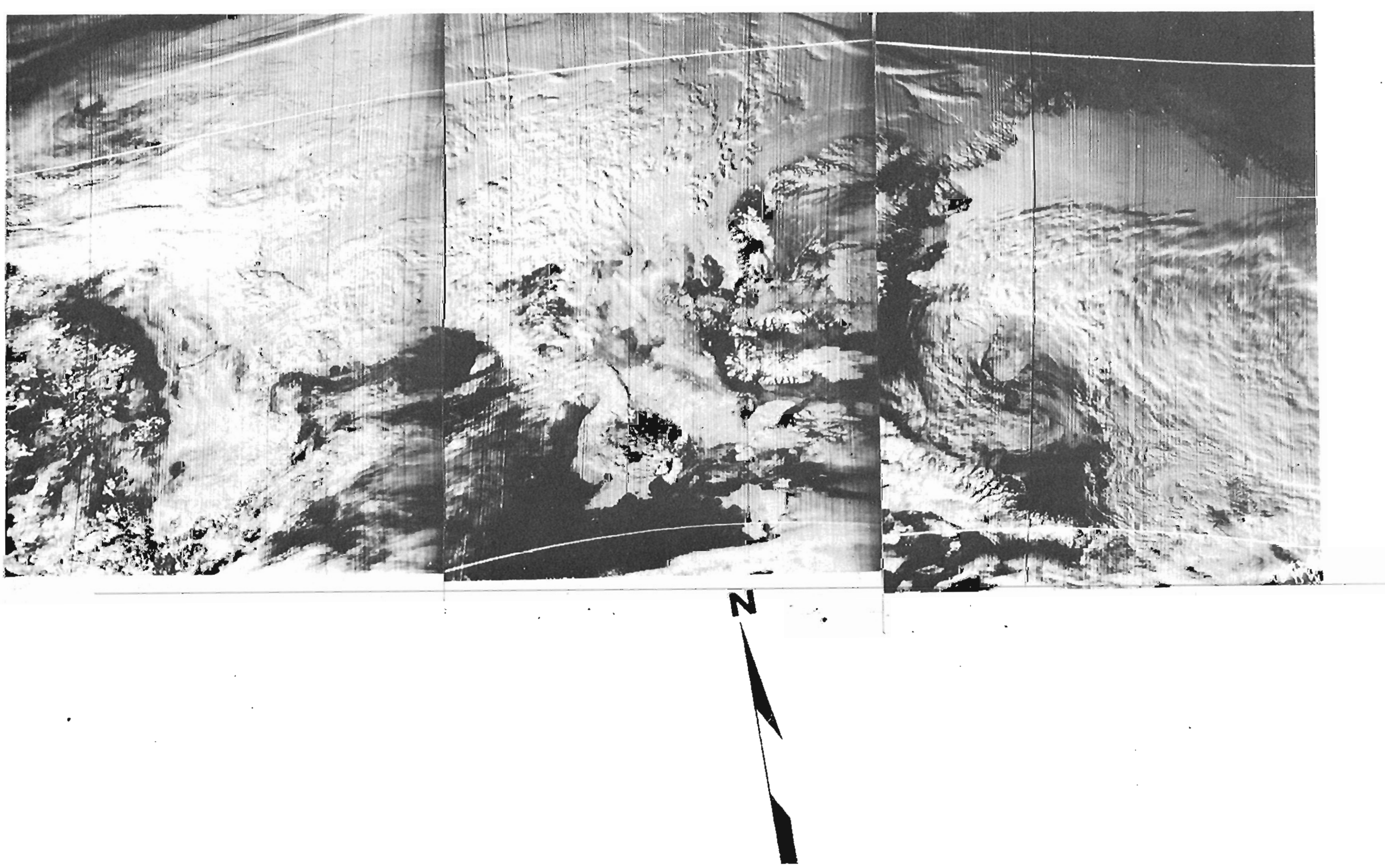

Figure 20. 4.

THE KOOTENAY FORMATION OF ALBERTA AND BRITISH COLUMBIA – A STRATIGRAPHIC SUMMARY

Project 750018

D.W. Gibson
Institute of Sedimentary and Petroleum Geology, Calgary

In 1976, field investigations were concluded on a project begun during the summer of 1975. This project was to describe the stratigraphy and sedimentology of the Jura-Cretaceous Kootenay Formation, in the Foothills and Front Ranges of the southern Canadian Rocky Mountains of Alberta and British Columbia. Thirty-six field sections and six drill cores were measured and sampled in detail in order to provide data on: (1) facies, facies relationships, and correlation of rock units; and (2) the distribution and continuity of coal seams in the Kootenay Formation. In addition, sedimentological data were gathered as an aid in describing and interpreting the history of deposition.

The following summary is a brief description of the Kootenay Formation in Alberta and British Columbia, emphasizing in particular such topics as nomenclature, unusual lithofacies, and the recognition and distribution of major or unusual rock units. A more detailed and comprehensive report on the Kootenay Formation, illustrating in detail all lithofacies, lithofacies variations, coal seam distribution, and depositional environments, will be submitted at a later date.

Kootenay Formation

The predominantly nonmarine Kootenay Formation occupies a narrow, linear band that comprises parts of the Rocky Mountain Foothills and Front Ranges of Alberta and British Columbia. The area is bounded by the 49th Parallel on the south, the Red Deer-Clearwater River area on the north, the Elk River-Kananaskis Valley area on the west, and the Blairmore-Bragg Creek area on the east. The formation is truncated progressively eastward by the Lower Cretaceous Blairmore Group whereas, northward, it becomes increasingly more marine from base to top, and eventually grades laterally into strata of the Nikanassin Formation in the vicinity of North Saskatchewan River.

The Kootenay Formation consists of an interstratified sequence of dark grey to grey-brown weathering siltstone, sandstone, mudstone, shale, conglomerate and coal, ranging in measured thickness from 1128 m (3700 ft.) in the Coal Creek area of the Fernie Basin to 85 m (280 ft.) at Limestone Mountain near Clearwater River in the north. The formation, as defined and recognized herein, conformably but abruptly overlies interbedded sandstone, siltstone and shale of the Jurassic "Passage Beds" of the Fernie Formation. The Kootenay is overlain by sandstone and conglomerate of the Cadomin Formation or by sandstone, siltstone, mudstone, and conglomerate of the Pocaterra Member of the Blairmore Group (Allan and Carr, 1947).

In the Coleman-Blairmore area in Alberta, the Kootenay Formation is subdivided into four members (Norris, 1959) with type section located on Grassy Mountain. However, as a result of the writer's regional investigation of the Kootenay, only the lowermost of these members has proven to be of regional significance. It is suggested that the nomenclature of the other three members be confined to the area east of the Lewis Thrust (for location, *see* Price, 1962), and to the region south of the Blairmore-Coleman area. To the west, in the economically important coal-bearing Fernie Basin, the Kootenay has been divided into three main stratigraphic units (Newmarch, 1953; Jansa, 1972, Fig. 1). The writer tentatively has adopted these subdivisions and the nomenclature of Jansa (1972) with one major modification (Fig. 21.1). Based upon recent results by the writer, it is suggested that the Kootenay Formation should be redefined to include additional older strata. Between the coal-bearing strata of the Kootenay and the dark grey siltstone and shale of the Fernie, a massive, cliff-forming sandstone forms a distinctive lithologic marker, recognizable at all areas of Kootenay exposure in Alberta and British Columbia. Many previous workers including Newmarch (1953), Norris (1959), and Jansa (1972) have placed the contact between the Kootenay and Fernie formations within the cliff-forming marker sandstone facies. As a result, part of the sandstone has been placed in the Fernie Formation, and part into the Kootenay Formation. The writer suggests that the entire cliff-forming sandstone be included as a distinct sedimentological-stratigraphic unit in the Kootenay Formation.

Basal Sandstone Member

The cliff-forming sandstone between the coal-bearing strata of the Kootenay Formation and the interbedded sandstone, siltstone, and shale of the "Passage Beds" of the Fernie Formation is termed tentatively the Basal Sandstone Member of the Kootenay Formation. At most localities in the report-area, the sandstone can be divided readily into two lithologic units, a lower Unit B, and an upper Unit A (Fig. 21.1). The Basal Sandstone Member ranges in thickness from a maximum of 79 m (260 ft.) near Mist Mountain in the Highwood Pass area of Alberta to 16 m (53 ft.) in the Bragg Creek area to the northeast.

Unit B. Unit B (Fig. 21.1) comprises a medium- to thick-bedded, fine- to medium-grained, orange- to yellow-brown weathering quartzose sandstone, ranging in thickness from 52 m (170 ft.) at Weary Creek in the upper Elk River valley to 5 m (15 ft.) at Mount Allan near Banff. The sandstone is argillaceous, slightly limonitic and slightly calcareous and, accordingly, is moderately to poorly indurated. Strata of Unit B have

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Geol. Surv. Can., Paper 77-1A (1977)

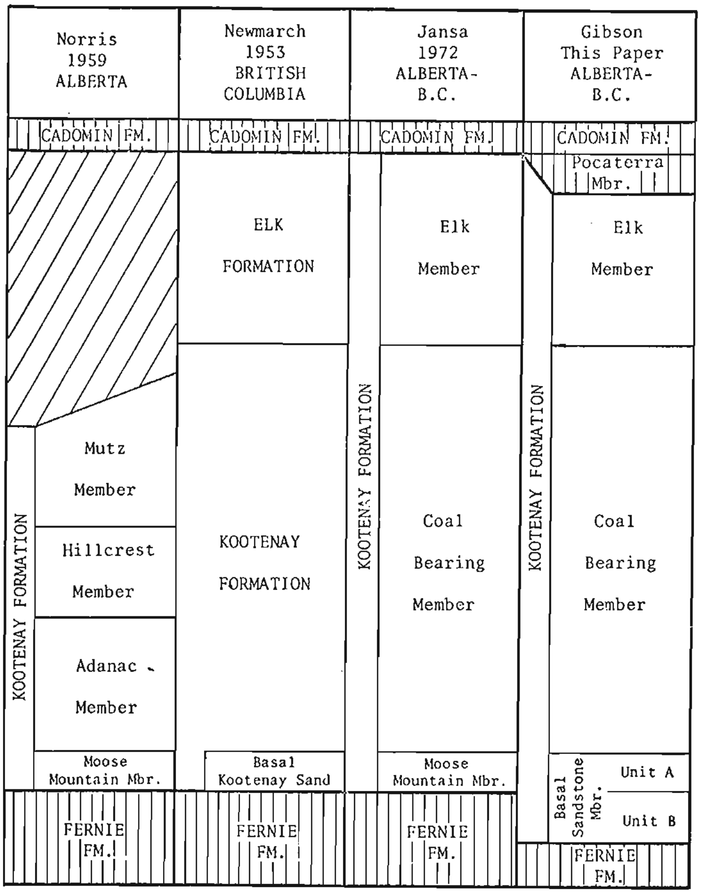

Figure 21. 1. Nomenclature chart illustrating and comparing the main Kootenay and stratigraphically adjacent formations and members recognized in the report-area.

been interpreted by some workers (*op. cit.*) as belonging to the "Passage Beds" of the Fernie Formation; however, as mentioned above, the writer considers the facies to be part of the Kootenay. The contact between Unit B and the underlying strata of the "Passage Beds" is placed at the base of the first continuous occurrence of cliff-forming sandstone devoid of any interbedded siltstone or silty shale.

Unit A. Unit A, commonly referred to as the Moose Mountain Member of the Kootenay Formation (Beach, 1943), consists of fine- to coarse-grained quartzose sandstone, ranging in thickness from a maximum of 35 m (116 ft.) at Mist Mountain near Highwood Pass to a minimum of 4 m (12 ft.) at Adanac strip mine south of Blairmore. The sandstone of Unit A, although compositionally similar to Unit B, is less argillaceous, less ferruginous, more siliceous, more indurated, and more resistant and cliff-forming than the sandstone of Unit B. In addition, strata of the unit weather much lighter grey than strata of Unit B, which serves as an additional criterion in separating the two sandstone units. The contact between Units A and B is not always obvious and must, at some localities, be placed within a gradational interval while, at others, it is considered unrecognizable, with the entire unit along with Unit B classified as undifferentiated Basal Sandstone Member.

Coal-Bearing Member

The Coal-Bearing Member (Jansa, 1972, Fig. 1), a name tentatively used by the writer, comprises a thick succession of grey to greyish brown, interbedded, carbonaceous-argillaceous sandstone, siltstone, mudstone, shale, and economically important thin to thick seams of low to medium volatile bituminous coal. In the Fernie Basin area of British Columbia, conglomerate and conglomeratic sandstone form conspicuous interbeds within the member such that, compositionally, the conglomerate and sandstone resemble those in the overlying Elk Member. Consequently, the occurrence of the conglomerate facies in the Coal-Bearing Member may create difficulty in defining a consistent lithological contact between the two members. Thick bedded, cliff-forming sandstone and conglomeratic sandstone were noted also at the base of the member at some localities on the east side of the Fernie Basin. These strata can be misidentified easily as part of the underlying Basal Sandstone Member if one is not careful. For example, at Mount Taylor, massive sandstone strata, in part conglomeratic, form a conspicuous cliff-forming lithofacies at the base of the member. However, upon close examination these sandstone strata are separated from the underlying sandstone beds of the Basal Sandstone Member by a recessive, commonly covered zone of black carbonaceous shale, with thin lenticular bands of vitreous coal. This unusual sandstone development at the base of the member was noted also at Marten Ridge, and at the headwaters of Leach Creek.

Coal seams form a conspicuous component of the member and, as such, have been actively explored and mined from the late 1800's to the present. In the Fernie Basin area, 14 major seams have been mined to varying degrees while, northward, coal seams of the member increase in number toward the upper Elk River valley, such that, in the Greenhills area, 40 seams of variable thickness have been recognized. However, only 15 of these coal seams appear to be of sufficient thickness and quality to warrant mining under present conditions. Eastward, the Coal-Bearing Member thins, as a result of basin convergence and pre-Blairmore erosion. Consequently, coal seams in the Alberta Foothills and Front Ranges are less numerous and generally more thinly developed.

The contact of the Coal-Bearing Member with the underlying Basal Sandstone Member is sharp and distinct, and is placed at the base of the first occurrence of dark grey carbonaceous siltstone, silty to coaly shale, or coal, overlying the massive cliff-forming sandstone of the Basal Sandstone Member. The contact with the overlying Elk Member generally is abrupt and easily recognized. It is placed at the base of the first major sandstone or conglomerate above the last or uppermost major coal seam in the Coal-Bearing Member. The criteria used to define this contact have been tested at all measured section localities by the writer, and have proved reliable in consistently placing the contact between the two members.

The Coal-Bearing Member ranges from 610 m (2000 ft.) thick at Coal Creek Mountain at Fernie to a minimum measured thickness of 24 m (80 ft.) at Limestone Mountain near Clearwater River, the northern boundary of the study-area.

Elk Member

Elk strata have long been recognized as a distinct and unusual conglomerate lithofacies exposed in the Fernie Basin of British Columbia. For this and other reasons, Newmarch (1953) considered the conglomerate lithofacies as a separate unit of the Kootenay and elevated it to formation status, designating Coal Creek Mountain at Fernie as the type section. The writer, however, has found that these quartz-chert pebble conglomerates do not extend for any distance beyond the Fernie Basin area. For example, in the upper Elk River valley and in most areas of the Alberta Foothills and Front Ranges, the conglomerate has not been observed. The conglomerate grades laterally into coarse grained sandstone and siltstone, or into an alternating or cyclical succession of sandstone, siltstone, and shale as occurs in the Banff-Red Deer River area to the north. The writer, however, suggests that the Elk Member does have regional significance and may warrant elevation to formational status, but for different reasons and characteristics than those suggested by Newmarch (1953). Douglas (1958) divided the Kootenay Formation in the Mount Head area of Alberta into two mappable units, a lower Unit A, and an upper Unit B. The latter unit is interpreted by the writer as homotaxial with the Elk Member of the Fernie Basin.

The Elk Member (Jansa, 1972) consists of a thick, generally cliff-forming succession of interbedded sandstone, siltstone, shale, conglomerate, and, at some localities, sporadic thin seams of coal. The Elk ranges in thickness from 457 m (1500 ft.) at the type section on Coal Creek Mountain to a minimum measured thickness of 28 m (91 ft.) at Cabin Ridge north of Coleman. It thins eastward as a result of pre-Blairmore erosion. The absence or conspicuous reduction in thickness and frequency of coal seams at most section localities is a notable lithologic characteristic of the member. The seams observed rarely exceed 0.5 m (2 ft.) in thickness and are confined usually to the thicker western sections of the report-area. In the upper Elk strata of the Fernie Basin and the Mount Allan area near Banff National Park, an unusual coal type was observed. It consists of thin seams – rarely exceeding 15 cm (6 in.) in thickness, composed of a compacted mass of coniferous? needles. These thin seams commonly have been referred to as "Needle Coal". The occurrence of "Needle Coal" in the Elk Member has proved to be useful in identifying isolated exposures of this unit.

The Elk Member is overlain erosionally by conglomeratic sandstone of the Cadomin Formation of the Blairmore Group. However, at some western sections on the report-area (i.e. the Sparwood, Marten, and Fernie Ridge areas of the Fernie Basin, and the Mount Lipsett, Highwood Pass, and Mount Allan areas of Alberta), the Elk Member is overlain abruptly but possibly conformably by sandstone, conglomerate, mudstone, siltstone and shale of Pocaterra Member of the Blairmore Group (Fig. 21.1). The Pocaterra Member was recognized and described first by Allan and Carr (1947) in the Highwood-Elbow River area of Alberta. This unusual lithofacies below the Cadomin Formation, although sporadic in distribution, has been recognized now as far south as the Fernie area of British Columbia.

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PHANEROZOIC STRATIGRAPHY AND SEDIMENTOLOGY OF SOMERSET ISLAND AND NORTHEASTERN BOOTHIA PENINSULA

Projects 750084, 740030

A.D. Miall and J.Wm. Kerr
Institute of Sedimentary and Petroleum Geology, Calgary

Introduction

Detailed studies of the stratigraphy and sedimentology of the Phanerozoic rocks were begun during the 1975 field season by Reinson *et al.* (1976) as part of Operation Boothia (Kerr, 1976) but were confined mainly to the Silurian to Devonian Cape Storm-Peel Sound sequence of northern Somerset Island. At the end of 1975, Miall replaced Reinson and assumed responsibility for the stratigraphic studies. The work was extended to the entire Phanerozoic succession of Somerset Island and northern Boothia Peninsula and was completed during a 7-week field season in 1976. Detailed ground studies were carried out by Miall, while mapping and correlation work were performed mainly by Kerr.

University of Ottawa graduate students or ex-students S.R. Williams, J. Savelle, M. Gibling and G. Narbonne have contributed unpublished information which the writers gratefully acknowledge.

A summary of the surface stratigraphy is given in Table 22.1.

Lang River Formation

The name Lang River Formation was given by Dixon (1973a, b) to a variegated succession of thin- to thick-bedded dolomite, stromatolitic dolomite, intraformational breccia and conglomerate, and (particularly at the base of the formation) cross-bedded sandstone that rests on the Precambrian basement on both the east and west flanks of the Boothia Uplift. The formation corresponds approximately to map-unit 8 of Blackadar and Christie (1963). Christie (1973) subdivided this map-unit into two new formations using type sections on southern Boothia Peninsula, but he did not attempt to map their distributions and the writers have not found it possible to do so on Somerset Island or on northern Boothia Peninsula.

The Lang River succession is varied, and characterized by marked and largely unpredictable lateral facies changes. However, it is a useful mapping unit which can be distinguished readily from the overlying Irene Bay and Allen Bay formations, and the formation will be retained for mapping purposes.

Partial or complete sections were measured through the Lang River Formation at nine localities along the east flank of Boothia Uplift (Fig. 22.1, Locs. 1, 2, 3, 7, 8, 9, 11, 13, 15). Those near Creswell Bay (Locs. 7, 8, 9) and Hunting River (Loc. 13) are structurally disturbed and a compilation of complete sections will follow the analysis of the regional mapping by Kerr. Some modifications of Dixon's (1973a) detailed work in these areas will be necessary in light of our more complete mapping coverage.

Table 22.1

Phanerozoic stratigraphy of Somerset Island and Northeastern Boothia Peninsula

Age	Formation		Lithology	Thickness (Metres)
Cretaceous-Tertiary	Eureka Sound		Sandstone, siltstone, shale	300
Lower Devonian	Peel Sound	4.	Polymict conglomerate	120
		3.	Pebbly sandstone, conglomerate	240
		2.	Dolomitic conglomerate	280
		1.	Sandstone, siltstone	60-400
Upper Silurian-Lower Devonian	Unnamed	2.	Siltstone, shale, dolomite	150-300
		1.	Dolomite, limestone	0-130
Upper Silurian	Read Bay		Limestone, rubbly, argillaceous	150-240
Lower-Upper Silurian	Cape Storm		Dolomite, thin bedded	120-260
Silurian	Cape Crauford		Dolomite, evaporite	50+
Upper Ordovician-Lower Silurian	Allen Bay		Dolomite, medium- to thick-bedded	340-540
Upper Ordovician	Irene Bay		Limestone	0-97
Upper Cambrian-Upper Ordovician	Lang River		Dolomite, sandstone, shale intraformational breccia	200-420

From: Report of Activities, Part A;
Geol. Surv. Can., Paper 77-1A (1977).

It has not been possible to detect any patterns in the thickness or facies variations except possibly a tendency for a westward thinning toward Boothia Uplift north of Creswell Bay. There are several possible reasons for this variability: (1) A variety of distinct but closely interrelated depositional environments ranging from supratidal to shallow subtidal are probably represented by the Lang River sediments. (2) The Precambrian unconformity surface may have been marked by erosional relief. (3) It is possible that gentle movement took place on the Boothia Uplift during earliest Paleozoic time, causing facies and thickness changes transverse to the Uplift. (4) It has yet to be demonstrated that the Lang River is completely conformable internally. Biostratigraphic evidence from the formation is very sparse and several periods of time are not represented by the available fossils. This may (as suggested by Christie, 1973) or may not indicate the presence of disconformities within the basal Phanerozoic succession.

Some lithologic similarities exist between the Lang River and the three Cambrian to Middle Ordovician units in northwestern Baffin Island described by Trettin (1969). Mayr (in press) correlated the subsurface section in the Garnier O-21 well (Fig. 22.1) with Trettin's units, and it is anticipated that further work on the surface outcrops of western Somerset Island will extend these correlations. By contrast, there is virtually no similarity between the Lang River and the vastly thicker lower Paleozoic succession on Cornwallis Island to the north, which serves to emphasize the very different tectonic settings of Somerset and Cornwallis islands at this time.

Irene Bay Formation

The Irene Bay Formation consists mainly of limestone, commonly bioclastic, and characterized by a suite of large invertebrate fossils known as the Arctic Ordovician fauna. The formation is distributed widely throughout the Franklinian Miogeosyncline (Kerr, 1967) but its recognition on Somerset Island marks the first record of this distinctive unit within the Arctic Platform.

The Irene Bay was studied at four localities, at Lang River (Fig. 22.1, Loc. 3) and three sections near Creswell Bay (Locs. 8, 9, 11). The formation has a maximum thickness of 97 m at Lang River, where it was described, but not identified, by Dixon (1973a, b), but elsewhere is considerably thinner and in many localities is absent altogether. North of Creswell Bay the formation thins westward, perhaps reflecting movement on Boothia Uplift. Other thickness variations may be, in part, the result of pre-Allen Bay erosion.

Allen Bay Formation

Blackadar and Christie (1963) mapped a succession of "light-weathering, relatively competent dolomite, dolomitic sandstone, and minor sandstone containing Ordovician and probably Silurian fossils" which they termed map-unit 9. Christie (1973) assigned these rocks to the new Franklin Strait Formation, designating a type section on western Boothia Peninsula. Dixon (1973a, b) studied the same rocks on Somerset and Prince of Wales islands and used the earlier name Allen Bay Formation (Thorsteinsson and Fortier, 1954). The latter is a well-established formation throughout the Arctic Islands and there is no reason not to use the same name on Somerset Island.

The Allen Bay Formation is recognizable readily by its pale weathering colours and its resistance to erosion. Complete sections are difficult to obtain owing to structural complications near Boothia Uplift, but three were studied during the 1976 field season (Fig. 22.1, Locs. 2, 3, 16). Partial sections were examined at several other localities, particularaly near Creswell Bay (Fig. 22.1, Locs. 8, 9, 10).

As summarized by Dixon (1973b, p. 139), the Allen Bay Formation probably represents a time of stability over Boothia Uplift, and of widespread establishment of tidal carbonate mud flats with the development of stromatolites. Occasional deepening to shallow subtidal conditions is suggested by horizons containing corals, nautiloids and brachiopods.

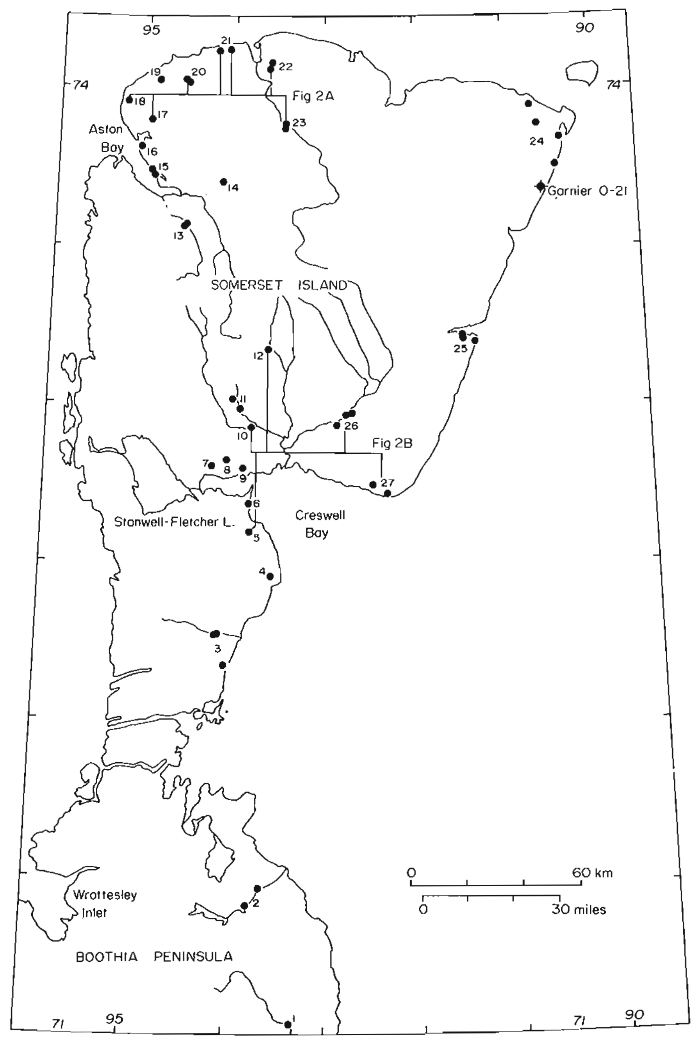

Figure 22.1. Location of the principal measured sections in the Phanerozoic rocks (including those studied by Reinson *et al.*, 1976).

Cape Crauford Formation

The Cape Crauford Formation was established in northwestern Baffin Island (Trettin, 1969) and is equivalent in age to the upper part of the Allen Bay Formation. It has been recognized in the Garnier O-21 well where it is 580 m in thickness (Mayr, in press), and the topmost beds are exposed along the east coast of Somerset Island between Port Leopold (Fig. 22.1, Loc. 24) and Batty Bay (Fig. 22.1, Loc. 25). The formation is distinguished from the Allen Bay by the presence of gypsum and a more varied succession of dolomitic rocks, including intraclast breccia, ripple-laminated stromatolitic and bioturbated units, and bioclastic dolarenite. The overlying Cape Storm Formation is very similar in lithology at Port Leopold except that it contains less gypsum, and the contact is drawn at the uppermost continuous evaporite bed as was done by Jones and Dixon (1975).

The Cape Crauford Formation probably represents a restricted, supratidal sabkha environment in part, and in part an intertidal environment with active tidal channels.

Cape Storm Formation

Blackadar and Christie (1963) recognized one map-unit (unit 11) between the pale weathering dolomites of the Allen Bay Formation and the red-weathering Peel Sound clastic rocks. The unit consists mainly of grey and greenish grey limestone and dolomite, and was correlated with the Read Bay Formation (Thorsteinsson and Fortier, 1954). Field work by S.R. Williams and J. Savelle of the University of Ottawa (unpublished data, 1968 to 1975) demonstrated that the lower part of map-unit 11 constitutes a separate, mappable formation, and Reinson *et al.* (1976) correlated it with the Cape Storm Formation (Kerr, 1975). Jones and Dixon (1975) carried out independent studies on the Silurian rocks of northern and northeastern Somerset Island and established a new unit, the Leopold Formation, which overlies the Cape Crauford Formation and which, on the basis of limited conodont and invertebrate evidence, was interpreted as a lateral equivalent of the Read Bay. Detailed mapping by Reinson *et al.* (1976) suggested that the Cape Storm and the Leopold are the same stratigraphic unit, and this was confirmed during the 1976 field season. Thickness and facies vary between eastern and western Somerset Island and possibly the unit is diachronous, but there is no need to treat these rocks as two formations. For the purposes of the present project the name Cape Storm will be used.

On eastern Somerset Island, the Cape Storm Formation consists mainly of laminated dolomite, sandy dolomite and minor limestone with intraclast breccia, ripple marks and stromatolites, and contains a limited fauna of gastropods, ostracodes and eurypterids. The formation differs from the Cape Crauford only in the absence of continuous evaporite beds. It grades up into the Read Bay Formation, such that the platy-weathering beds of the Cape Storm and the rubbly-weathering Read Bay limestones are interbedded over a vertical interval of up to 30 m. Several sections have been examined near Port Leopold (Loc. 24), the area studied in detail by Jones and Dixon (1975). Other sections were visited between Batty Bay (Loc. 25) and Fury Point (Loc. 27), and this area currently is being investigated in detail by G. Narbonne (University of Ottawa).

On western Somerset Island, partial or complete sections were examined by Kerr and Reinson in 1975 in the general area of Aston Bay (Locs. 14, 18) and by Miall in 1976 to the north and south of Creswell Bay (Locs. 2, 3, 10). Exposures in western Somerset Island contain shale and argillaceous dolomite, imparting a grey to greenish grey colour to the rocks, in contrast to the buff-brown colours of exposures in the eastern part of the island. The Cape Storm also is thinner near Boothia Uplift than farther east. The sedimentological and paleogeographical implications of these differences have yet to be analyzed.

Read Bay Formation

As discussed in the previous section, map-unit 11 of Blackadar and Christie (1963) has been redefined during the present project. The unit was correlated originally with the Read Bay Formation, but laminated dolomites forming the lower part of the unit now are assigned to the Cape Storm Formation, and similar rocks in the upper part of the map-unit are included now in a new, unnamed formation, described below. The Read Bay, as redefined for the purpose of this project, consists of a monotonous succession of rubbly-weathering argillaceous limestone, the weathering character reflecting the pervasive presence of oscillation ripple marks and bioturbation. Oolitic and bioclastic limestone are common. An abundant invertebrate fauna is present, as on Cornwallis Island (Thorsteinsson, 1958). Sections were measured at several localities along the east flank of Boothia Uplift by Reinson (Locs. 18, 23) and Miall (Locs. 2, 5, 6).

The Read Bay Formation represents open-marine, shallow subtidal conditions and is the only formation in the project area, other than the Allen Bay, that does not contain some clear evidence of the influence of Boothia Uplift on thickness or lithology. A general northward thickening is evident, from Aughterston River (Loc. 2) to Pressure Point (Loc. 18) but this probably reflects the northward tilt of the Arctic Platform toward the Franklinian Geosyncline.

Unnamed Formation

On Somerset Island there is a gradual upward transition from the Read Bay limestones into the Peel Sound sandstones and conglomerates. In earlier studies (Thorsteinsson and Tozer, 1963; Blackadar and Christie, 1963; Dineley, 1966; Brown *et al.*, 1969), the clastic units were included in the Peel Sound and the carbonates were assigned to either the Peel Sound or the Read Bay. The distinctiveness of the transition unit was recognized by all these workers, but none studied the island in sufficient detail to justify the introduction of a new formation. This has now been

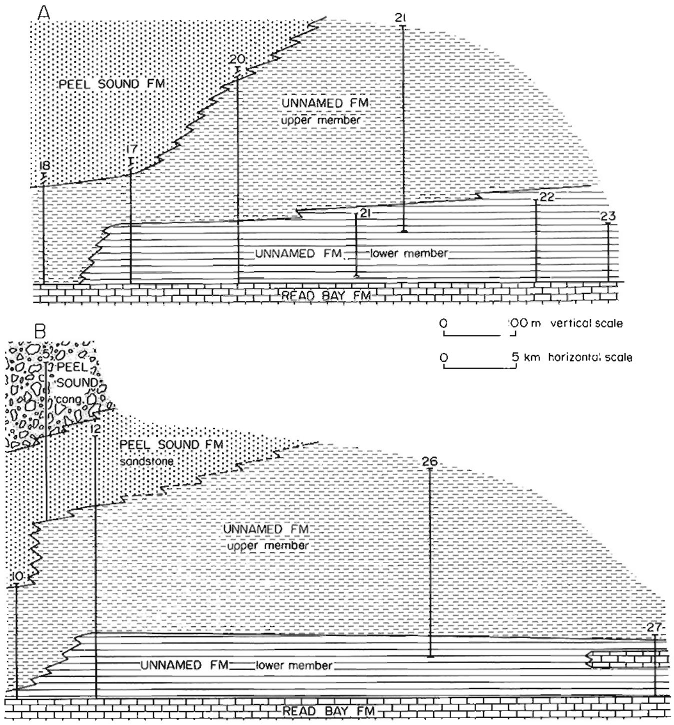

Figure 22.2. Stratigraphic relationships within the Peel Sound Formation and the unnamed formation. Lines of section are shown on Figure 22.1. The top of the Read Bay Formation is used as datum, although it cannot yet be demonstrated that the contact is a time plane.

done and a formal definition will shortly be proposed, in co-operation with M. Gibling (University of Ottawa), who has been working independently in the area since 1973.

The formation contains two members, which show a partial lateral facies transition into each other and into the overlying and underlying units. This relationship was documented by Gibling in 1973-1976 (unpublished data) and by Kerr and Reinson in 1975 (unpublished data) in northern Somerset Island (Locs. 17, 18, 20, 23). Additional information was obtained by Miall in this area (Locs. 21, 22) and the same relationships were established by Miall in southern Somerset Island in 1976 (Locs. 4, 10, 12, 26, 27). At the same time, the facies transitions were traced between the detailed sections by Kerr, using aerial photograph interpretation and helicopter traverses.

The stratigraphic relationships of the unnamed formation are shown in Figure 22.2. The facies changes demonstrate a gradual eastward encroachment of progressively more continental environments into the Read Bay sea. The lower member of the unnamed formation consists of platy weathering, laminated, grey or buff dolomite with a limited fauna of ostracodes, gastropods and rare brachiopods. Vertebrates are abundant in some outcrops. Bioclastic limestone, some beds containing oncolites or stromatoporoids, and stromatolitic dolomite, are present in sections on the east side of Creswell Syncline (Loc. 26). At Fury Point (Loc. 27), the lower member contains several beds of rubbly limestone indistinguishable from those of the Read Bay, suggesting a lateral as well as a vertical gradation into that formation. The same lateral transition into the Read Bay also has been mapped in much of northern Somerset Island and, east of Location 23, rubbly limestone predominates over laminated dolomite. The lower member is interpreted as being predominantly intertidal in origin. Bioclastic limestone units generally have scoured bases, and probably represent tidal channels or storm deposits. They normally grade up into laminated dolomites, suggesting an encroachment of tidal carbonate mud flat conditions.

The upper member of the unnamed formation is characterized by the presence of dolomitic siltstone and mudstone with a purple, red or green colour. The lithologies of the lower member also are present, with the exception of the Read Bay-type subtidal marine limestones. The clastic units contain abundant desiccation features, rare halite casts and gypsum nodules, and are interpreted as the product of fluvial, sheet-flood sedimentation on a supratidal flat, probably under arid or semi-arid conditions. Data from numerous sections have been subjected to Markov chain analysis (method described by Miall, 1973) to test for cyclicity; results are given in Figure 22.3. The succession is dominated by a carbonate-clastic cycle ranging from 2 to 10 m thickness. Laminated dolomite units with flat or scoured bases (Cl) grading up into mottled siltstone (Fl), constitute the commonest cyclic type. The cycles are interpreted as representing a regressive process, intertidal carbonate mud flat conditions being replaced upward by a supratidal alluvial flat environment as fluvial sheet floods caused an eastward progradation of the terrigenous sediments. The cycles commonly are capped by purple silty mudstone (Fm), which probably represents the deposits of ephemeral lagoons or playa lakes. Stromatolite beds (Cs) occur near the top of the dolomite units in some cycles and probably represent a high intertidal environment. Some cycles begin with a thin unit of bioclastic limestone (Cb), which invariably rests on a scoured base. In Figure 23.3, the upward transition from Cb to Cl is shown as occurring with greater than random frequency, but no transition into Cb is indicated. This implies that the occurrence of facies Cb is a random event, probably representing the erosion of the tidal flats or the cutting of a new tidal channel during a storm or exceptional high tide. Once the unusual event passed, however, the evolutionary, regressive sequence of events became re-established. Tidal range may be approximately equal to the thickness of the dolomite units, which averages 2.7 m. Individual cycles may have considerable lateral persistence; one dolomite unit has been traced for more than 20 km near East Creswell River.

As shown in Figure 22.2, the intertidal to supratidal conditions represented by the upper member appeared first in the west and gradually spread eastward over much of the area that is now Somerset Island. The sediment source presumably was Boothia Uplift, as indicated by the nature and direction of the facies changes. The relief of the uplift cannot have been great, however, for even in the westernmost exposures of the formation clastic material rarely is coarser than silt grade.

Peel Sound Formation

The unnamed formation passes laterally into, and is overlain by, red-weathering sandstone and conglomerate of the Peel Sound Formation (Fig. 22.2). In the original definition of the formation (Thorsteinsson and Tozer, 1963), the base was placed at the first red clastic unit, but the writers have redefined the Peel Sound so that the base lies at the first appearance of red sandstone. An underlying succession of interbedded dolomite and red siltstone is reassigned to the unnamed formation (described above). Dineley (1966) and Brown *et al.* (1969) recognized four members within the Peel Sound Formation. The lowermost unit has been redefined to coincide with the redefinition of the Peel Sound Formation, but the remaining members have not been modified by the writers. Their thicknesses and lithologies are given in Table 22.1 and are described below.

The Peel Sound Formation is well exposed in northwestern Somerset Island, and smaller areas of outcrop are present north and west of Creswell Bay. Detailed sections were measured at Localities 5, 12, 19 and 20, and the four members of the formation were mapped in the Cape Anne-Pressure Point area for the first time (Fig. 22.4).

The lowermost member of the Peel Sound (Dp1) is a succession of red sandstone and siltstone containing an abundance of trough and planar cross-bedding, parting lineation and ripple marks. Vertebrate remains are common. A study of the successions using Markov chain analysis reveals a variety of cyclic types, all of which are characterized by an upward decrease in grain size from sandstone to siltstone and an upward decrease in size of sedimentary structures. Virtually all cycles

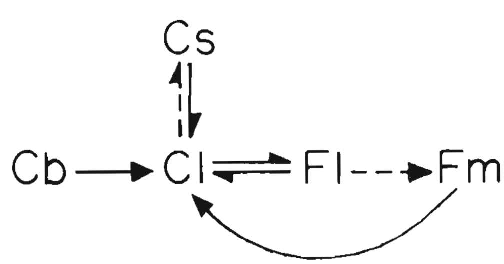

Figure 22.3. Transition path diagram, upper member of the unnamed formation, constructed from the Markov difference matrix (Miall, 1973). Principal transition paths are shown by solid arrows, less common paths by dashed arrows. Facies codes are explained in the text. Only those transitions which occur with a greater than random frequency are shown.

begin with a scoured base. Cycle thicknesses range up to approximately 8 m but 1 to 2 m is typical. The sediments are interpreted as the deposits of a braided river system, each cycle representing either lateral channel migration and bar accretion, or vertical channel aggradation. Most of the sedimentation took place by ripple and dune development in small, shallow channels; cycle thickness probably corresponds closely to channel depth. Paleocurrent determinations in these rocks (165 paleocurrent readings) indicate transport directions away from Boothia Uplift towards the southeast, east or northeast.

Member 2 of the Peel Sound is present only in northern Somerset Island. It is an oligomict conglomerate, consisting of dolmicrite clasts in a matrix of carbonate and quartz sand. Maximum clast size is 23 cm. Brown *et al.* (1969) recorded the presence of fossils derived from the Allen Bay Formation in some of the clasts and this formation, plus the Lang River, probably formed the main sediment source. The conglomerate is thought to represent the deposits of subaerial alluvial fans that prograded eastward from the rising Boothia Uplift. It disappears eastward in an abrupt lateral facies change into members 1 and 3 of the Peel Sound (Fig. 22.4).

Member 3 consists of medium- to very coarse-sandstone, pebbly sandstone and thin conglomerate beds. Clasts are more varied in type than in Member 2, and include Precambrian gneiss, quartzite from the Aston Formation (Proterozoic) and carbonate sediments from the Hunting Formation (Proterozoic) and the lower Paleozoic rocks. Planar and trough cross-beds are common, and occur in sets ranging up to 40 cm thick. In northern Somerset Island, where the member reaches its thickest development (it has not been separately mapped in the Creswell Bay region), paleocurrent determinations indicate northeast to northwest transport directions.

In the vicinity of Locality 19, Member 3 includes several units characterized by very large scale planar cross-beds, ranging up to 6 m in thickness (some of the outcrops of these units were very kindly shown to Miall by M. Gibling). These consist of well-sorted, finely laminated, pebble-free, sandstone, in which grain size ranges up to very coarse sand. Angular grains up to 2 mm in diameter were recorded in a few laminae. The orientation of these structures is different from that of the smaller scale cross-beds in the Peel Sound; transport directions range between north and west, with a mean of northwest. The structures are interpreted as the deposits of large dunes or sand waves that formed in deep fluvial channels, similar to those described in the Brahmaputra River by Coleman (1969). They probably represent the deposits of trunk streams flowing more or less parallel to the structural and depositional strike, as do many large modern rivers such as the Ganges, Indus, and Mackenzie. The interbedded, pebbly sands with smaller scale cross-beds presumably represent the deposits of tributary streams flowing perpendicular to depositional strike. The paleogeographic implications of this interpretation are that the eastern part of the Somerset Island area may have become land by the time Member 3 was deposited, and that somewhere north of Somerset

Island, where north-flowing trunk streams must have debouched into the sea, a large delta system may have developed. No traces of such a delta have yet been identified, although a unit such as the Bathurst Island Formation (Kerr and Christie, 1965; Kerr, 1974) is a possible candidate. Subsurface work by U. Mayr (pers. comm., 1976) shows that the formation is unusually thick (> 2000 m) in southeastern Bathurst Island, and that it could represent a distal delta facies.

Member 4 of the Peel Sound Formation is present in northern Somerset Island (Fig. 22.4) and to the west of Creswell Bay (Loc. 5). It consists of polymict conglomerate (clast types as in Member 3) and thin beds of red sandstone. Clasts reach 30 cm in diameter. The mode of origin of these rocks probably was very similar to that of Member 2, but continued erosion on the Boothia Uplift caused the stripping away of the Phanerozoic cover and the derivation of abundant detritus from the crystalline Precambrian basement. Several small outliers of the Peel Sound Formation, a few miles to the west of Stanwell Fletcher Lake, probably date from this period. The rocks of the outliers consist of immature, poorly sorted, purple and red, pebbly and silty sandstone and siltstone resting in hollows within the rugged metamorphic terrain. Sparse cross-bedding evidence indicates

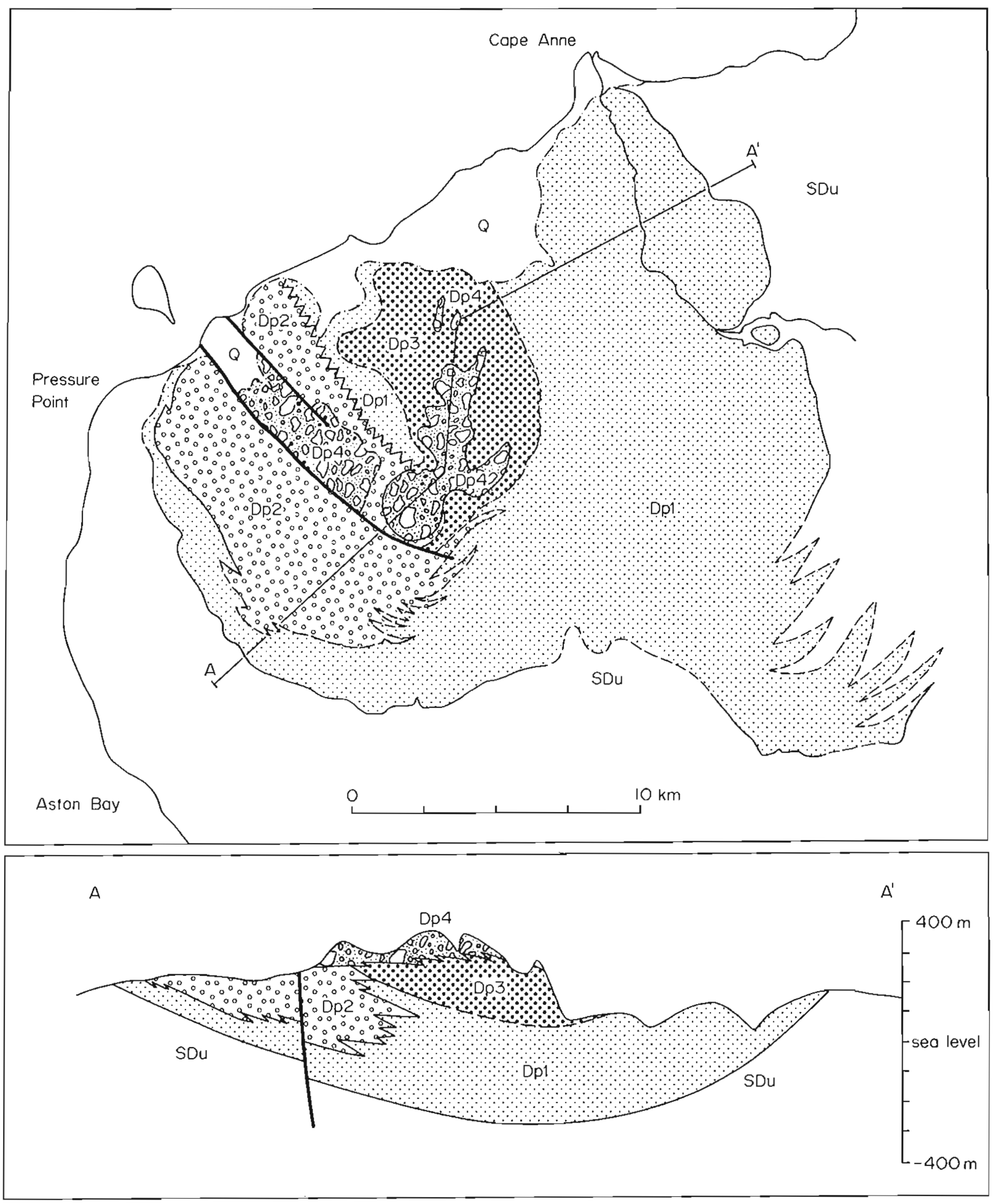

Figure 22.4.

Map and cross-section of the Cape Anne-Syncline area of northwest Somerset Island, showing facies relationships within the Peel Sound Formation. The members are numbered as in Table 22.1.

Dp: Peel Sound Formation
SDu: unnamed formation
Q: Quaternary

northward transport directions corresponding, no doubt, only to the local variation in direction of one of the eastward-flowing rivers that emptied onto the Somerset Island alluvial plain.

The base of Member 4 cuts down into younger rocks of the Peel Sound toward the west and, between Localities 17 and 19, rests directly on Member 2 (Fig. 22.4). The cause of this minor unconformity probably was uplift taking place to the west during deposition of Members 3 and 4 to the east. A similar type of syndepositional structure was described by Miall (1970) in the Peel Sound of Prince of Wales Island, and such structures appear to be common in areas of coarse clastic sedimentation adjacent to rapidly rising areas (Bryhni and Skjerlie, 1975; Riba, 1976).

Other aspects of Peel Sound sedimentation are currently under study by M. Gibling (University of Ottawa), including the depositional processes and clast fabric of the conglomerates and, in co-operation with D. Elliott (University of Bristol), the ecology of the vertebrates, which reach their greatest abundance in Member 1 of the formation.

Cretaceous-Tertiary Sediments

Numerous fault-bounded outliers of Cretaceous-Tertiary sediments are present in Somerset Island. Most are small in areal extent, such as those near Cunningham Inlet (Hopkins, 1971) and north of Creswell Bay (Dixon *et al.*, 1973). Larger outliers are present north of Stanwell Fletcher Lake (Dineley and Rust, 1968) and east of Wrottesley Inlet (discovered in 1976). The sediments near Creswell Bay correspond in age and lithology to the marine Kanguk Formation (Upper Cretaceous), but most are largely nonmarine and are assigned to the Eureka Sound Formation.

The rocks occupying the graben north of Stanwell Fletcher Lake consist of a succession of sandstone, siltstone, shale and conglomerate, named the Idlorak Formation by Dineley and Rust (1968). The formation contains marine fossils at the base, but passes up into a thick succession of cross-bedded sandstone and minor conglomerate containing only derived fossil material. Dineley and Rust (1968) interpreted the Idlorak as deltaic in origin. An examination of the sandstone by Miall suggested a close analogy with certain modern braided river environments dominated by dunes and linguoid or transverse bars. The complete succession, therefore, represents a regressive sequence.

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23. STRUCTURAL GEOLOGY OF SOMERSET ISLAND AND BOOTHIA PENINSULA, DISTRICT OF FRANKLIN

Project 740030

J. Wm. Kerr and C. D. S. deVries[1]
Institute of Sedimentary and Petroleum Geology

Introduction

The geology of Somerset Island and Boothia Peninsula (Fig. 23.1) is dominated by the Boothia Uplift (Kerr and Christie, 1965), a north-trending positive basement controlled tectonic feature. The Boothia Uplift is redefined by Kerr (in press). In the report-area, it includes a core of crystalline basement rocks called the Boothia Horst, and the overlying and adjacent sedimentary superstructure of folded Paleozoic rocks called the Cornwallis Fold Belt. The horst is part of the Churchill Province of the Canadian Shield. Positive and negative movements of the horst produced the structures of the fold belt. Extensive normal faulting that affected the area after activity of the uplift is related to continental drift.

A prime objective of the present study is to locate faults on Somerset Island and Boothia Peninsula to assist in choosing a route for a proposed natural gas pipeline. The work on the crystalline rocks was done by de Vries, and on the sedimentary rocks by Kerr. The present paper reports the result of 1976 fieldwork, and expands upon an earlier paper (Kerr and de Vries, 1976).

The Crystalline Precambrian Shield

The entire area of the Canadian Shield exposed on Somerset Island has now been mapped, with study being concentrated in the southern part where exposure is good. In southern Somerset Island, the various phases of folding can be distinguished clearly by mapping relatively well exposed calcsilicate and metabasite markers. The entire Shield area of Somerset Island was affected by late, regionally pervasive north-south trending folds. In the south these folds are relatively open, and permit the recognition of earlier fold phases, as well as the interference patterns resulting from late phase overprinting of earlier folds. By contrast, in northern Somerset Island the late folding structures are tight, and tend to obscure evidence of earlier structures.

The various fold phases have a slightly different designation in this report than that used in an earlier report (Kerr and de Vries, 1976). In this report, F_1 refers to the rare, oldest set of tight or isoclinal folds. The set referred to as F_2 is newly recognized; these folds are locally developed, open folds that trend approximately east-west. In this report, F_3 refers to the regionally pervasive structures that were termed F_2 earlier (Kerr and de Vries (1976). The elongate domal and basinal structures in the Bellot Strait – Fitz Roy Inlet region are the result of interference between east-west trending F_2 structures and north-south trending F_3 folds.

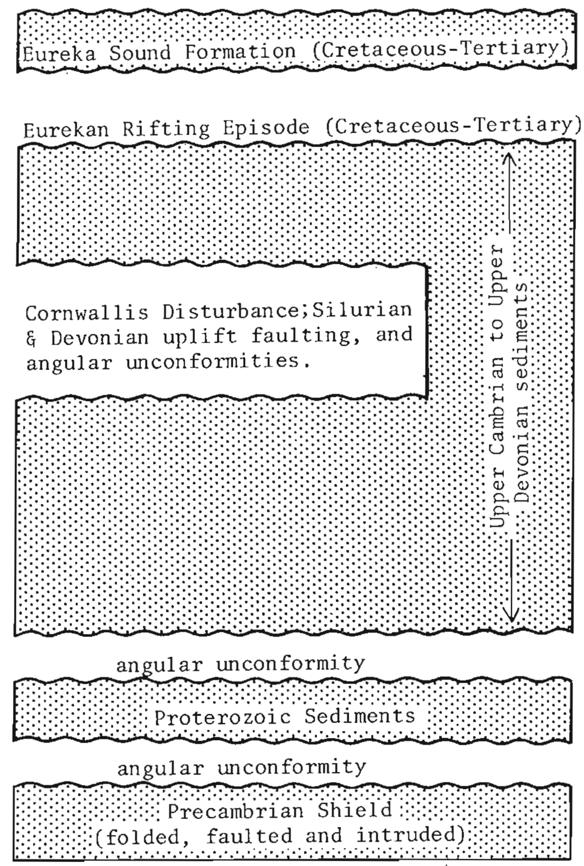

Figure 23.1. The main divisions in the stratigraphic column and the tectonic events of Somerset Island and Boothia Peninsula.

Mesoscopic features in isolated outcrops attest to the complex sedimentary, structural, metamorphic and plutonic history of the crystalline rocks now forming the core of the Boothia Uplift. The combined intensity of F_1 and F_3 deformational events, however, has tended to transpose initial angular discordances between rock units by extensive flowage parallel to the axial surfaces of the structures.

Two lithological units in the Shield that are of sedimentary origin and can be traced for considerable distances around megascopic structures, are the calc-silicates and marbles, and the garnet sillimanite gneisses. These units have been thinned tectonically so that they

[1]Department of Geology, University of Calgary, Calgary, Alberta

From: Report of Activities, Part A;
Geol. Surv. Can., Paper 77-1A (1977).

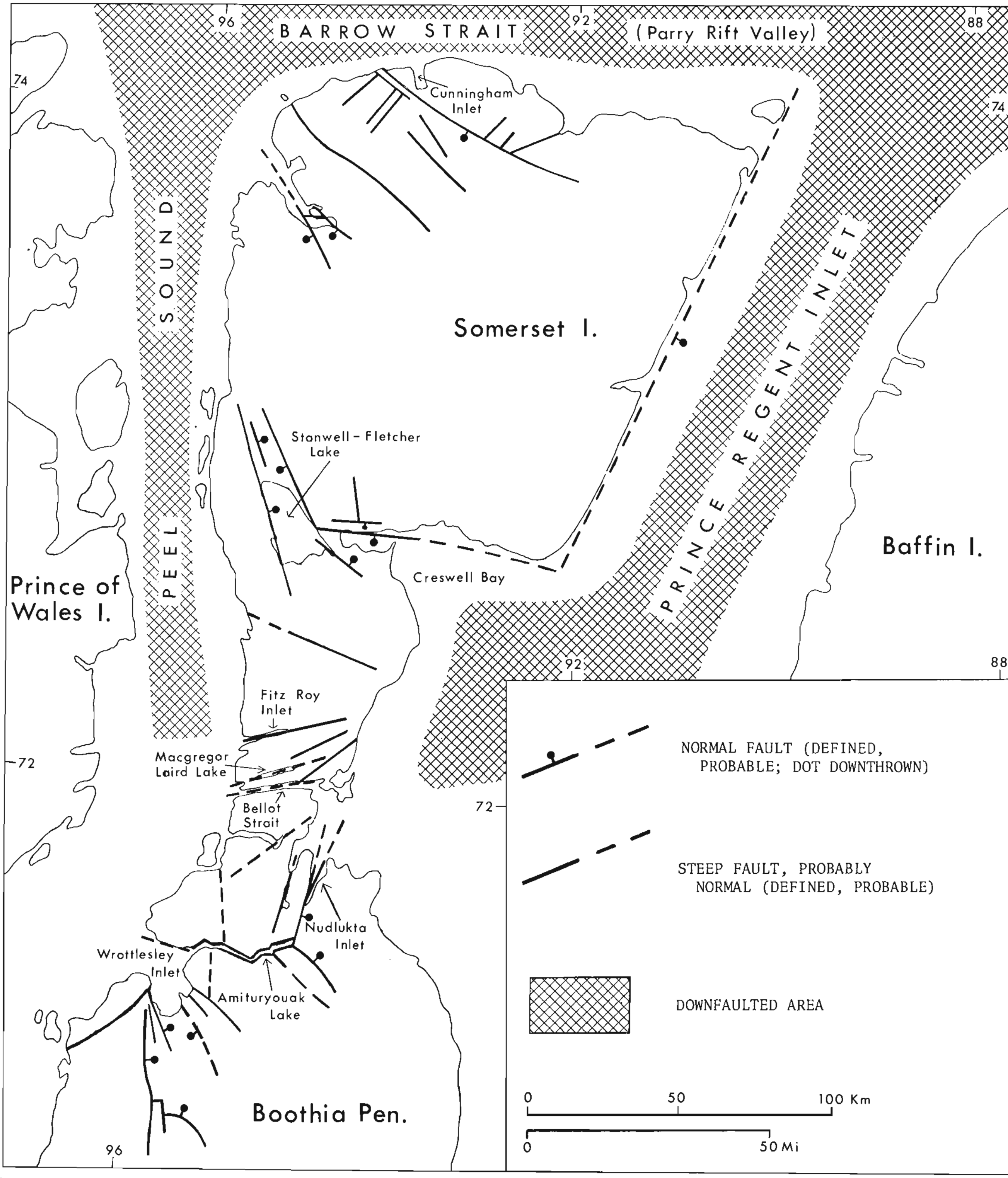

Figure 23. 2. Structures formed on Somerset Island and Boothia Peninsula during the Eurekan Rifting Episode.

commonly disappear altogether along strike. They, nevertheless, have been traced over remarkably long distances; in places they are not thicker than 20 cm. Calcsilicate bands now have been mapped over the entire length of the exposed Shield of Somerset Island.

Four lithological units probably can be regarded as pre-F_3 intrusives into the layered gneisses, although discordant relationships can rarely be demonstrated in the field.

1) A thick, poorly foliated, monotonous, brown-weathering quartz-feldspar gneiss, in which biotite and hornblende sometimes occur, is present at the coast west of Macgregor Laird Lake, and along the limbs of F_3 structures farther east. The relatively abrupt weakening or disappearance of the quartz foliation in these rocks near the contact with well foliated gneisses, and the monotonous character of the mineralogy of this unit suggest that it may have originated as an acid intrusive. The unit can be traced around several F_3 synforms and antiforms, but appears to be concordant with the surrounding foliated gneisses.

2) Numerous metabasites, that range from 50-cm-thick agmatitic bands to 100-m-thick metabasite sheets, are present over the entire map-area. They commonly display a pronounced mineralogical change from a coarse hornblendite core to a finer grained pyroxene and plagioclase-rich border zone. At two localities the metabasites have discordant relationships with the surrounding gneisses, suggesting that the metabasites originated as intrusive sheets.

At least some of the metabasites were emplaced prior to the F_1 event since, at one locality, a thick metabasite occupies the core of an F_1 antiform refolded by an F_3 synform.

3) Coarse pegmatitic bodies, both discordant and concordant, are present throughout the entire map-area, although they are not abundant.

4) Thin, conformable lenses of serpentinized ultramafic rocks have been found in the Bellot Strait – Fitz Roy Inlet region. A sheet-like network of serpentine surrounding remnants of coarse orthopyroxene is oriented approximately parallel to S_3 surfaces, suggesting that the ultramafic unit was emplaced before F_3 deformation ceased.

Migmatization has affected all pre-F_3 rock units with the possible exception of the ultramafics. The amount of leucosome varies greatly with the rock type. Garnet-sillimanite gneisses in some cases, for example, contain up to 60 per cent quartzofeldspathic leucosome that is distributed as thin, irregular layers in the gneiss. Thin metabasites commonly display agmatitic textures, while the leucosome in the thicker metabasites commonly is concentrated in discrete veins.

Evidence from isolated outcrops suggests that probably more than one episode of migmatization affected the rocks since F_1 deformation took place. Early layered rock units, including calcsilicates and sillimanite-garnet gneisses are involved in tight to isoclinal F_1 folds. These mesocopic structures are cut discordantly by fine- to medium-grained quartzofeldspathic mobilizates, which also occur as concordant layers parallel to the overall lithological layering. This unit, in turn, is cut by a set of coarse grained pegmatitic sills and dykes that are oriented approximately parallel to S_3 surfaces.

The Cape Granite granite is a mesozonal pluton (Kerr and de Vries, 1976) that occurs at Cape Granite on northwestern Somerset Island and postdates F_3 deformation. Late Proterozoic diabase dykes, sills and plugs, that occur over the entire exposed Shield, are the youngest intrusive events to affect the Shield of Somerset Island.

Folds in the Crystalline Shield

All three phases of folding recognized to date, F_1, F_2, and F_3, are represented by large, macroscopic structures in the southern part of Somerset Island. The F_3 structures, however, are the most important regionally, and determine the predominantly north-south trending structural grain of the Precambrian Shield of Somerset Island.

The F_1 folds are tight to isoclinal structures that probably occur throughout Somerset Island, but can be positively identified only in hinge regions of broad, open F_3 folds.

A well-developed quartz foliation, defined by platy quartz, is present in all quartz-rich rocks. Since the quartz foliation is oriented parallel to axial surfaces of F_3 folds, and occasionally can be seen to cut across limbs of F_1 folds, it was thought to have been formed during F_3 deformation (Brown *et al.*, 1969; Kerr and de Vries, 1967). In some hinge areas of open F_3 folds, however, two sets of quartz foliation are present, one parallel to the axial plane of F_3 folds, the other parallel to gently dipping axial surfaces of F_1 isoclines. It seems that some of the platy quartz that developed during F_1 survived reorientation during penetrative F_3 deformation.

A prominent quartz lineation, at the intersection of the platy quartz foliation and the lithological layering, is present at many localities on the Shield, and often can be demonstrated to be parallel to F_3 fold hinges. In the area west of Stanwell-Fletcher Lake, the quartz lineations consistently plunge gently southward, while F_3 hinges plunge northward, again suggesting that some of the quartz foliation formed prior to F_3 deformation.

The complex fold pattern that was produced by interference between F_1 and F_3 perhaps is exemplified best by a large-scale, doubly plunging, "banana-shaped" structure north of Macgregor Laird Lake. The F_1 axial plane can be traced for a distance of 3 km around several F_3 hinges. The F_1 axial plane is parallel approximately to the lithological contacts and the tight F_1 hinges can be recognized only by the presence of calcsilicate markers. The F_1 fold has been refolded by a number of north-trending F_3 folds with steep axial planes and variably plunging axes.

The original orientation of F_1 structures is unknown, and the attitude of only a few F_1 hinges could be determined. Field evidence, however, seems to suggest that F_1 folds are present over an extensive area, and tight F_1 folds perhaps may have dominated the structure of the map-area before the advent of F_3 deformation.

The F_2 structures are east-west trending, open folds with steep axial planes that can be recognized by their characteristic domal or basinal interference pattern with F_3 folds. No mesoscopic folds are associated with F_2 structures. The shape of the domes and basins is determined largely by the tightness of F_3 folds. Two oval-shaped basins outcrop west of Macgregor Laird Lake. A good example of an elongated dome is the structure north of Fitz Roy Inlet along Peel Sound, while a nearly circular dome, outlined by a calcsilicate band, occurs 40 km to the north. The F_2 structures, probably originated as gentle, east-west oriented warps in F_1 isoclines.

The F_3 structures, which are dominant and trend north-south, display almost the entire spectrum of fold styles from open concentric to tight similar folds. Open F_3 folds are more abundant in the southern part of Somerset Island than farther north.

Axial planes of F_3 folds invariably are steep to vertical and, in mesoscopic folds, platy quartz commonly marks the axial trace. The profiles of many major F_3 folds show that parasitic folds generally are restricted to a narrow hinge zone and are absent from the limbs.

Fold axes of F_3 folds generally plunge between 10 and 30 degrees northward, but plunge reversals occur in the domal and basinal F_2/F_3 interference structures. An extremely variable pattern of F_3 plunges was observed in the large-scale "banana-shaped" F_1/F_3 interference structure north of Macgregor Laird Lake referred to previously. This pattern is attributed to a combination of two factors: the difference in attitude of beds separated by the F_1 axial plane at the onset of F_3 folding; and the ductile behavior of marbles and calcsilicates in the cores of folds.

Faults in the Crystalline Shield

Two prominant topographic lineaments in the southern part of Somerset Island are east-northeast trending faults with an apparent right-lateral strike-slip displacement in the crystalline rocks. At Macgregor Laird Lake, correlation of lithological markers and traces of axial planes of folds across the lake indicate right-lateral displacement within the Shield. The displacement decreases eastward, being 2.5 km near the west end, 2.2 km in the middle, and 0.3 km east of the lake. The fault in Fitz Roy Inlet had approximately 2 km of right-lateral displacement within the Shield.

In both of the strike-slip faults mentioned above, nearly all of the displacement cited occurred prior to the Paleozoic column, whose base is Upper Cambrian (Miall and Kerr, this publ., rep. 22). On their east ends, however, the two faults offset the Paleozoic rocks with a small amount of mainly vertical displacement. This vertical displacement is considered to be a rejuvenation in Cretaceous or Tertiary time (*see* below).

Bellot Strait appears to occupy a shear zone parallel to its trend, that developed, presumably, in Precambrian time. No estimate of the sense or amount of displacement can be given because markers were not traced to the south. A minimum age for the Precambrian faulting along the three east-northeast trending structures in the Shield of southern Somerset Island described above would be set by the shearing and veining of a diabase plug near the eastern entrance to Bellot Strait.

Sedimentary Cover Rocks

The sedimentary cover rocks of Somerset Island and Boothia Peninsula lie unconformably upon the crystalline Shield and range in age from Late Cambrian to Late Cretaceous. They have been summarized by Reinson *et al.* (1976), Kerr (in press), and by Miall and Kerr (this publ., rep. 22).

The main divisions in the stratigraphic column and tectonic events that occurred on Somerset Island and Boothia Peninsula are shown in Figure 23.1. The times in which rock was being deposited are shown by pattern, and the times of tectonic disturbance are shown by the intervening blank spaces.

All structures that now exist in the sedimentary cover rocks of Somerset Island and Boothia Peninsula were controlled in a large way by the prominent gneissic trends in the underlying Precambrian crystalline basement. Nearly all faults that cut the sedimentary cover rocks follow older faults in the Shield that were reactivated.

Faults of the Cornwallis Disturbance

The Cornwallis Disturbance of Early Silurian to Late Devonian time affected much of Boothia Peninsula, and western Somerset Island. The structures that formed during that disturbance are described by Kerr (in press). In this area, they are dominantly steep to vertical faults and flexures, mainly north-trending. The forces causing that disturbance originated in or beneath the crystalline basement and extended upward into the sedimentary cover rocks. Between the Cornwallis Disturbance and the Eurekan Rifting Episode, Somerset Island and Boothia Peninsula apparently were not deformed structurally (Kerr, in press).

Faults of the Eurekan Rifting Episode

The Eurekan Rifting Episode is a tectonic event defined by Kerr (in press), in which the central Canadian Arctic was subjected to extension and major block faulting. Rift valleys and their branches were down dropped to form linear marine channels. A most striking downfaulted feature is Parry Submarine Rift Valley (Kerr, 1967, in press), which trends east-west and lies just north of Somerset Island (Fig. 23.2). In the central Canadian Arctic, the Eurekan Rifting Episode began in Late Cretaceous (Maestrichtian) time. There probably was additional faulting within Tertiary time, and this may continue to the present.

All faults that are shown on the map of Somerset Island and Boothia Peninsula (Fig. 23.2) are considered to have been active during the Eurekan Rifting Episode.

Older faults that were not reactivated during the episode are not shown. The Eurekan Rifting Episode preceded and coincided with deposition of the Eureka Sound Formation. Many of the faults of the Eurekan Rifting Episode (Figure 23.2) displace the Cretaceous Eureka Sound Formation. This includes the main fault south of Cunningham Inlet, where the Eureka Sound Formation was reported by Hopkins (1971), the graben at Stanwell-Fletcher Lake (Dineley and Rust, 1968), the short east-trending fault north of Creswell Bay (Dixon *et al.*, 1973), and a major fault at Wrottlesley Inlet. Other faults displace the Shield and the Paleozoic column only, but are attributed to the episode because of their configuration, extending into the present day marine channels. Nearly all faults of the Eurekan Rifting Episode appear to have rejuvenated older structures, or at least have been guided by the older structures in the Shield and Paleozoic cover.

Several major faults follow coastlines, with the coastal side downdropped relatively. The east-west fault bordering Creswell Bay is well documented on the west. It had vertical displacement only, with the south side down. It has been traced as far east as 93°40'W, but probably continues eastward. This fault may connect with another long, straight fault that trends along the east coast of Somerset Island. That eastern coastline, which probably is the straightest coastline in the Canadian Arctic Islands, is presumed to be faulted, mainly because of its straightness.

The Eureka South Formation is downfaulted against the crystalline shield on the west side of Wrottlesley Inlet on western Boothia Peninsula (Fig. 23.2). South of the inlet is a large low-lying valley, that is presumed to be large graben. This graben appears to connect with a large downfaulted area in Peel Sound.

The faults at Nudlukta Inlet cut Paleozoic rocks and appear to have been active during the Cornwallis Disturbance. Their configuration, extending northward to a channel, suggests reactivation in the Eurekan Episode. Amituryouak Lake contains downfaulted Paleozoic rocks at one location. It appears to have connected faults on either side of Boothia Peninsula during the Eurekan Rifting Episode. The main northwest-trending faults of northern Somerset Island have displacements that increase toward the north. They are considered to be southward projections from Parry Submarine Rift Valley (Kerr, in press).

The Eurekan Rifting Episode resulted from continental drift. The faults of that episode are westward extensions into the central Canadian Arctic from older and more substantial spreading centres farther east, Baffin Bay and Labrador Sea (Kerr, 1967). The stress configuration that produced the Eurekan Rifting Episode, that is the tendency for rotation apart of continental blocks, may still exist today. Thus, the faults of the Eurekan Rifting Episode (Fig. 23.2) are potential modern dry fault zones. These faults should be considered in pipeline or other construction projects on Somerset Island and Boothia Peninsula.

There is a concentration of earthquake epicentres in the area of Barrow Strait, with a more diffuse trend through Somerset Island, and down Boothia Peninsula (Basham *et al.*, in press). According to D.A. Forsyth (pers. comm., 1976), well over 90 per cent of the earthquakes north of latitude 60 degrees in Canada are of magnitude <5, and this may indicate that stresses are not large enough to effectively "creak" existing structures. We concur and suggest that current seismicity, which appears to coincide with the regions deformed during the Eurekan Rifting Episode, may be very mildly augmenting the deformation of the Eurekan episode.

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TWO STAGES OF FAULTING, VIRGINIA FALLS MAP-AREA, DISTRICT OF MACKENZIE

Project 750055

Donald G. Cook
Institute of Sedimentary and Petroleum Geology, Calgary

Introduction

Virginia Falls map-area (NTS 95F) (Douglas and Norris, 1960) displays a prominent set of north-northeast trending faults, and a less prominent but significant set of northwest-trending faults. Relationships between the two sets can be determined in a relatively small area (Fig. 24.1) in the east-central part of Virginia Falls map-area. There, the north-northeast trending faults are steeply dipping with moderate reverse displacements. They may have originated as right-lateral wrench faults which subsequently were rejuvenated as reverse faults. The north-northeast trending faults are overridden by and therefore predate the Tundra Thrust, a northwest-trending thrust fault probably having several miles of displacement.

North-northeast Trending Faults

In the area shown (Fig. 24.1) the largest fault of the north-northeast trending set is Gate Fault. It terminates southward at Funeral Thrust (*see* Douglas and Norris, 1960) but its age relative to Funeral Thrust cannot be determined. In its southern exposures Gate Fault is a simple structure, dipping about 70 degrees west, that intersects bedding at a high angle in both hanging-wall and footwall. If movements were entirely dip-slip then translation cannot greatly exceed a stratigraphic separation of 1000 to 4000 feet. Northward, Gate Fault has drag-folds in both hanging-wall and footwall (e.g. locality A, Fig. 24.1) that affect only rocks adjacent to the fault. Northeast of Prairie Creek, Gate Fault cuts obliquely across an anticline-syncline pair comprising folds that appear to be larger than mere drag folds (locality B, Fig. 24.1). The fault juxtaposes the east limb of the anticline with the axial area of the syncline and, because they appear to be disrupted by Gate Fault, the folds probably predate the fault. The fault must extend into the core of the syncline; it is unexposed there, and may die out, but probably extends beneath Tundra Thrust. Tundra Thrust clearly overrides the syncline and probably overrides Gate Fault as well.

Right lateral wrench movement on Gate Fault is suggested by the right-lateral offsets of stratigraphic markers (locality B, Fig. 24.1). For example, the Headless and Nahanni formations, as exposed, are offset 5 or 6 miles. Such offset could of course be accomplished simply by erosion of a plunging anticline in the hanging-wall. Other weak evidence for lateral movement on this set of faults occurs on the adjacent fault to the east. At locality C (Fig. 24.1), deep-water shale, graptolitic limestone, and silty dolomite are juxtaposed with approximately time-equivalent shelf carbonates of the Sombre Formation across a steeply dipping north-northeast fault. That relationship is more easily explained by wrench than by dip-slip displacements. Right-lateral displacements on northerly trending faults have been reported elsewhere in the region. One example, is Hayhook Fault, about 120 miles to the northwest in Wrigley Lake and Glacier Lake map-areas. Gabrielse *et al.* (1973, p. 111) suggested right-lateral movement of 9 to 12 miles on that fault.

If Gate Fault and others originated as wrench faults, the reverse offsets which they now display may have developed at a later stage. In the area shown in Figure 24.1, the rejuvenated movements would all be west-side-up, whereas regionally some north-northeast trending faults are east-dipping, and motion has been east-side-up; for example, South Headless Fault (*see* Douglas and Norris, 1960). The suggestion that some compressional deformation in this area developed along previously formed structures is not new. The possibility that a pre-Laramide fracture system affected Laramide deformation was discussed by Douglas *et al.* (1970, p. 473). R.J.W. Douglas suggested to the writer (pers. com., 1971) that North Headless and South Headless Faults with their opposing dips and senses of movement (*see* Douglas and Norris, 1960), were localized by early structures.

Tundra Thrust

Tundra Thrust is a large fault which extends approximately 50 miles from about Deadman Valley to the south to within a few miles of North Nahanni River to the north (Douglas and Norris, 1960, p. 22). Tundra Thrust dips about 45 degrees northeastward and has the appearance of a classic bedding plane thrust, maintaining essentially the same stratigraphic sequence in its hanging-wall for most of its length. Displacement on the thrust may be several miles because, in places, it superimposes shelf carbonates of the Delorme, Camsell and Sombre formations in its hanging-wall with equivalent unnamed deep-water shale, graptolotic limestone, siltstone, silty dolomite, and dolomite in its footwall. Intermediate slope deposits including debris flow breccias occur at only one locality in the hanging-wall.

Tundra Thrust overrides one of the north-northeast trending faults, overrides the syncline which is cut by Gate Fault, and appears to override Gate Fault itself. Tundra Thrust clearly postdates the set of north-northeast trending steeply dipping faults.

Ages of Deformation

The ages of deformation phases cannot be sharply defined. All are post-Late Devonian and pre-Quaternary.

From: Report of Activities, Part A;
Geol. Surv. Can., Paper 77-1A (1977)

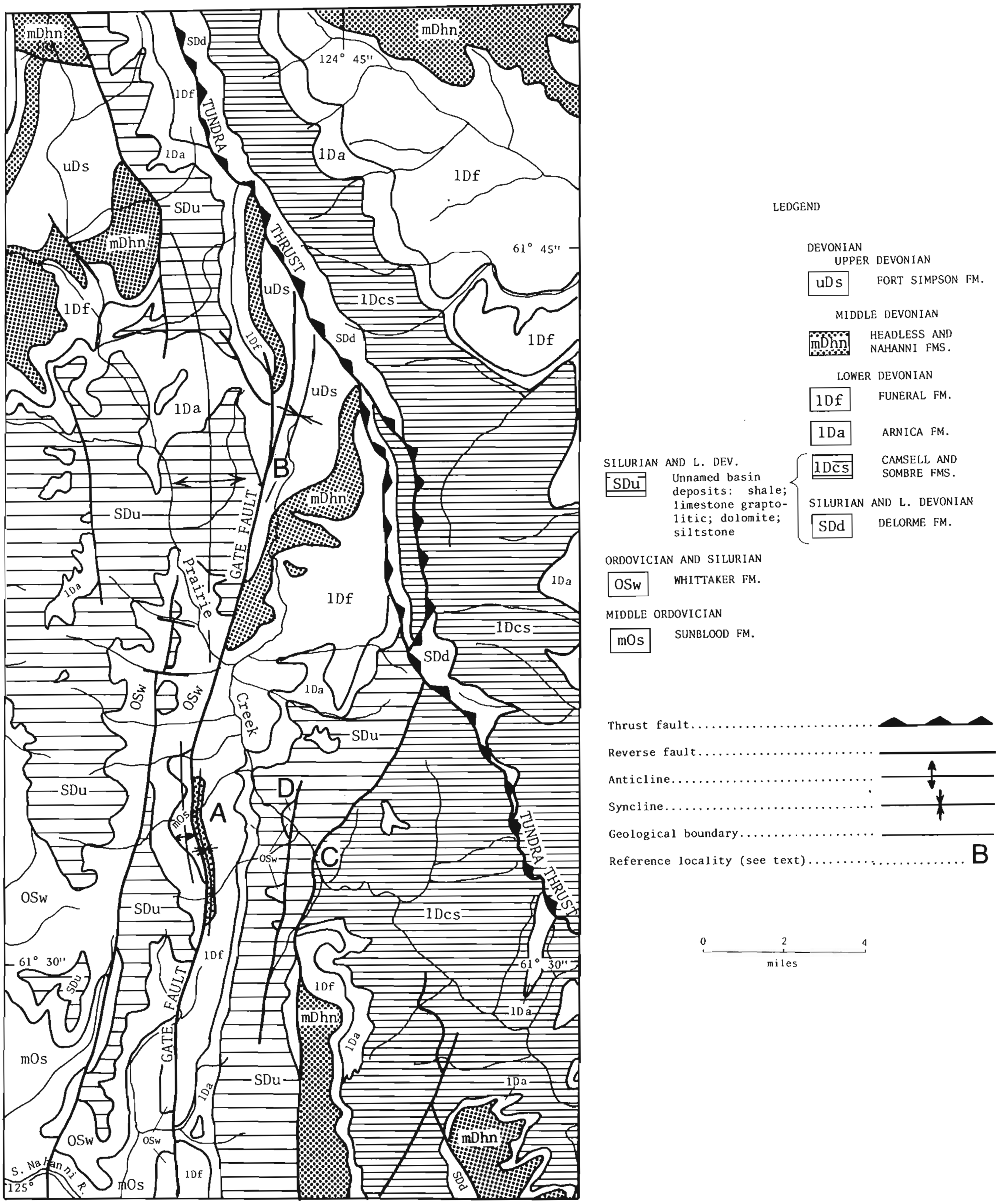

Figure 24.1. Sketch map showing relationships between north-northeast trending and northwest trending faults.

Conclusions

From the foregoing it can be stated that two distinct classes of fault exist and that the steeply dipping north-northeast trending faults are postdated by the more gently dipping Tundra Thrust.

Inter-relationships between various structures permit suggesting the following sequence of deformational events as a working model to be tested as detailed mapping progresses.

1. Development of north-trending folds.
2. Development of north-northeast trending, near-vertical, right-lateral wrench faults resulting in disruption of folds by Gate Fault, and juxtaposition of facies by another fault.
3. Northeast-southwest compression resulting in development of the northwest trending Tundra Thrust and major translation thereon. Simultaneous development of reverse movements on existing wrench faults.

Economic Considerations

The Cadillac lead-zinc deposit occurs in the area of Figure 24.1 (locality D). The deposit is a high-grade vein which dips eastward but probably is related to an adjacent westward-dipping, north-northeast trending fault. Where observed by the writer, the mineralized vein occurs in cherty dolomite of the Whittaker Formation, but it is not known if all mineralization is confined to that formation. If it is so restricted, then the Whittaker Formation adjacent to north-northeast trending faults becomes an obvious exploration target.

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GEOLOGY OF THE ATHAPUSCOW AULACOGEN, EAST ARM OF GREAT SLAVE LAKE, DISTRICT OF MACKENZIE

Project 660009

P.F. Hoffman, I.R. Bell, R.S. Hildebrand[1] and L. Thorstad[2]
Regional and Economic Geology Division

Introduction

This project was originally undertaken in 1966-67 as a study of Proterozoic sedimentary rocks in the East Arm of Great Slave Lake. From this work came a detailed description of the stratigraphy (Hoffman, 1968), a reconstruction of the depositional history (*idem*, 1969), an interpretation of the regional tectonics (*idem*, 1973), and a facies analysis of the important stromatolite-bearing formations (*idem*, 1974). The area is interpreted as an aulacogen, or failed rift arm, genetically related to the Coronation Geosyncline of late Aphebian age (Hoffman and others, 1974).

Many aspects of the structural and magmatic development of the aulacogen were still poorly understood, however, because of the lack of systematic large-scale mapping. The best maps, based on the canoe-reconnaissance work of Stockwell (1936), were published at a scale of four miles equals one inch. To improve them, the project was reactivated and, during the 1976 field season, all the Proterozoic rocks were remapped for publication at 1:50 000 scale. Most of the Archean rocks in the East Arm had been mapped in 1967-68 by the late E.W. Reinhardt, and his meticulous but largely unpublished work is being incorporated in the maps now in preparation. It is expected that the new 1:50 000 scale maps will be placed on open file early in 1977.

This report summarizes only new data and concepts and, for a more general description of the area, the reader is referred to the earlier publications mentioned above. Among the many highlights of the new mapping project, the following have especially intriguing regional implications:

(1) There are many huge nappes, up to 70 km in length, of recumbently-folded sedimentary rocks that moved, at the end of Great Slave Supergroup deposition,

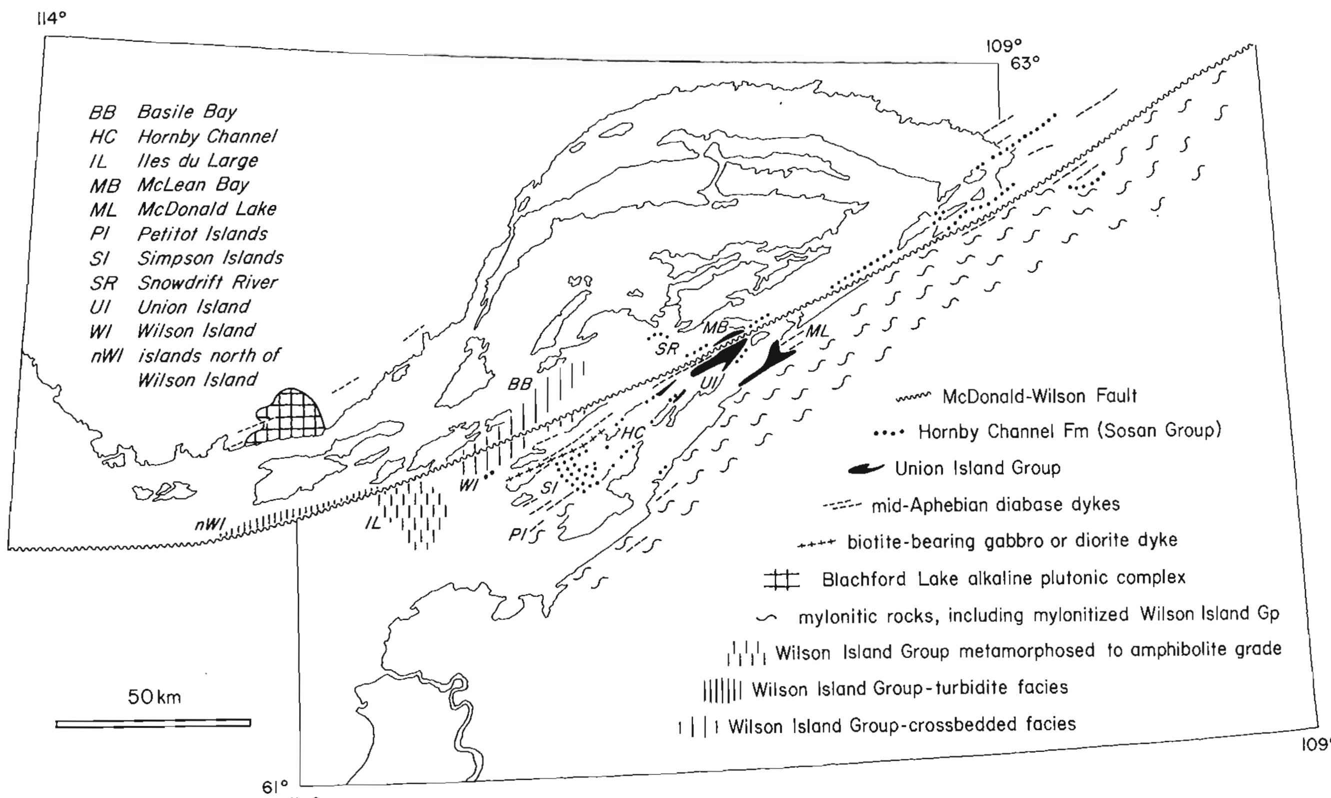

Figure 25.1.

Distribution of various pre-Kahochella Group rocks with the effects of strike-slip on the McDonald-Wilson Fault removed.

[1]University of California at Santa Barbara, Goleta, CA.

[2]University of British Columbia, Vancouver, B.C.

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northwestward into the East Arm from the region of tectonic denudation south of the McDonald-Wilson Fault. The nappes contain stratigraphic units that are correlative with, but of more basinal facies than, the subjacent autochthon. They indicate that the axial trough of the aulacogen originally lay south of the McDonald-Wilson Fault.

(2) The chaotic megabreccia, or olistostrome, of the Stark Formation is now believed to result from solution of a thick sequence of salt, deposited above the basinal carbonates of the Pethei Group. Solution collapse occurred in advance of the prograding fan-delta that deposited the overlying red clastic sediments of the Christie Bay Group. The internal structure of the megabreccia was subsequently complicated by its involvement in nappe tectonics, and later by the forceful intrusion of giant laccoliths, some more than 20 km in diameter, of diorite and monzonite.

(3) More than 35 volcanic centres have been located and are found in every formation of the Sosan and Kahochella groups. Most are basaltic and range from simple breccia pipes, through small (less than 1 km in diameter) submarine pyroclastic cones, to larger (more than 5 km) multicycle complexes or reworked tuffs and lavas. Many centres in the Kahochella Group occur along old fault lines, where displacements occurred during deposition of the underlying Sosan Group. A line of silicic centres extends from Seton Island to near Sachowia Point, where large rhyolite domes and ash-flow tuffs occur. In general, volcanic rocks are more extensive in the autochthon than in the nappes, indicating that volcanism was concentrated at the margin of the aulacogen, not in the axial trough.

(4) The Wilson Island Group, possibly of early or middle Aphebian age, records a stage of rifting, beginning with bimodal volcanic rocks and fanglomerate, that predates the main development of the aulacogen in the late Aphebian. The group was metamorphosed, locally to amphibolite grade, and involved in regional mylonitization of mid-Aphebian age, before a second cycle of rifting occurred, which initiated the aulacogen.

(5) An important swarm of diabase dykes, of mid-Aphebian age, extends the entire length of the East Arm. The dykes intrude the Archean basement rocks only, but are presumed to postdate the deformation of the Wilson Island Group. They pass beneath the sedimentary rocks of the aulacogen and are probably a manifestation of the cycle of rifting that initiated the aulacogen. An unusual biotite-bearing dyke, previously dated at 2170-2200 million years, is cut by the dyke swarm and provides a maximum age for the aulacogen.

(6) There is about 75 km of dextral strike-slip, estimated from displacements of Wilson Island and Union Island Group rocks, on what is here termed the McDonald-Wilson Fault. West of the Snowdrift River, the main locus of movement diverges from the McDonald Fault and skirts the north shore of Simpson Island and Wilson Island. The often photographed fault-line scarp on the south shore of McDonald Lake marks a mere splay fault of little strike-slip displacement. The extensive belt of mylonites south of the East Arm is not genetically related to the McDonald-Wilson Fault. Most of the Archean granites along the fault are relatively undeformed and, where mylonites do occur, they are clearly truncated by, and therefore older than, the fault.

Archean Basement Rocks and the southward extent of the Slave Province

Archean basement underlies all the Proterozoic rocks of the aulacogen. Most of it consists of massive adamellite and granite but, in the western Simpson Islands, there is a northwest-trending belt of gneisses and migmatite. The gneisses are an extension of the Archean metasedimentary belt exposed on the north shore of Hearne Channel near Narrow Island. Foliation in the gneisses dips gently to the northeast, projecting beneath the massive granitic rocks. Dips are about the same as those in the overlying Sosan Group (Table 25.1) and, therefore, the gneisses must have been nearly horizontal at the time of Sosan deposition.

Most of the Archean rocks along the McDonald-Wilson Fault are remarkably little deformed and the massive granitic rocks extend southward, as much as 10 km, to a northeast-trending belt of intensely lineated mylonites (Fig. 25.1). North of this belt, Archean radiometric ages have been obtained from micas even using the K-Ar method. So far as deformation of the basement is concerned, it is useful to consider that the Slave Province extends as far south as the north edge of the mylonite belt.

Wilson Island Group and the First Cycle of Rifting

This is an important group, over 6 km in thickness, of early or middle Aphebian sedimentary and volcanic rocks (Table 25.1). The oldest rocks, exposed south of Wilson Island, are an intimately mixed assemblage of basalt and rhyolite flows, and conglomerate containing volcanic, granitic and gneissic clasts. The granitic and gneissic clasts are derived from the underlying Archean basement, although no basal unconformity is exposed. On Wilson Island itself, the volcanic rocks are overlain by a thick, homoclinal, vertically dipping, northward facing sequence of cross-bedded quartzite, feldspathic quartzite, and dolomitic quartzite. Eliminating the displacement on the McDonald-Wilson Fault, the sequence on Wilson Island is juxtaposed against the other main outcrop belt of Wilson Island Group, that around Basile Bay (Fig. 25.1). There, stratigraphically above the sequence on Wilson Island, a sequence of impure dolomite, argillaceous quartzite, and argillite with local flows of porphyritic basalt is preserved in the core of a major synclinorium. Using the same pre-fault reconstruction, the quartzose turbidites exposed on the islands north of Wilson Island, a facies quite distinct from that on Wilson Island itself, are placed far to the west. The transition from the eastern cross-bedded facies to the western turbidite facies is consistent with the group's westerly directed paleocurrents (Yeo, 1976).

The Wilson Island Group is everywhere significantly more metamorphosed than the adjacent Great Slave Supergroup. At all of the localities mentioned above, the group is of greenschist grade. On the Iles du Large (Fig. 25.1), amphibolite grade is attained and the pelitic rocks contain prominent metacrysts tentatively identified as andalusite and staurolite. On some of the islands, notably Butte Island, the Wilson Island Group is intruded by deformed stocks and dykes of pink biotite adamellite, which have been sampled for geochronological studies.

On the Petitot Islands, a fault-bounded sliver of Wilson Island Group has suffered severe cataclastic deformation, as have the adjacent Archean gneisses. This area is probably an extension of the mylonite belt south of the McDonald Fault (Fig. 25.1). The Great Slave Supergroup, however, is not mylonitized in this area. Furthermore, on an island at the western end of Inconnu Channel, a conglomerate near the base of the Sosan Group, which there lies unconformably on Archean gneiss, contains mylonitized (and nonmylonitized) clasts of both Archean rocks and Wilson Island Group quartzite.

Despite the fact that the Wilson Island Group is nowhere exposed in unfaulted contact with either the Archean basement or the Great Slave Supergroup, the following conclusions can be made in light of the relations outlined above (Table 25.1):

(1) The Wilson Island Group is younger than the Archean and older than the Great Slave Supergroup.
(2) Mylonitization postdates the Wilson Island Group and predates the Great Slave Supergroup.
(3) Metamorphism of the Wilson Island Group probably also predates the Great Slave Group, although there is some uncertainty on this point because the juxtaposition of metamorphosed Wilson Island Group and unmetamorphosed Great Slave Supergroup may be the result of younger nappe tectonics. If, however, metamorphism can be proved to predate mylonitization, then it must also predate the Great Slave Supergroup.

Yeo (1976) has suggested, on the basis of paleocurrent data, that the Wilson Island Group was deposited in a precursor trough coincident with the Athapuscow Aulacogen. That this trough originated by rifting is indicated by the occurrence, at the base of the Wilson Island Group, of the classic rift-valley assemblage of compositionally bimodal volcanic rocks and conglomerate. This cycle of rifting is separated, however, from the main development of the aulacogen by a period of intense mylonitization and, probably, thermal metamorphism of mid-Aphebian age. The earlier cycle of rifting may not bear the same genetic relationship to the Coronation Geosyncline as does the later but, by providing a zone of crustal weakness, it may ultimately have predetermined the site of the aulacogen.

Mid-Aphebian Diabase Dyke Swarm and the Beginnings of the Second Cycle of Rifting

An important swarm of mid-Aphebian diabase dykes extends, on an east-northeast trend, the entire length of the East Arm (Fig. 25.1). The dykes intrude only the Archean basement rocks, and are most numerous at the northeast end of Simpson Island and south of the McDonald Fault east of the Snowdrift River. On the southern Simpson Islands, they are overlain by the Sosan Group and, southwest of McDonald Lake, they pass beneath the Union Island Group (Table 25.1). Dykes of the same trend intrude the Blachford Lake alkaline plutonic complex (Davidson, 1972) north of Hearne Channel and, south of Hornby Channel, they intrude mylonitic gneisses (Reinhardt, 1972).

In the west-central Simpson Islands, there is an unusual dyke, almost 30 km long, of biotite-bearing gabbro or diorite (Fig. 25.1). The biotite, which is of igneous origin, has been dated radiometrically by the K-Ar method at 2200 (Burwash and Baadsgaard, 1962, p. 28) and 2170 (Leech *et al.*, 1963, p. 61) million years. There has long been confusion, however, concerning the relation of this dyke to the Sosan Group, with which, in its eastern part, it is in fault contact. Reinhardt's unpublished maps clearly show, and we have confirmed, that this dyke is cut by numerous diabases of the swarm that passes beneath the Sosan Group and must, therefore, be older. Part of the confusion arose from reports that a "trachytic" border phase of the dyke closely resembles other dykes, associated with diatreme breccias (*see* Reinhardt, 1972) occurring along old fault lines, that do intrude the Sosan Group. Mapping shows that the "trachyte" occurs only where the dyke is faulted against the Sosan Group and is not a normal border phase of the dyke. The "trachyte", if it is related to the diatreme breccias, probably intrudes the fault and is, therefore, younger than the biotite-bearing dyke.

If we have interpreted these relations correctly, it can be concluded that:

(1) The diabase dyke swarm is younger, although probably not very much younger, than 2170-2200 million years.

(2) The dyke swarm is older, although probably not very much older, than the Union Island Group and Great Slave Supergroup.

(3) The dyke swarm apparently post-dates the mylonite belt south of the McDonald Fault and, therefore, should be younger than the Wilson Island Group. There is uncertainty on this point, however, because the dyke swarm is not known to intrude the Wilson Island Group itself. The group is cut by basic dykes, but they are metamorphosed and of more northerly trend. They could be local feeders to the upper Wilson Island Group basalt flows. Whether involvement of the group in nappe tectonics may explain this anomaly we cannot say, but there remains some ambiguity concerning the relative ages of dyke emplacement, mylonitization, metamorphism, and the Wilson Island Group. For now, we consider the dyke swarm, including the dated biotite-bearing dyke, to mark the beginnings of the second cycle of rifting in the East Arm (Table 25.1).

Union Island Group:
Internal Stratigraphy and Age Relations

This is an interesting group of stagnant basin sediments, subaqueous basalts and intrusive gabbros. They are the oldest supracrustal rocks in the aulacogen that can be correlated with rocks in the Coronation Geosyncline and were the first to be deposited in the second cycle of rifting. It is exposed almost exclusively in the vicinity of Union Island and, previously, its internal stratigraphy and relation to the Great Slave Supergroup were uncertain.

The group was deposited on an irregular surface of Archean granitic rocks and is not cut by the mid-Aphebian diabase dyke swarm. Five mappable stratigraphic units are recognized (Fig. 25.2):

(1) The lower dolomite is unbedded and characteristically has complex vein systems of alternating quartz and radiaxial dolomite. Some veins are filled in part by clastic sediments. Ancient hills of granite, flanked by narrow aprons of fossil talus breccia, stick up into the dolomite and, in places, even into the overlying black shale. Beds of quartzite and quartz-pebble conglomerate occur in depressions on the granite surface at the base of the dolomite.

(2) The black shale is fissile, highly carbonaceous, and contains abundant iron sulphide nodules. In the lower part, there are many beds of black dolomite, commonly with shale rip-up clasts. In the upper part, there are quartzitic turbidites. Southwest of McDonald Lake, on the south shore of the East Arm, the black shale is intruded by two gabbro sills, the lower of which is glomero-porphyritic.

(3) The volcanic rocks, which do not everywhere occur at the same stratigraphic horizon and which are missing altogether north of Union Island, consist of wedge-shaped units amygdaloidal flows and flow breccia, pillow lavas and pillow breccia, coarse poorly bedded aquagene tuff, and laminated tuffaceous sediments. All are basaltic in composition, save for rare tuff beds which may contain more silicic pyroclasts. Many of the pillows have cement-filled centres, and were presumably lava tubes.

(4) The upper dolomite is distinguished from the lower by being well bedded and, for the most part, laminated. Slump breccia beds are common but sedimentary structures resulting from traction currents are rare. There are many reddish mudstone beds and, in places, the dolomite contains lenses of pyritic quartz-pebble conglomerate.

(5) The upper red and green mudstone is laminated, contains abundant slump scars, and has soft-sediment thrust and extension faults. There is no evidence of subaerial exposure.

The Union Island Group is overlain unconformably by the Sosan Group (Table 25.1). On an island northeast of Union Island, the basal Sosan is exposed in contact with the red and green mudstone unit. On north-central Union Island, although the contact is covered, the Sosan truncates the black shale and lower dolomite. On the south side of Simpson Island, across Hornby Channel from central Union Island, the basal Sosan overlies the upper dolomite, although here also the contact is covered. Nevertheless, we believe the stratigraphic position of the group to be established and it will be formally subdivided into formations.

Sosan Group:
Facies Relations and Contemporaneous Block Faulting

The Sosan Group is formally subdivided into four formations, which occur in simple stratigraphic order at the type section north of Lac Duhamel, from the top:

(4) Akaitcho River Formation: red micaceous sandstone,
(3) Kluziai Formation: white to pink quartzite,
(2) Duhamel Formation: stromatolitic dolomite, and
(1) Hornby Channel Formation: pebbly feldspathic granulestone.

Locally, there is a newly discovered unit, possibly correlative with the Quadyuk Formation of the Goulburn Group at Bathurst Inlet (Campbell and Cecile, 1976), of laminated argillaceous dolomite that overlies the Akaitcho River Formation.

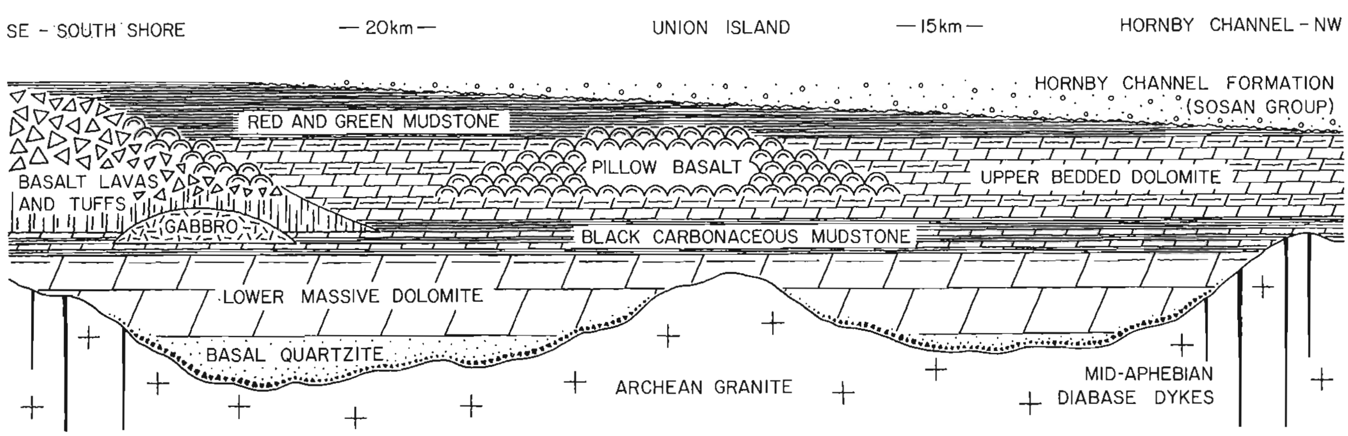

Figure 25.2. Stratigraphy and facies relations in the Union Island Group.

Table 25.1

Summary of Proterozoic sedimentation, magmatism and tectonics in the East Arm of Great Slave Lake

	diabase dykes (Mackenzie Swarm)	
	funnel-shaped gabbro intrusion	
Et-then Group	minor basalt	strike-slip faulting
	unconformity	
		folding
	diorite-monzonite laccoliths	
		movement of nappes
Christie Bay Group	basalt	
Stark megabreccia	minor basalt	
Pethei Group		
	gabbro intrusions	
Kahochella Group	basalt and rhyolite	
Sosan Group	basalt and felsic porphyry	block faulting
	gabbro intrusions	block faulting
Union Island Group	basalt	
	unconformity	
	diabase dyke swarm	
	biotite-bearing gabbro or diorite dyke	
		mylonitization
	adamellite stocks	metamorphism
Wilson Island Group	basalt and rhyolite	
	unconformity	
	Archean basement rocks	

Regional facies analysis is greatly complicated by four factors. Firstly, there is considerably more intertonguing of the principal facies that define each formation than is apparent in the type section. Kluziai-like tongues are common near the top of the Akaitcho River Formation, Hornby Channel-like tongues occur in the Kluziai Formation and *vice versa*, both Hornby Channel and Kluziai-like tongues occur in the Duhamel Formation, which is of limited areal distribution. The facies relations are especially complex around Charlton Bay, at the extreme northeast end of the East Arm, where the group as a whole is relatively thin. Nevertheless, it has been decided to retain the formational nomenclature and, so far as is possible at 1:50 000 scale, to map the intertonguing relations. The intertonguing is a reflection of the complexity and instability of depositional environments that range from braided rivers (Hornby Channel facies), barrier islands and distributary-mouth bars (Kluziai facies), interdeltaic tidal flats (Duhamel facies), to barrier-protected lagoons and the nearshore open shelf (Akaitcho River facies).

Secondly, there are old faults, apparently active during Sosan deposition, across which there are abrupt changes in stratigraphy. For example, there is a northeast-trending line of volcanic centres between Lac Duhamel and McLean Bay north of which, the Hornby Channel Formation is more than 200 m thick. Only 2 km south of Lac Duhamel, however, on a ridge where Archean granite is exposed, the formation is as little as 2 m thick. South of the ridge, the formation is again relatively thick. The Duhamel Formation undergoes comparable changes in thickness over this paleotopographic high. Another old fault, also marked by a line of volcanic centres, occurs at Taltheilei Narrows and bounds a paleotopographic high east of the narrows (Fig. 25.3), against which the Sosan Group must abut in the subsurface. The complexities of the block-fault topography that probably existed during Sosan deposition, especially the lower Sosan, is only hinted at in the limited areas in which the group is now exposed.

Thirdly, although most of the Sosan exposures are autochthonous, the 20-km-long outcrop belt that passes between Bunting Lake and Moose Lake is a nappe (Fig. 25.4). A 2-km-long klippe of Sosan quartzite, possibly once part of the same nappe, rests on the Stark megabreccia south of Meridian Lake. The northwestward transport of these allochthonous masses must be considered in making paleogeographic reconstructions of the Sosan Group.

Fourthly, the effects of strike-slip displacement on the McDonald-Wilson Fault must be accounted for. Palinspastic restoration of the fault effectively moves the important Sosan exposures of the Hornby Channel region much closer to those of the type area around the Snowdrift River (Fig. 25.1). It also explains why the Sosan sediments around Seton Island, almost overwhelmed by volcanic products, are so different from the volcanic-free Sosan rocks located only a short distance away on the other side of the fault.

Kahochella Group and the Seton Volcanic Centres

The Kahochella Group is formally subdivided into three sedimentary formations, from the top:

(3) Charlton Bay Formation: green concretionary shale,
(2) McLeod Bay Formation: red concretionary shale, and
(1) Gibralter Formation: red shale and siltstone.

Volcanic rocks, occurring in both this and the Sosan Group, have collectively been referred to the Seton Formation. Most of the new information concerns the distinction in facies between allochthonous and autochthonous sections, and the areal and stratigraphic distribution of volcanic rocks.

The difference in sedimentary facies between the autochthonous rocks and those in the nappes is not great. In general, the McLeod Bay Formation is thinner in the nappes, and both the Charlton Bay and Gibralter formations are thicker and contain siltstone turbidites. The turbidites in the Gibralter Formation are especially numerous in the 25-km-long nappe that is best exposed between Hair Lake and Bunting Lake, near the east end of the aulacogen (Fig. 25.4).

The regional distribution of volcanic centres is significant (Fig. 25.3). Most are concentrated either along the northern margin of the axial trough of the aulacogen, or along the old fault line that bounds the basement high east of Taltheilei Narrows. Rocks of the axial trough itself, whether in the autochthon south of the McDonald-Wilson Fault or in the nappes to the north, are relatively deficient in volcanic products. In general, the volume of volcanic rocks decreases from west to east, that is, away from the mouth of the aulacogen, and there are no centres at all on the platform north of Christie Bay, east of Taltheilei Narrows.

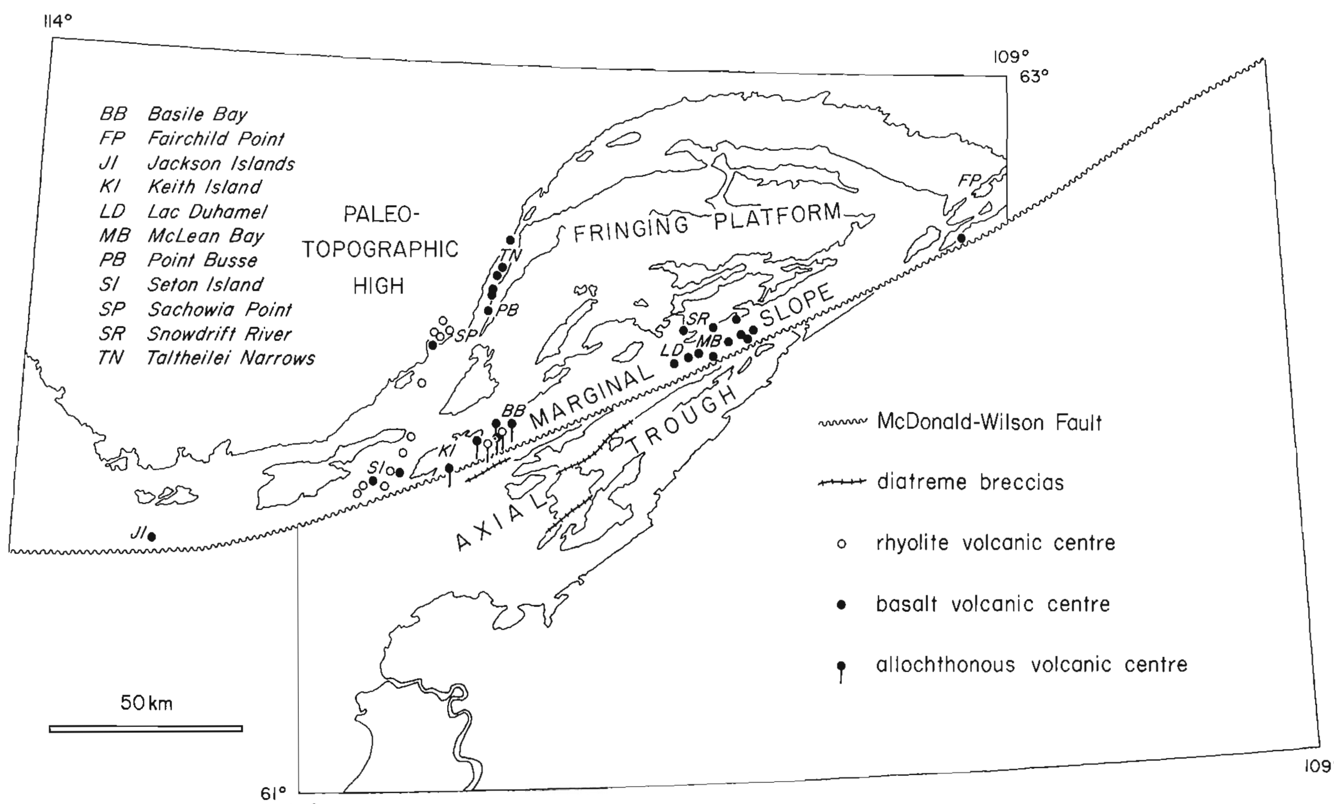

Figure 25.3. Regional distribution of Seton volcanic centres and their relation to major tectonic elements in the aulacogen.

Several clusters of volcanic centres are worth describing individually. In the Snowdrift River area (Fig. 25.3), there are no less than twelve volcanic breccia pipes and eight submarine pyroclastic cones, not to mention numerous tuffaceous beds of unknown derivation. All are basaltic in composition. Half of the pipes intrude the Akaitcho River Formation but their total stratigraphic range is from the Kluziai to McLeod Bay Formation. Of the cones, four occur in the Charlton Bay Formation, all located north of McLean Bay, three are in the lower part of the Gibralter Formation, and one occurs just above the base of the Akaitcho River Formation. Most of the cones are less than 1 km in diameter. A dark green concretionary shale, lithologically not unlike the Charlton Bay Formation, occurs at the base of the Gibralter Formation, and may be related to volcanism.

A line of six, possibly seven, basaltic breccia pipes occur between Taltheilei Narrows and Point Busse (Fig. 25.3). The pipes are irregular in detail and the southern four all intrude the Gibralter Formation. Basalt flows occur at the base of the formation and beds of lapilli tuff, some with large-scale cross-bedding, are associated with green shale and beds of red granular ironstone at the top of the formation. The pipes may well have fed the upper tuffs. The most northerly pipes, those on the west side of Taltheilei Narrows, are mineralized and very poorly exposed. They intrude the Akaitcho River Formation and, if they are the same age as those to the south, are more deeply eroded. This suggests that the southern pipes may be mineralized at depth.

A unique line of silicic, probably rhyolitic, centres extends from near Sachowia Point to Seton Island (Fig. 25.3). West of Sachowia Point, a complex of flow-banded rhyolite domes and intimately-related welded ash-flow tuffs cover an area of almost 35 km^2. The rhyolite overlies basalt flows, probably correlative with those near Taltheilei Narrows, and are intruded by basic dykes and sills. Basic and silicic rocks are also intimately associated on Seton Island and nearby islands, the thickest volcanic complex in the East Arm. Five funnel-shaped bodies of albite porphyry intrude an assemblage of basalt flows, tuffs and their erosion products, which are intercalated with sediments of the Kluziai and Akaitcho River formations. The felsic porphyry bodies also contributed detritus and must, therefore, have been unroofed during volcanism or had extrusive equivalents.

Nappes bearing volcanic rocks deposited in the axial trough of the aulacogen occur in a structurally complex belt between Keith Island and Basile Bay (Fig. 25.4). There are two major nappes, both of which contain locally thick piles of submarine volcanic rocks in the upper Gibralter and McLeod Bay formations. Amygdaloidal pillow basalts and pillow breccias are by

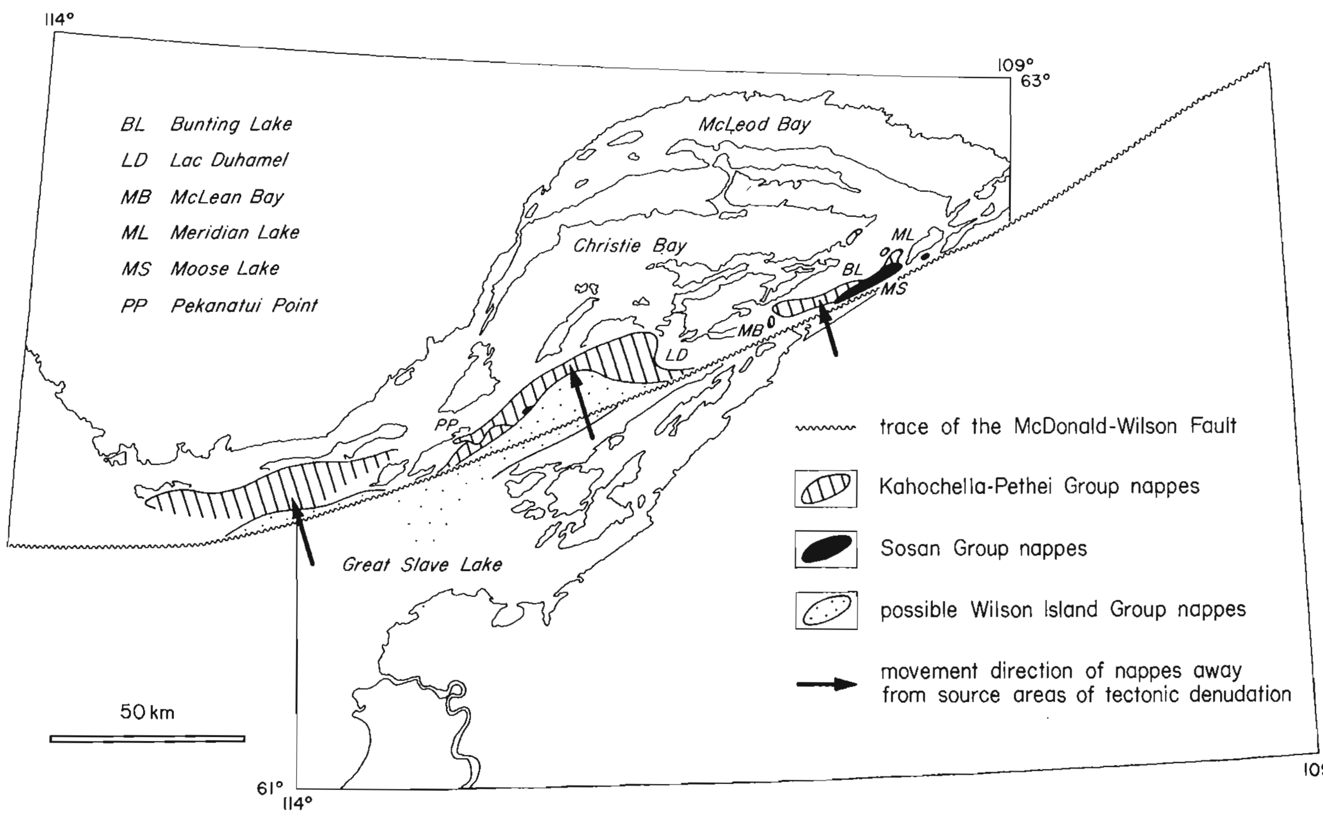

Figure 25.4. Distribution of major nappes in the East Arm and their source areas of tectonic denudation.

far the most voluminous volcanic rocks, but felsic, probably rhyolitic, domes and their erosion products also occur. Most of the shales in the vicinity of the volcanic piles are green, rather than the usual red, and beds of red granular ironstone, commonly associated with red and white chert lenses, are spatially related to the volcanic piles.

Intrusive gabbros, of irregular shape and variable texture, are closely associated with Seton volcanic rocks, both in the nappes and the autochthon. They are prominently exposed on the Jackson Islands, near the west end of the East Arm, where they intrude basalt flows, tuffs and a breccia pipe in the Duhamel and Kluziai formations; and on Fairchild Point, near the east end of the East Arm, where they intrude tuffaceous sandstones of the Kluziai and Akaitcho River formations. They are cut by many high-angle oblique-slip faults, which serves to distinguish them from the much younger dykes and sills of the Mackenzie Swarm. They constitute an important set of basic intrusions, probably closely related in age and origin to the Seton volcanic rocks.

Pethei Group and the Paleobathymetric Zonation of the Aulacogen

Facies analysis of this group led to the recognition of three paleobathymetric zones (Fig. 25.3), from the south:

(1) An axial trough, where tongues of greywacke turbidites (Blanchet Formation), derived from the west, are interbedded with deepwater stromatolitic mudstone (McLean Formation) and rhythmically laminated limestone (Pekanatui Point Formation),

(2) A marginal slope, where the McLean and Pekanatui Point formations occur without greywacke turbidites and are more calcareous than in the axial trough,

(3) A fringing platform, where two formations of shallow-water stromatolitic dolomite and limestone occur at the top of the marginal slope. The Taltheilei Formation, the older, occurs on the flanks of the old structural high east of Taltheilei Narrows but the Wildbread Formation, the younger, extends almost the entire length of the East Arm.

An appreciation of this zonation is essential in interpreting the relative movements of the various nappes. The largest nappes have transported Pethei of the axial trough northward onto the autochthonous rocks of the marginal slope. Smaller nappes of marginal slope facies have moved onto the fringing platform. Movement of the nappes has telescoped the facies changes, such that the axial trough was originally some 20 km south of the platform edge, not 10 km as previously reported. Mapping has also revealed that greywacke turbidites in the axial trough extend almost to Meridian Lake, 60 km farther to the northeast than previously known.

Stark Formation: Salt Solution-Collapse Megabreccia?

The Stark megabreccia, or olistostrome, is the most difficult formation in the East Arm to interpret, and is one of the most important because of the occurrence of a very similar megabreccia at a precisely correlative stratigraphic level in the Goulburn Group at Bathurst Inlet, 500 km to the north (Campbell and Cecile, 1976). The megabreccia rests sharply on the Pethei Group and consists of chaotically dispersed blocks of stromatolitic limestone and dolomite, containing abundant shallow-water sedimentary structures, in a brecciated matrix of red mudstone. The carbonate blocks, which are less than 50 m thick but range up to 1 km in length, are indigenous to the Stark Formation. None of the blocks is derived from the underlying Pethei Group, nor does the brecciation extend downwards into the Pethei. The carbonate of the blocks was intimately interbedded with the mudstone before brecciation. In a few places, notably 1.5 km south of the Snowdrift townsite, there are rare, large blocks of pillow basalt and basalt breccia.

The top of the megabreccia is generally indistinct and varies considerably in stratigraphic level. Overlying the main part of the megabreccia, that which contains the carbonate blocks, is an interval of thin bedded, red mudstone. The mudstone contains ripple marks and mudcracks, and becomes much more sandy toward the western end of the East Arm. In many places, notably in Tochatwi Bay, it contains numerous beds of carbonate-pebble conglomerate and granulestone, clearly derived from carbonate blocks in the underlying megabreccia, some of which project as pinnacles into the mudstone from below. The mudstone is overlain conformably by cross-bedded, lithic and feldspathic red sandstone of the Tochatwi Formation, which also contains beds of carbonate granulestone derived from blocks in the megabreccia. Both the mudstone and the sandstone are themselves extensively brecciated in places, notably at the east end of Tochatwi Bay and on the Caribou Islands.

It is difficult to measure the thickness of the megabreccia, but the Stark Formation as a whole ranges from about 500 to 700 m. There are no complete sections of the formation in which the megabreccia is not at least a prominent lower member. Thus, the extent of the megabreccia in the East Arm alone is 250 km.

Salt crystal casts occur both above and below the megabreccia. The uppermost bedding surface of the Pekanatui Point Formation, at the top of the Pethei Group, is littered with hopper-shaped salt casts and the red mudstones, above the megabreccia, contain highly skeletal salt casts, up to several centimetres in size, the poikilotopic nature of which suggests *in situ* precipitation from hypersaline groundwater.

Another unique feature is the presence of travertine-like coatings, up to 1 m thick, that were precipitated during brecciation. They are mostly well laminated

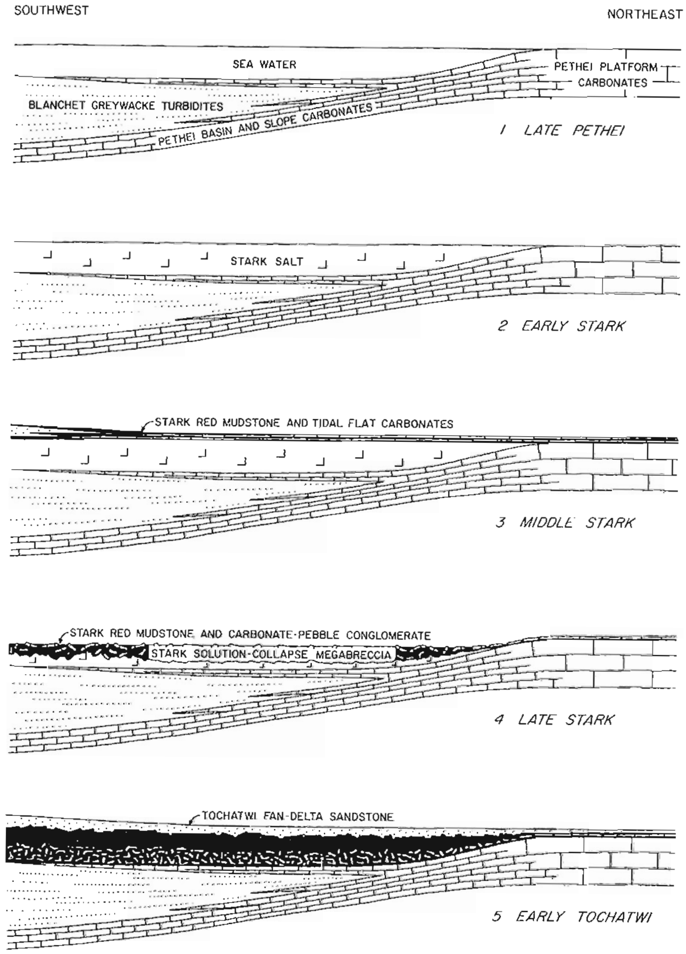

Figure 25.5. Genetic model for development of the Stark megabreccia by salt solution collapse.

and contain breccia layers that are draped by succeeding laminations. They are commonly riddled with small prismatic crystals of low-temperature quartz. The travertine occurs as detached blocks within the megabreccia, as discordant coatings that accreted downwards from the undersides of other blocks, and as a more or less continuous encrustation on the upper surface of the Pethei Group at the base of the megabreccia.

Origin of the megabreccia by salt solution-collapse was previously discounted because of the abundance of contorted and overturned carbonate blocks indicative of lateral transport by slumping. Much of this is probably related, however, to subsequent involvement of the megabreccia in nappe tectonics and the forceful intrusion of igneous laccoliths (*see* below). The absence of blocks derived from older formations, which make up the nappes, indicate that it is not a type of "precursory olistostrome" (*see* Elter and Trevisan, 1973) related to nappe formation. That the laccoliths are not ultimately responsible for the megabreccia is indicated by the fact that they do not control its distribution. Finally, neither nappe formation nor the intrusion of laccoliths, as possible causes of the megabreccia, can explain its occurrence in the Goulburn Group, whereas the deposition of salt may be a regional phenomenon.

According to the genetic model we now favour (Fig. 25.5), the end of Pethei carbonate deposition was marked by the precipitation of probably several tens of metres of salt, a thickness sufficient to nearly fill the axial trough of the aulacogen. Red mudstone and shallow-water carbonates of the Stark Formation were deposited on top of the salt in front of the alluvial Tochatwi fan-delta that was prograding into the East Arm from the west. Downward percolation of relatively fresh water from the fan-delta began to dissolve the salt at depth, resulting in subsidence by solution collapse in advance of the fan-delta. This produced a karst-like surface, on which detritus derived locally from tilted blocks of resistant carbonate was mixed with far-travelled sand and mud of the fan-delta. Solution of salt produced ephemeral caves, in which travertine was precipitated. As the caves were destroyed by downward collapse of the overlying sediments, hypersaline solutions were displaced upwards and, far in front of the fan-delta, precipitated poikilotopic salt crystals in the red mudstones above the main part of the megabreccia. As solution of salt and progradation of the fan-delta were contemporaneous, the upper Stark and lower Tochatwi were involved, to varying degrees, in the collapse breccias. Rotation of very large blocks during the final stages of solution collapse may be responsible for the aberrant paleomagnetic results obtained even from seemingly unbrecciated outcrops of these formations (*see* Bingham and Evans, 1976; Evans and Bingham, 1976).

Christie Bay Group: A new formation above the Pearson Basalt

This group consists of red nonmarine detrital sediments and subaerial basalt flows. Previously, the basalt flows, named Pearson Formation, were believed to be the youngest unit. On the south side of Tochatwi Bay, at latitude 62°33', longitude 110°14', about 200 m of buff feldspathic sandstone, with an interval of

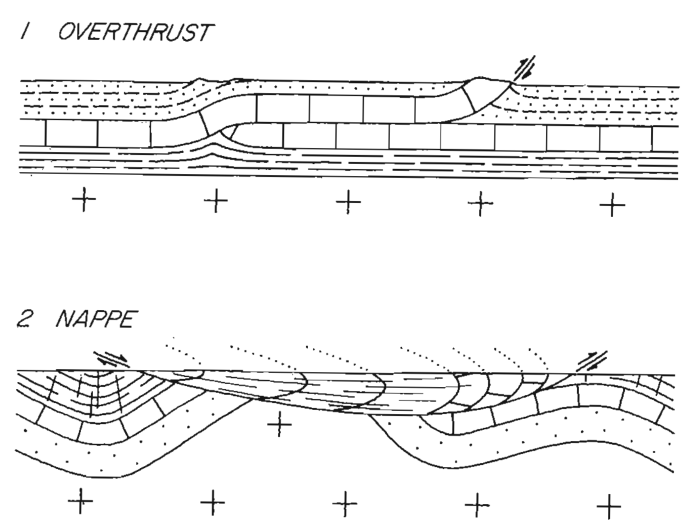

Figure 25.6. Distinction between a typical listric thrust fault and a nappe of the East Arm type.

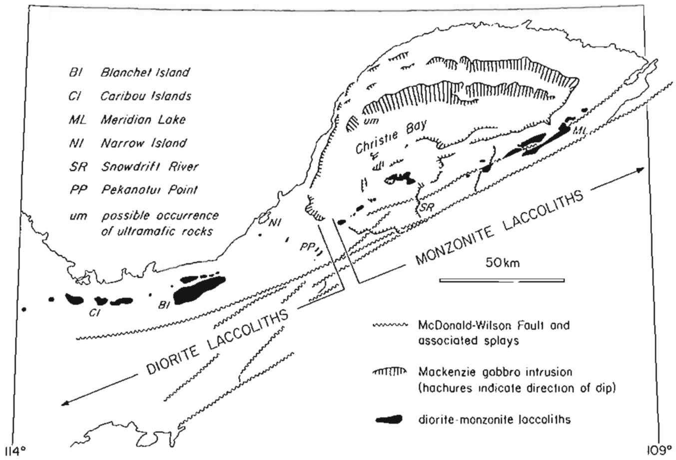

Figure 25.7. Distribution of diorite and monzonite laccoliths and the funnel-shaped gabbro intrusion centred at Christie Bay.

laminated mudstone at its base, overlies the Pearson Formation. This new formation may be correlative with the Amagok or uppermost Brown Sound Formation of the Goulburn Group at Bathurst Inlet (Campbell and Cecile, 1976).

Nappe Tectonics and the Tectonic Denudation of the Axial Trough of the Aulacogen

Nappes are the most spectacular structural feature of the East Arm, although their presence was unsuspected when mapping began. Huge rock masses, up to 70 km in length and 10 km in width, have been transported northwestward out of what was the axial trough of the aulacogen, now a zone of tectonic denudation. The transported rocks extend stratigraphically from Sosan Group to Stark Formation, and the Wilson Island Group may be involved as well. The nappes were emplaced before intrusion of the diorite-monzonite laccoliths, deposition of the Et-then Group, or strike-slip on the high-angle McDonald-Wilson Fault. There are important differences between the geometries of the East Arm nappes and such belts of listric thrust faults as occur, for example, in the foreland of the Coronation Geosyncline (Fig. 25.6). The nappes are unrooted at the present level of erosion, which means that they have a trailing edge as well as a leading edge. Rocks within the nappes are steeply homoclinal or recumbently folded, and their facing and vergence is invariably in the direction of transport. The detachment surface, on which the nappe has moved, is generally discordant with the strata both above and below. The nappes have either or both younger-over-older and older-over-younger relations with the rocks beneath them. In cases of younger-over-older superposition, the missing rocks are presumed to have been removed in an earlier nappe, moving in front of the nappe in question. This is unlike the situation in most listric thrust belts, where older-over-younger relations are the rule.

The three largest nappes (Fig. 25.4) are composed of Kahochella and Pethei Group rocks. The Pethei is of the axial trough facies, whereas that in the underlying autochthon is of the marginal slope. The Pethei generally occurs at the front of the nappes, where the crumpling is most severe, and the Kahochella trails behind. Folds are mostly recumbent, commonly overturned, and cleavage is gently dipping. This contrasts with the upright folds and weak high-angle cleavage in the autochthon. Of the three largest nappes, the central one is the most interesting. Near its eastern end, around Lac Duhamel, it transgresses autochthonous rocks ranging stratigraphically from Archean granite to Stark megabreccia. Near its western end, around Pekanatui Point, there is a stacked pair of nappes, each with a complete Kahochella-Pethei succession and both having internal thrust faults. Rocks both above and below the basal detachment surface of large nappes may be highly altered, and this zone is commonly complicated by many smaller nappes, torn from either the nappe or the autochthon. Intense alteration of autochthonous rocks north of McLean Bay may be related to a western extension, now removed by erosion, of the 35-km-long nappe between McLean Bay and Meridian Lake, the easternmost of the three largest nappes. Using reconstructions eliminating strike-slip on the McDonald-Wilson Fault system, it can be seen that the easternmost of the three largest nappes is derived from the region south of McDonald Lake, the central one from the Simpson Islands area, and the westernmost from the area covered by Great Slave Lake west of the Outpost Islands.

Overlying the Kahochella-Pethei nappes, and in places overlapping onto the autochthon, are a few smaller nappes made up of Sosan Group. The nappe that angles between Bunting Lake and Moose Lake is by far the largest of these (Fig. 25.4). It is at least 20 km in length, 27 km if the small klippe south of Meridian Lake is included, and contains a very thick section of the Hornby Channel Formation. Most of the other Sosan exposures, however, are autochthonous.

It is possible that the Wilson Island Group is allochthonous. This would help explain why it is nowhere seen in unfaulted contact with the Archean basement or the younger Sosan Group. If so, the fact that the grade of metamorphism is higher in the Wilson Island Group than in the adjacent Great Slave Supergroup cannot be used to determine the age of metamorphism. It is difficult to prove that the Wilson Island Group is involved in nappe tectonics because most of the critical contacts are either underwater or obscured by high-angle faults unrelated to nappe formation.

The age of nappe formation seems tightly constrained. Although the Stark megabreccia is the youngest formation carried in the nappes, their age is believed to post-date the entire Great Slave Supergroup. It seems improbable that the nappes could have been emplaced during Great Slave sedimentation and leave no evidence in the form of precursory fanglomerates or olistostromes. The Stark megabreccia itself contains no blocks of pre-Stark age and is very unlikely, therefore, to be derived from the nappes. The sandstones above the Stark are generally far too feldspathic to have a sedimentary

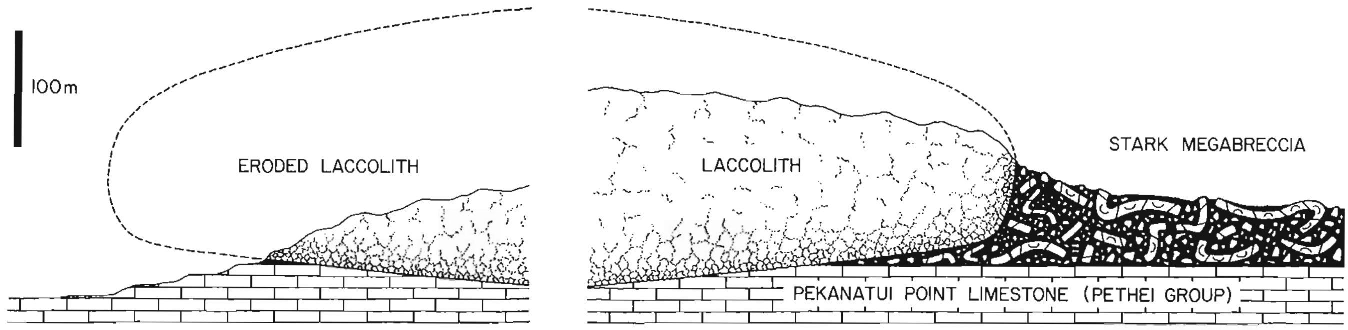

Figure 25.8. Intrusive relations of a typical East Arm laccolith, showing the distinction between a true lateral margin (right) and an eroded margin (left).

provenance as would be required if the nappes were being emplaced at that time. The minimum age of nappe development is constrained by the intrusion of diorite-monzonite laccoliths (*see* below), many of which override contacts between allochthonous and autochthonous rocks.

It is too soon for us to speculate on the mechanics of nappe formation, especially on the contentious issue of the extent to which the nappes were gravity driven. However, it can be concluded that they are fundamentally related to uplift and tectonic denudation of what had previously been the axial trough of the aulacogen.

Diorite-Monzonite Laccoliths: Composition, shape and age of intrusion

There is a string of more than twenty igneous intrusions, the largest 25 km in length, that extends almost the entire length of the East Arm (Fig. 25.7). Compositionally, there are two distinct sets. Those west of longitude 111°30' are hornblende-biotite diorites although, at their borders, they contain more acidic dykes and veins. Those to the east are mostly porphyritic monzonites with phenocrysts, in rapidly decreasing order of abundance, of plagioclase, hornblende and quartz. Both sets are compositionally and texturally similar to certain synvolcanic intrusions in the Great Bear Batholith (*see* Hoffman and McGlynn, in press), and may be about the same age.

The vast majority of intrusions are localized by the Pethei-Stark contact (Fig. 25.8). They tend to have flat bottoms that are nearly concordant with limestone of the Pekanatui Point Formation. In places, thin thrust slices of limestone have been driven out from beneath the laccoliths into the overlying Stark megabreccia. The laccoliths themselves have steep-walled lateral margins that abut against, and have pushed aside, the megabreccia. The tops of the laccoliths are rarely exposed but contain numerous large blocks of highly altered megabreccia. The margins of the laccoliths are characterized by autoclastic breccias and detached igneous blocks are locally abundant in the Stark megabreccia within a few metres of the contact. Such contacts are especially well exposed along the west shores of Blanchet Island and the Caribou Islands. The contrast between steep contacts with Stark megabreccia and flat contacts with Pethei limestone serve to distinguish the original lateral margins of laccoliths from merely erosional margins. The small sill-like diorite intrusions in the McLeod Bay and Charlton formations south of Pekanatui Point, and an intrusion in the Seton Formation on a solitary small island in Hearne Channel south of Narrow Island, are exceptional in their stratigraphic positions.

The fact that the laccoliths are so strongly localized stratigraphically tempts one to presume that they predate the nappes. However, several of the laccoliths override contacts between allochthonous and autochthonous rocks, thus necessitating that the laccoliths postdate nappe formation. This is especially well shown in the area west of Meridian Lake. In many areas, steep dips in the Pethei Group flatten beneath the laccoliths, suggesting that folding, as opposed to napping, continued after intrusion of the laccoliths.

McDonald-Wilson Fault: Strike-slip displacement and relation to Et-then sedimentation

A complex system of dextral strike-slip faults became active after intrusion of the laccoliths and was accompanied by nonmarine sedimentation of the Et-then Group. Although some of the old fault lines, active during Sosan sedimentation, were reactivated at this time, there is no evidence of strike-slip movement during Great Slave Supergroup deposition.

The bulk of strike-slip displacement occurred along a fault line that skirts the north shore of Wilson Island and Simpson Island, and the north shore of McDonald Lake, joining the McDonald Fault itself near the Snowdrift River (Fig. 25.8). The famous scarp on the south side of McDonald Lake is a splay fault that extends, west of the lake, through the main outcrop area of the Union Island Group. In common with other splays, it is probably an oblique-slip fault and has relatively little net strike-slip displacement. The only other exposures of Union Island rocks occur in windows through the Et-then Group south of McLean Bay. Matching these with the area west of McDonald Lake requires 70-90 km of strike-slip on the McDonald-Wilson Fault (Fig. 25.1). Displacement of this magnitude nicely resolves otherwise puzzling relations between dissimilar outcrop belts of the Sosan and Wilson Island groups (*see* above), and agrees with the more conservative of two estimates based on the matching of aeromagnetic anomalies northeast of the East Arm (Thomas *et al.*, 1976).

The massive Archean granites exposed along much of the length of the McDonald-Wilson Fault are chloritized but not penetratively deformed. The belt of mid-Aphebian mylonites (Fig. 25.1) is clearly truncated by, and not genetically related to the fault.

The alluvial fan sediments of the Et-then Group were deposited in down-dropped fault blocks during strike-slip movement. Although the conglomeratic Murky Formation is cut by the main fault and its splays, there are great changes in its thickness across the fault lines. Even more impressive is the way sandstones high in the Preble Formation overstep several splay faults, proving that movement on those splays ceased before the end of sedimentation. These relations are especially well shown north of McDonald Lake.

Basic Intrusions of the Mackenzie Swarm and the Christie Bay Gravity Anomaly

These are the only rocks not cut by strike-slip faults. Centred at Christie Bay is a tiered, crudely funnel-shaped, intrusion of gabbro, 125 km in maximum diameter (Fig. 25.8). The intrusion takes the form of anastomosing sills on the north side and inward-dipping dykes elsewhere. Deeply weathered outcrops of a possibly ultramafic phase of the intrusion occur on Pethei Peninsula at latitude 62°37'30", longitude 111°16'. If ultramafic rocks occur more extensively at depth, they may account for the 30 mgal positive gravity anomaly associated with the intrusion (Hornal and Boyd, 1972). The intrusion is cut by northwest-trending dykes of the Mackenzie Swarm.

Regional Tectonics and implications for the Aulacogen Hypothesis

The original reasons for hypothesizing the existence of an aulacogen – the longitudinal paleocurrents, the unusual ubiquity of volcanic rocks, the evidence of early rifting in the form of old faults and dyke swarms, and the stratigraphic correlations with other basins in the foreland of the Coronation Geosyncline – have all been strengthened by the new mapping. Nevertheless, certain newly discovered features may require modification of earlier interpretations of the paleogeography and time of initial rifting of the aulacogen.

Discovery of the nappes led to the realization that the axial trough of the aulacogen originally lay south of the McDonald-Wilson Fault (Fig. 25.3), perhaps coincident with the densest part of the mid-Aphebian dyke swarm. This means that the aulacogen was a somewhat larger structure than previously shown and that the East Arm is merely part of its northern margin. The nappes were presumably triggered by uplift south of the McDonald-Wilson Fault but whether this uplift is genetically related to the aulacogen or a manifestation of the so-called Hudsonian Orogeny that affected the entire Churchill Province is not known. However, medial uplifts of this type seem to be a common feature of old aulacogens, those in the Wichita Aulacogen of Oklahoma (Walper, 1976) being especially well known.

An important aspect of the aulacogen hypothesis is that it be cogenetic with the Coronation Geosyncline. The interpretation of the Wilson Island Group as being the result of a first stage of rifting in the East Arm complicates the question of the origin of the aulacogen. No rocks correlative with the Wilson Island Group are known in the geosyncline and, therefore, cogenecity cannot be argued until the second stage of rifting in the East Arm, that beginning with the mid-Aphebian dyke swarm. Whether the Wilson Island Group is truly part of the aulacogen, or the product of an entirely earlier cycle of rifting that merely provided a zone of crustal weakness later occupied by the aulacogen, is hard to say, especially so as the structural position and age of metamorphism of the Wilson Island Group is uncertain.

The possibility that the diorite-monzonite laccoliths may be related to the Great Bear Batholith has interesting implications. The batholith (*see* Hoffman and McGlynn, in press), is believed to have been generated above an east-dipping subduction zone at the old continental margin. Are the laccoliths what happens when a subduction zone passes beneath an aulacogen?

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NEW DATA ON CORRELATION OF THE LITTLE DAL FORMATION AND A REVISION OF PROTEROZOIC MAP-UNIT 'H5'

Project 730057

J.D. Aitken
Institute of Sedimentary and Petroleum Geology, Calgary

Prior to 1976, a widespread, thick, and economically important stratigraphic interval in the Proterozoic succession in Mount Eduni, Bonnet Plume Lake, Upper Ramparts River and Sans Sault Rapids map-areas, District of Mackenzie (106 A, B, G, H) was dealt with as "Helikian(?) map-unit H5" in a number of Geological Survey of Canada reports (Aitken and Cook, 1974a, b, 1975). Knowledge of the interval was fragmentary, and important implications of the summary by Aitken *et al.* (1973) remained undocumented. For these reasons, the writer concentrated his efforts during the short 1976 field season (July 15-August 15) on the H5 map-unit.

The principal findings of this study are that:

a. Map-unit H5 is a lithostratigraphic unit of group rank about 1350 m (4400 ft.) thick, divisible into five or more widespread formations.
b. Each of the subunits comprising the map-unit is a regional, "blanket" formation, instead of being characterized by variability in the sense of lateral facies changes, as previously emphasized.
c. Contrary to previous speculations, the type Little Dal Formation is not equivalent to any part of unit H5, but is instead another "blanket" formation overlying unit H5.

The reasons for the published errors in regard to the interpretation of map-unit H5 (aside from the scanty observational base) are to be found in, first, the regional, low-angle unconformities beneath the Hadrynian Rapitan Group and the Cambro-Ordovician Franklin Mountain Formation, which result in those units overlying different parts of the unit H5-Little Dal succession from place to place and, second, the fact that the lower three formations of unit H5 are confined to the country north of Plateau Fault, whereas the upper two formations, with minor exceptions, occur only in the Plateau Thrust Sheet itself. Those same circumstances probably account for a miscorrelation of the type Little Dal Formation in Wrigley Lake and Glacier Lake map-areas by Gabrielse *et al.* (1973).

Stratigraphy of map-unit H5

Map-unit H5, with an aggregate thickness of at least 1350 m (4400 ft.), conformably overlies the Katherine Group and is overlain conformably by the Little Dal Formation, *sensu stricto*, that is, by the feature-forming carbonate formation mapped as Little Dal Formation in the immediate hanging wall of Plateau Fault in Wrigley Lake (Gabrielse *et al.*, 1973), Mount Eduni and Bonnet Plume Lake map-areas (Aitken and Cook, 1974a, b). Unit H5 is readily divisible into five formations that are practicable map-units at a scale of 1: 125 000 (some "lumping" might be advisable at 1: 250 000 scale); these are named here, informally and temporarily, in upward succession, the "mudcracked subunit", the "basinal sequence", the "grainstone subunit", the "gypsum subunit", and the "rusty shale subunit".

The following descriptions are based mainly on a cluster of measured sections centred at Latitude 64°47'N, Longitude 129°45'W in Mount Eduni map-area ("A", Fig. 26.1), but the formations were studied and mapped from the south boundary of Mount Eduni map-area to the west boundaries of Bonnet Plume Lake and Upper Ramparts River map-areas.

"Mudcracked subunit"

The basal, "mudcracked subunit" of unit H5, 40 to 60 m (130-200 ft.) thick, is less resistant than the uppermost Katherine Group quartzites, upon which it lies conformably, but more resistant than the lower part of the overlying "basinal sequence" (Fig. 26.2). It consists primarily of sandstones and mudrocks in a coarsening-upward, shallowing-upward sequence. The shale and mudstone are mainly dark grey, shedding black talus. The sandstone is mainly pale grey, very fine and fine grained; rip-up breccias of shale chips are common at the bases of the sandstone units. A few medium and thick beds of brown-weathering, dolomitized grainstone, including oolites, occur in many sections. The subunit is named for the abundance of mudcracks, which increase in abundance and coarseness upward.

"Basinal sequence"

The "basinal sequence" conformably overlying the "mudcracked subunit" comprises several different lithologies, each of which may replace the other laterally (Figs. 26.2, 26.3). It is 250 to 400 m (820-1310 ft.) thick. The sequence is characterized by limestone nodules and nodular beds that are clearly of diagenetic origin. The most striking manifestation of this lithofacies is the weakly resistant "Dead End Shale" (obsolete) of Nauss (*see* Aitken *et al.*, 1973, p. 16), brick red to deep brownish red, calcareous laminated shale with abundant ellipsoidal nodules and nodular thin beds of red to pinkish grey, microcrystalline limestone. The "Dead End Shale" is not, however, a useful stratigraphic unit, because the red facies passes at many places both laterally and vertically into beds with grey limestone nodules in dark grey shale; furthermore, more than one red interval occurs in some sections. Resistant intervals of a rhythmite lithofacies (many of the bodies

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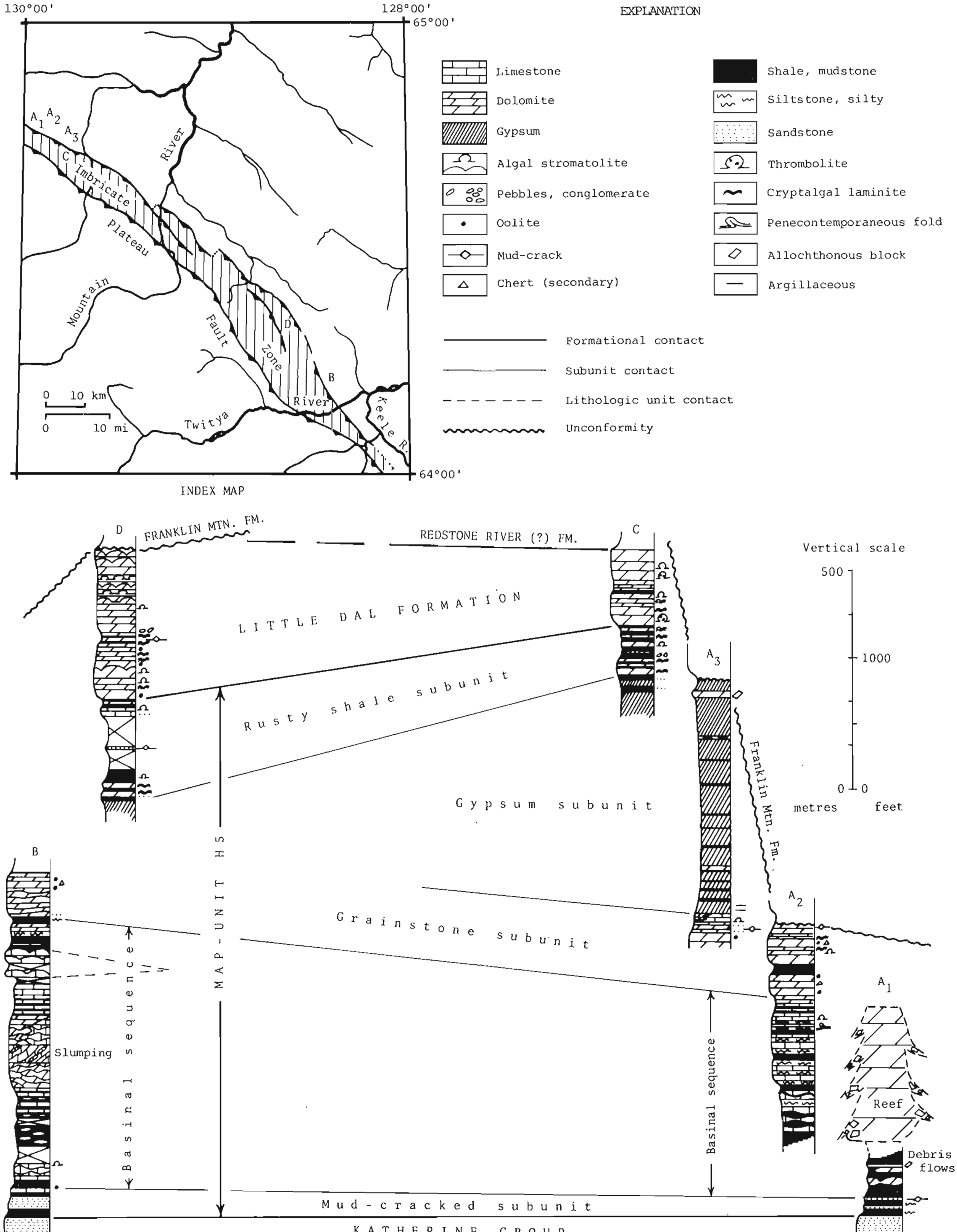

Figure 26.1. Correlation of map-unit H5 and Little Dal Formation, on lines A-B and C-D.

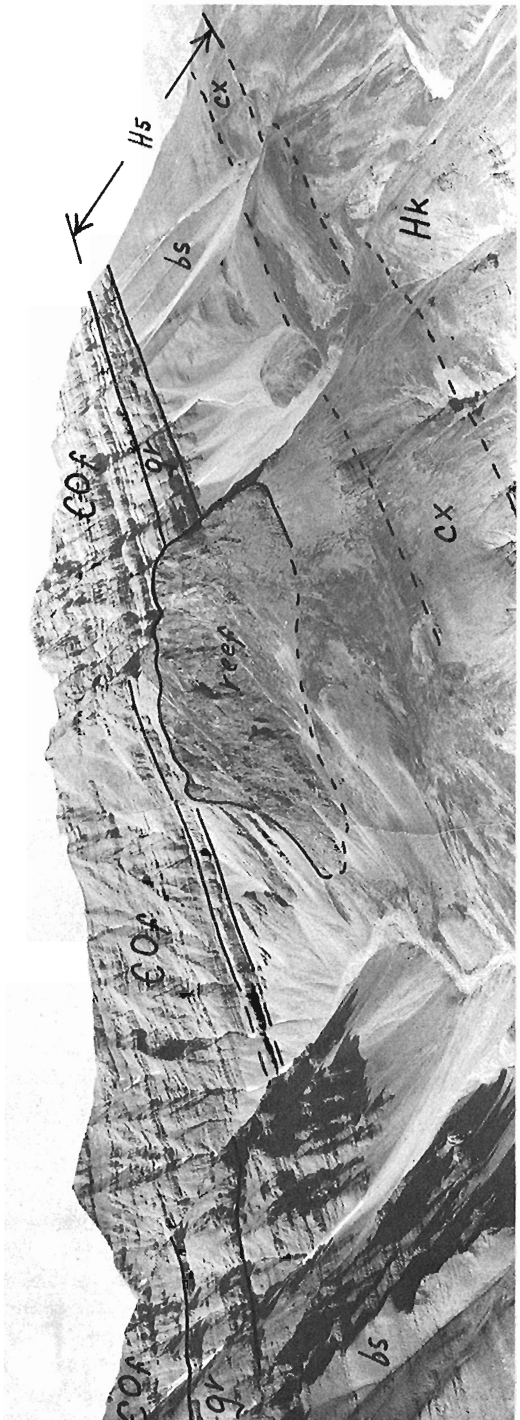

Figure 26.2. Succession Katherine Group (Hk)-map-unit H5-Franklin Mountain Formation (ЄOf), localities A_1 and A_2 (*see* Fig. 26.1). Note the contrast in the resistance of the "basinal sequence" (bs) between left foreground and right background; this is due to the presence of thick rhythmite intervals at the former, and their absence at the latter. The stromatolitic reef has its base low in the "basinal sequence" and its top just below the "grainstone subunit" (gr). Note the steeply dipping flank beds. cx – "mud cracked subunit".

mapped as "H5 carbonates" in manuscript maps placed on Open File) interrupt the nodular facies, and locally can be seen to form tongues passing into nodular facies. The rhythmites consist of lenticular to nodular, but uniformly thin beds of silty microcrystalline limestone separated by thinner beds of grey to dark grey, carbonaceous, calcareous silty shale or shaly siltstone. The rhythmites are locally dolomitized, especially at the top of the "basinal sequence". At and near the base of the "basinal sequence", dark grey to black shale is prominent, as are intervals comprising medium and thick, nodular beds of microcrystalline limestone, locally dolomitized; "molar-tooth" structure is present locally. Thin intervals of gravity-slumped and penecontemporaneously folded beds, and of debris flows, occur here and there in the sequence.

The "basinal sequence" is host to spectacular carbonate reefs (Figs. 26.2, 26.3), some nearly 300 m (1000 ft.) high. The reef-cores are built of a variety of cryptalgal stromatolites, and are enveloped in a thick mantle of coarse reef-derived talus. The reefs at the head of Gayna River are mostly limestone, while those in the Mountain River drainage are mainly dolomitized, with variable, locally excellent, preservation of depositional textures. The reef belt is incompletely defined, but appears to trend slightly more southerly than the tectonic trend. At the head of Gayna River, it wholly occupies the first anticline northeast of Plateau Fault but, southeastward, is confined increasingly to the south limb, and at Mountain River has passed from sight, presumably down-dip in the south limb.

Apart from the belt of reefs, the only significant change noted in the "basinal sequence" is the appearance, at the most southerly section studied ("B", Fig. 26.1), of a resistant member of stromatolitic grey limestone (locally, orange-weathering dolomite) near the top. This is Unit 6 of the sequence incorrectly assigned to the Little Dal Formation, at Section U-10, Deca Creek (Aitken *et al.*, 1973, p. 93). It probably correlates with stromatolitic dolomite and limestone also mistakenly assigned to the Little Dal at Little Bear River (Sec. U-13, *ibid.*, p. 110, 111).

"Grainstone subunit"

The "basinal sequence" passes gradationally into the overlying, feature-forming "grainstone subunit" (Figs. 26.2, 26.3), 230 to 270 m (750-890 ft.) thick, and host to the more important zinc-lead showings at the RT property at Gayna River. This subunit is characterized by an abundance of dolomitized grainstones, notably ooid grainstones, that are mostly thick bedded. Preservation of depositional texture is variable and commonly poor, but scattered nodules of grey and white chert have preserved the original limestone textures perfectly. Other lithologies in the "grainstone subunit" include beds interpreted as dolomitized calcisiltite, biostromes of cryptalgal stromatolites, "molar-tooth" limestone and dolomite, and minor intervals and interbeds of grey shale. Included with the "grainstone subunit" is an upper member of platy to flaggy, yellow-weathering, microcrystalline, partly sandy dolomite with large mudcracks that forms a useful marker interval. The platy dolomites include minor simple stromatolites and, at the top, salt-crystal casts.

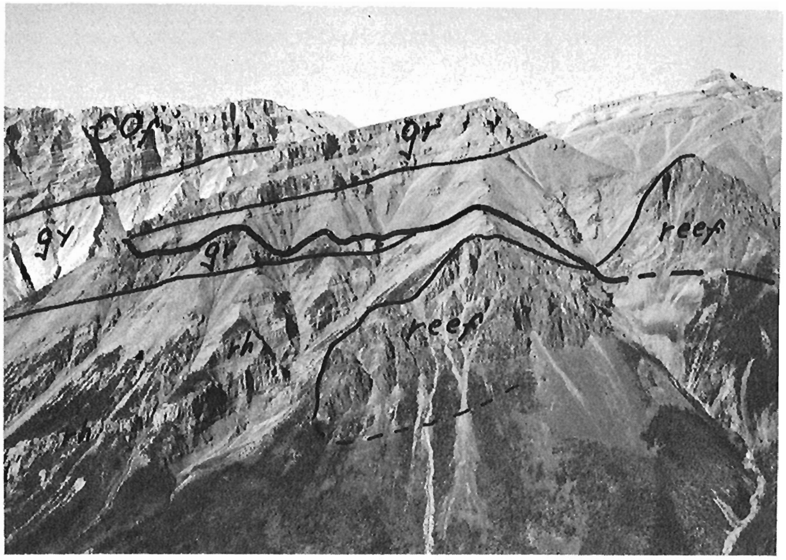

Figure 26.3

Map-unit H5 at locality A_3 (*see* Fig. 26.1). The "grainstone subunit" (gr) dips beneath the "gypsum subunit" (gy) in left background. Note the low-angle unconformity beneath the Franklin Mountain Formation (ЄOf). The two reef-masses are the faulted halves of a single reef. rh — rhythmite intervals in the "basinal sequence".

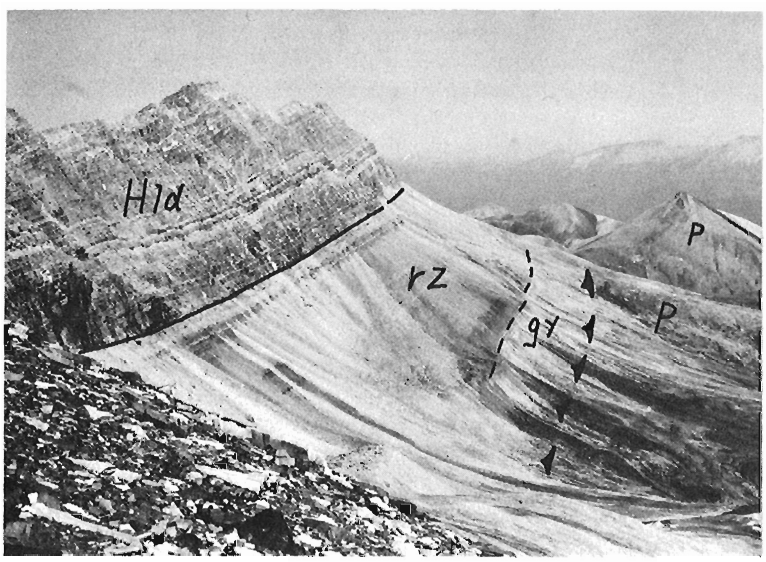

Figure 26.4

Map-unit H5 and Little Dal Formation (*sensu stricto*) at locality D (*see* Fig. 26.1).

gy — "gypsum subunit" (mostly covered by talus);
rz — "rusty shale subunit";
Hld — Little Dal Formation;
P — Paleozoic rocks in footwall of unnamed thrust fault.

"Gypsum subunit"

The "grainstone subunit" is conformably succeeded by the "gypsum subunit" (Figs. 26.3, 26.4), consisting almost entirely of gypsum that is white or near-white, except for two or three red intervals and a zone at the base that is black and shaly. The only non-sulphate beds are rare, very thin beds and laminae of grey gypsiferous shale and, at the top of the subunit, two or three thin beds of red, gypsiferous sandstone. Laminated, nodular, and enterolithic fabrics alternate throughout the subunit which, in view of the total absence of such shallow-water indicators as mudcracks and ripple marks and the paucity of detrital material, is probably an example of an evaporite body of subtidal origin.

The "gypsum subunit" is 531.8 m (1740 ft.) thick at the single section measured to date ("A_3", Fig. 26.1). This is less than the depositional thickness, because the subunit is seen to be truncated by the sub-Franklin Mountain unconformity at the measured section (Fig. 26.3). It is worth emphasizing here that the "gypsum subunit" is a regional, "blanket" deposit that everywhere overlies the "grainstone subunit" and underlies the "rusty shale subunit" (barring erosional removal of the latter). Previous speculations (e.g. Aitken *et al.*, 1973) regarding a facies change of map-unit H5 to a gypsiferous facies are unfounded.

"Rusty shale subunit"

Fifth and highest of the formational subdivisions of the unit H5 "group" is the "rusty shale subunit", 110 to 230 m (335-700 ft.) thick. It succeeds the gypsum with abrupt gradation and is intermediate in resistance between the "gypsum subunit" below and the Little Dal Formation (*sensu stricto*) above (Fig. 26.4). The formation consists largely of weakly to highly fissile shale that is grey, greenish grey and black, with one or two purple intervals. Next in abundance is dolomite, or less commonly relict limestone. Cryptalgal laminates are prominent near the base, where they are associated with a few beds of intraclast grainstone or wackestone and mat-chip breccia. Higher, one or two cryptalgal stromatolite biostromes or horizons of bioherms occur. Minor in amount, but characteristic of the subunit, are beds of red and white, very fine grained sandstone with well developed ripple marks and distinctive, crescentic to vermiform mudcracks.

The unit gives the impression of being notably iron-rich. Pyrite is prominent in all lithologies, the stromatolitic dolomites weather to a deep rusty shade, and nearly all of the shales display pronounced rusty weathering.

The "rusty shales subunit" is overlain conformably, at an interbedded contact, by massive carbonates at the base of the Little Dal Formation (*sensu stricto*, *see* below).

Corollaries of the new stratigraphic data

The new data have important implications with regard to geological maps of the region, both those formally published and those released on Open File in manuscript form.

First, it should be noted that the type section of the Little Dal Formation is in the Plateau Thrust Sheet near Coates (formerly Little Dal) Lake in Glacier Lake map-area (Gabrielse *et al.*, 1973, p. 16 and Map 1314A). Second, with minor exceptions, the stratigraphic succession above Plateau Fault in Mount Eduni and Bonnet Plume map-areas is: "gypsum subunit" (H5), overlain by "rusty shale subunit" (H5), overlain by Little Dal Formation in mapped continuity with the type section. It is the writer's opinion, subject to confirmation by examination of exposures in southern Wrigley Lake and northern Glacier Lake map-areas, that the type Little Dal overlies the "rusty shale subunit" of map-unit H5 or its equivalent. If this is correct, then rocks mapped as Little Dal Formation over wide areas, being pre-gypsum, are incorrectly mapped, and are in fact feature-forming carbonate units of map-unit H5.

In earlier mapping (Aitken and Cook, 1974a, b, 1975) the writer repeated an error apparently made by Gabrielse *et al.* (1973) when limestones on the southwest side of Tigonankweine Range were "tentatively included in the Little Dal Formation" (*ibid.*). Two totally distinct, feature-forming carbonate intervals (Little Dal Formation *sensu stricto* and the "grainstone subunit" and/or rhythmite intervals in the "basinal sequence"), separated by more than 700 m (2297 ft.) of mainly non-carbonate strata (Fig. 26.1), were assumed to be the same formation, although in detail they do not resemble one another closely. The Helikian succession on the southwest flank of Tigonankweine Range now is known not to contain any stratigraphic unit equivalent to the Little Dal Formation of the Plateau Thrust Sheet.

The following errata in published and Open File maps should be noted:

1. The Little Dal Formation (*sensu stricto*) does not occur in Carcajou Canyon map-area. The feature-forming carbonates so mapped are the "grainstone subunit" of map-unit H5 and/or intervals of feature-forming rhythmites or stromatolitic carbonates in the "basinal sequence".
2. In Mount Eduni and Bonnet Plume Lake map-areas, the Little Dal (*sensu stricto*) occurs northeast of the Plateau Sheet and the sub-Plateau Fault imbricate zone in one limited area only, namely, in the south limb of the first anticline in front of Plateau Fault, from the head of Gayna River intermittently westward. In this area, the stratigraphic sequence, "grainstone subunit"-"gypsum subunit"-"rusty shale subunit"-Little Dal Formation can be observed directly.
3. In no place (so far as presently known) has Little Dal (*sensu stricto*) been mapped as "H5 carbonates".
4. Four different lithologic units have been shown as mappable carbonate units within map-unit H5, namely:
    a. Intervals of feature-forming rhythmites in the "basinal sequence".
    b. Stromatolitic reefs in the "basinal sequence".
    c. A prominent stromatolitic member within the "basinal sequence".
    d. The "grainstone subunit".

A group name for the interval now treated as "map-unit H5", and formational names for each of the five subunits described here will be formally proposed in a forthcoming publication.

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27. REDSTONE RIVER FORMATION (UPPER PROTEROZOIC) IN MOUNT EDUNI AND BONNET PLUME LAKE MAP-AREAS, DISTRICT OF MACKENZIE

Project 730057

J.D. Aitken
Institute of Sedimentary and Petroleum Geology, Calgary

Field work in Mount Eduni and Bonnet Plume Lake map-areas (106A, B) was carried out in 1969 and 1970, prior to publication of complete descriptions of the Redstone River and Coppercap formations or the underlying Little Dal Formation (Gabrielse *et al.*, 1973). Manuscript maps of the two areas, released on Open File (Aitken and Cook, 1974), do not record the presence of the Redstone River Formation, and there has been concern that this important, commonly copper-bearing unit had been overlooked.

In 1976, in the course of field work directed mainly to other questions, the writer, with the kind assistance and co-operation of G. Eisbacher, undertook a rapid check of structural panels containing the Little Dal Formation (*sensu stricto*; *see* Aitken, 1977) with the specific objective of determining the presence or absence of Redstone River and Coppercap Formations above the Little Dal. The possible economic significance of the findings dictates publication of this short note.

Both Redstone River and Coppercap were indeed found, as shown in Figure 27.1 (to be used in conjunction with the maps on Open File). With a single, tentative exception, both formations outcrop only along the leading edge of the Plateau Thrust Sheet. Their distribution is discontinuous; this is due largely to removal at the sub-Rapitan erosion surface, which was bevelled across a faulted terrain, but is due also in part to distinct depositional lenticularity of the Redstone River and Coppercap formations.

Recognition of the Redstone River Formation was based on:

a. Position above complete sections of Little Dal Formation, *sensu stricto*.

b. Presence of poorly exposed, pink, calcareous siltstone and sandstone.

c. Presence of gypsum, or collapse-breccia related to gypsum.

d. Position (commonly in a poorly exposed interval) beneath recognizable Coppercap Formation.

e. Presence of coarse, reddish conglomerate.

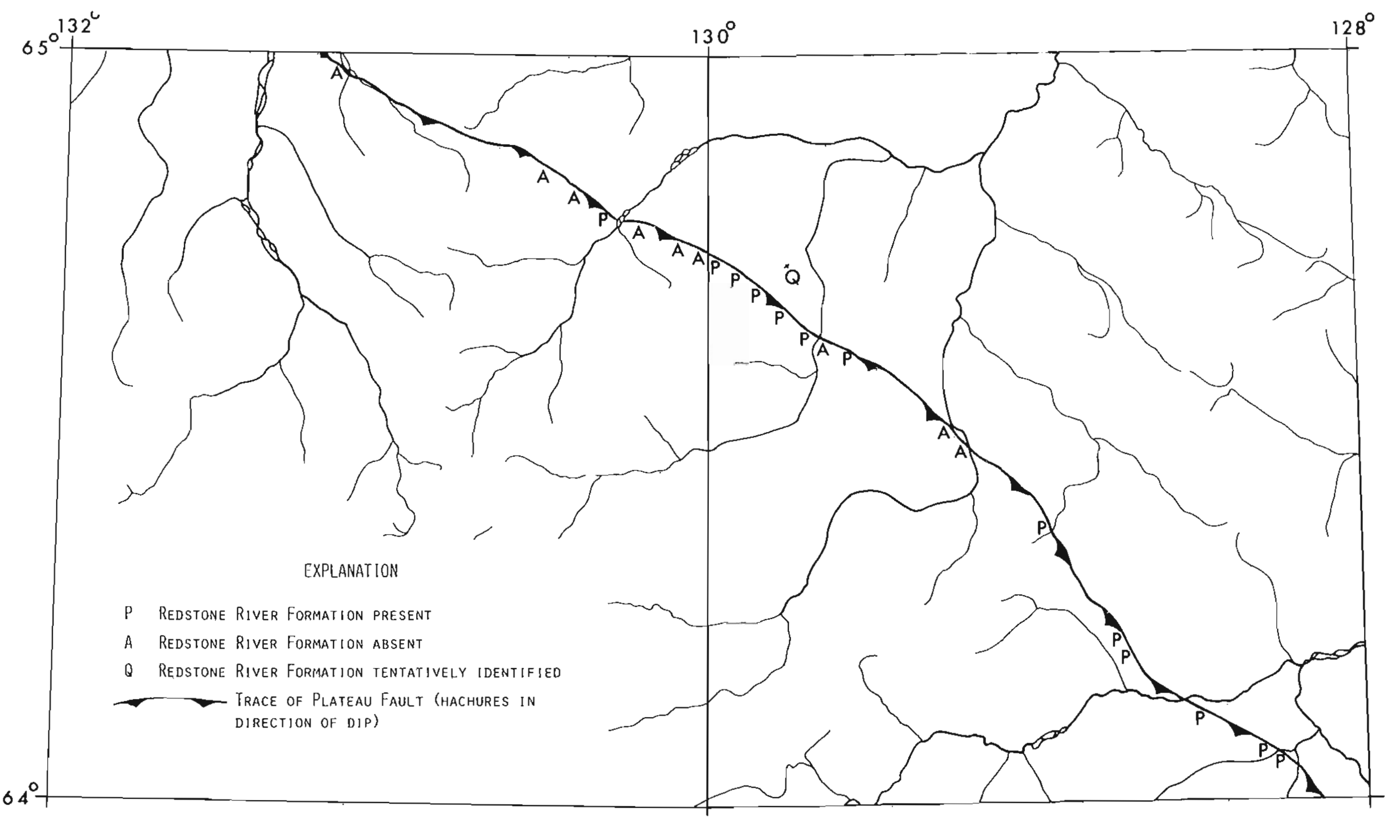

Figure 27.1. Observations of the presence or absence of Redstone River Formation in Mount Eduni and Bonnet Plume Lake map-areas.

From: Report of Activities, Part A;
Geol. Surv. Can., Paper 77-1A (1977)

Recognition of the Coppercap Formation was based on:

a. Position relative to Redstone River and Little Dal formations.
b. A distinctive cycle beginning with black laminites (limestone and/or dolomite) at the base, associated with flat-pebble breccias, grading upward into pale-coloured, resistant dolomites that are cryptalgal laminites.

The single, tentative occurrence of the Redstone River Formation outside the Plateau Thrust Sheet ("Q", Fig. 27.1) is only 25.5 m (84 ft.) thick. The interval is assigned tentatively to the Redstone River on the basis of an association of coarse conglomerate, similar to that of a Redstone River section in the Plateau Sheet nearby, with red shales.

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PERMIAN CONODONTS IN THE CANADIAN ARCTIC ARCHIPELAGO – BIOSTRATIGRAPHIC DISCUSSION

Project 680064

H. Kozur[1] and W.W. Nassichuk
Institute of Sedimentary and Petroleum Geology, Calgary

Introduction

Biostratigraphic and chronostratigraphic relationships of Permian rocks in the Sverdrup Basin have been interpreted, in the main, through studies of fusulinaceans, ammonoids and brachiopods. Fusulinaceans are particularly abundant in strata of Asselian to early Artinskian ages which represent an interval of widespread carbonate deposition in the basin, but they are absent from all younger Permian rocks in the basin; that is, these ranging from middle Artinskian to Kazanian. Rocks of the latter ages are predominantly sandstones with minor limestones and they contain abundant brachiopods and fewer ammonoids, all indicative of a Boreal Province. Little is known of the distribution and diversity of Permian conodonts within the Sverdrup Basin but Bender (1973) reported the presence of well-preserved but unidentified representatives of the group from several Lower Permian localities in Ellesmere Island and Axel Heiberg Island. Similarly, Uyeno (pers. comm., 1968) identified a single specimen of *Gondolella* sp. from Upper Permian strata near M'Clintock Inlet, northern Ellesmere Island. In the present study, single samples from the Assistance Formation (Kungurian) near the eastern edge of the Sverdrup Basin and from the lowest part of the Degerböls Formation (Kungurian-Kazanian) in the interior of the basin have yielded several species of the cosmopolitan *Gondolella* Stauffer and Plummer which can be identified readily by the outline and sculpture of the platform. The purpose of this report is to compare representatives of *Gondolella* from the Assistance and Degerböls formations with species that occur in nearly contemporaneous strata in the Phosphoria Formation in Idaho and Wyoming and the Road Canyon Formation in Texas. Conodonts are documented insufficiently from the type Permian near the Lower-Upper boundary in the Ural Mountains and, thus, age relationships cannot be compared directly. Conodonts from the Assistance Formation (Boreal), however, appear to be considerably younger than those from the lowermost Kungurian. They resemble species from Chihsian or Kubergandinian strata (Tethyan) in the Pamir Mountains in the southern Soviet Union (Tadzik SSR) directly north of Afghanistan. According to Leven (1967), the boundary between the Lower Permian and Upper Permian series lies above the Chihsian and below the Kubergandinian.

Assistance Formation and equivalent strata - biostratigraphic review

The Assistance Formation was defined by Harker and Thorsteinsson (1960) for a succession of sandstones, siltstones and minor limestones at Grinnell Peninsula, Devon Island near the southern margin of the Sverdrup Basin. Complementary information concerning relationships of the Assistance Formation was provided by Thorsteinsson (1974) and Nassichuk (1975). The formation extends discontinuously along the southern margin of the basin where, generally, it is less than 60 m thick. Along the eastern margin of the basin the formation contains a greater proportion of limestone and thicknesses may exceed 300 m. In the type area, the Assistance Formation overlies the Belcher Channel Formation (Asselian-Artinskian) which contains skeletal limestone and minor sandstone. Locally, on parts of Melville Island and Ellesmere Island, a deltaic succession of quartzose sandstones, the Sabine Bay Formation (Artinskian), occurs between the Belcher Channel and Assistance formations. Basinward, the Assistance Formation grades to the van Hauen Formation (Artinskian-Kungurian) which contains dark grey shales, siltstones and bedded chert. The van Hauen is overlain by the Degerböls Formation but, locally, where the van Hauen is absent, the lower part of the Degerböls may be equivalent to the van Hauen Formation.

The Assistance Formation is replete with a great variety of fossils, some of which have been described in the literature; included are brachiopods (Harker and Thorsteinsson, 1960; Waterhouse, 1969, 1971; Sarytcheva and Waterhouse, 1972), ammonoids (Nassichuk *et al.*, 1966; Nassichuk, 1970), scaphopods (Nassichuk and Hodgkinson, in press), nautiloids, gastropods, conularids, foraminifers and plants. A single fragmentary conodont specimen, *Gondolella* sp. indet. has been found in the type section of the Assistance Formation but the state of preservation of the specimen precludes more than simple documentation of the occurrence.

The age of the Assistance Formation is latest Early Permian (*sensu* Furnish, 1973) but correlation with successions in the type area for the Permian remains obscure. In the latter area, that is, in the Ural Mountains, uppermost Lower Permian strata are assigned to the Kungurian Stage and lowermost Upper Permian to the Ufimian Stage. Stratotypes for the Kungurian and Ulfimian Stages contain beds of lagoonal and continental origin with only few strata reflecting open-marine conditions between the western edge of Priurals and the eastern edge of the Russian Platform. With the exception of ostracodes and brachiopods at particular levels, both stratotypes lack distinctive marine faunas suitable for intercontinental correlation over much of their extent; ostracode zonation was summarized by Gorsky and Guseva (1973) and brachiopod zonation by Waterhouse (1976).

Various suggestions have been made in the literature to combine the Kungurian and Ufimian stages into a single stage (Stepanov, 1967; Gorsky, and Guseva, 1973)

[1]Democratic Republic Germany, Meiningen, Staatliche Museen

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but the paucity of marine faunas, particularly in the upper half, reduces effectiveness of the interval as a standard for correlation. Waterhouse (1976) favours combining Kungurian and Ufimian into a single stage for which the name "Kungurian" is retained and, on the basis of brachiopods, which may be identified on a global basis, recognizes two substages; a lower substage (Fillipovian) and an upper substage (Irenian). In the Waterhouse (*ibid.*) scheme, Ufimian beds are contained within the upper Irenian. Nassichuk *et al.* (1966) examined Permian ammonoids from Arctic Canada, including *Daubichites fortieri* (Harker) and *Sverdrupites harkeri* (Ruzhencev) from the type section of the Assistance Formation and concluded that the Assistance probably was equivalent to Kungurian strata in the Urals. Similarly, these authors related the Assistance Formation to the Leonardian Road Canyon and Brushy Canyon Formations in West Texas and to the Meade Peak Member of the Phosphoria Formation in Idaho. According to Pnev (pers. comm., 1975) and others, numerous Soviet stratigraphers hold the opinion that much of the Kungurian in the Urals may be equivalent to "Artinskian" outside of the Urals.

Ammonoids are exceedingly rare in Kungurian (*sensu stricto*) strata in the Urals and only three occurrences are known from the middle part of the stage; included are *Paragastrioceras*, *Uraloceras* and ?*Daubichites*. According to Bogoslovskaya (pers. comm., 1975), Kungurian ammonoids can be distinguished from Artinskian ammonoids only at the species level. Nassichuk (1970) suggested that certain ammonoids from typical Assistance rocks show affinities with Roadian species from the Phosphoria Formation (*Daubichites fortieri* and *Sverdrupites harkeri*), whereas others show affinities with slightly older Leonardian species [*Synartinskia belcheri* Nassichuk, *Medlicottia* aff. *M. orbignyana* (Verneuil) and *Popanoceras* cf. *P. sobolewskyanum* (Verneuil)]. Waterhouse (1976) suggested that brachiopods from the type Assistance Formation are early Kungurian (Fillipovian) in age and can be distinguished clearly from older Artinskian species. Brachiopods from the type section of the Assistance Formation include: *Arctitreta pearyi* (Whitfield), *Neochonetes* spp., *Thuleproductus arcticum* (Whitfield), *Muirwoodia mammatus* (Keys.), *Pseudosyrinx* spp., *Spiriferella loveni* (Diener) and *Neospirifer striatoplicatus* (Gobbett).

In the Pamir Mountains, as is the case in the Ural Mountains, Artinskian strata can be identified on the basis of ammonoids and fusulinaceans. Unlike the Urals succession, however, where Artinskian strata are overlain by sparsely fossiliferous lagoonal and continental deposits with only few limestone beds, post-Artinskian strata in the Pamirs, consisting mainly of limestones, contain an abundant and varied Tethyan fusulinacean fauna (Leven, 1967). Leven (*ibid.*) placed the Lower-Upper Permian boundary above the fusulinacean Zone of *Misellina* and below the Zone of *Cancellina*. The *Misellina* and *Cancellina* Zones correspond to the Chihsian and Kuberganinian Stages, respectively (Leven, 1975).

Conodonts from GSC locality C-32487 - Assistance Formation

GSC locality C-32487 occurs in the Assistance Formation on Hamilton Peninsula, north of Canon Fiord, west-central Ellesmere Island (Lat. 80°02'N, Long. 81°44'W) (Fig. 28.1). Conodonts were recovered from a large talus block situated 40 m above the base of the formation; the talus block originated from a prominent ledge 43 m above the base of the formation. The same horizon yielded the Roadian ammonoid *Daubichites fortieri* which also occurs in the type section of the Assistance Formation. In the vicinity of locality C-32487, the Assistance Formation is 110 m thick and contains recessive, glauconitic and quartzose sandstones and an equal proportion of more resistant thin to medium bedded sandy limestone intervals. Sandstone beds contain *Spirophyton* and limestone beds contain abundant brachiopods and bryozoans. The Assistance Formation rests on quartzose sandstones of the Sabine Bay Formation and is overlain by glauconitic sandstones, and lesser amounts of conglomerates and limestones of the Trold Fiord Formation.

Residues from 10 kg of limestone yielded 6 conodont specimens. Included are well-preserved specimens of *Gondolella idahoensis* Youngquist, Hawley and Miller, 1951, as well as 3 fragments which are determined to be morphologically intermediate between *G. idahoensis* and *G. nankingensis* Cheng, 1960 or *G. serrata* Clark and Ethington, 1962. One of the latter fragments has a serrate margin and, therefore, more closely resembles *G. nankingensis* or *G. serrata* than it does *G. idahoensis*. A second fragment consists of a small middle portion of platform; the margin is non-serrate but shallow transverse ribs are developed. Since none of the fragmentary serrate or weakly ribbed specimens shows posterior regions, it cannot be determined whether they are more closely related to *G. nankingensis* or to *G. seratta.*

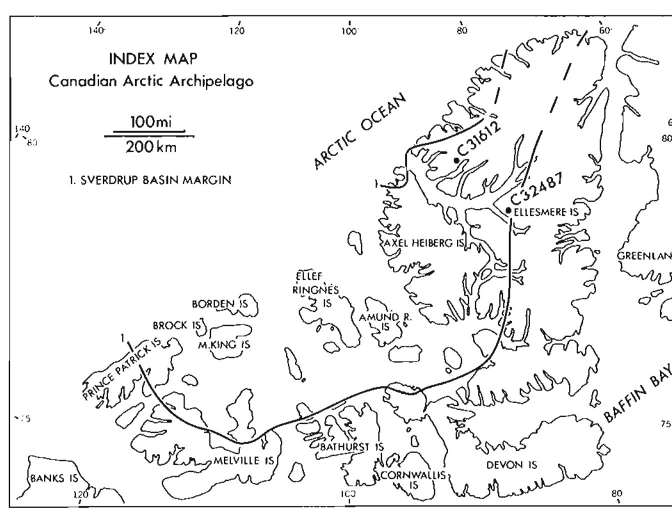

Figure 28.1. Index map showing conodont localities C-32487 (Assistance Formation) and C-31612 (Degerböls Formation).

In stratigraphic sections in Pamir, *Gondolella idahoensis* is the most abundant conodont in the uppermost Lower Permian *Misellina* Zone (Chihsian Stage). In Leonardian beds below the *Misellina* Zone, that is, in the upper part of the ammonoid-bearing "Buz-tere" beds, highly developed representatives of *Gondolella bisselli* Clark and Behnken, 1971 occur. Many of the latter are in fact morphologically intermediate between *G. bisselli* and *G. idahoensis; Gondolella bisselli sensu stricto* generally ranges from Sakmarina to lower Kungurian in the Soviet Union. In Spitzbergen, *G. bisselli* occurs in lower Svalbardian (=lower Kungurian) strata in the Spirifer Limestone (Malkowski and Szaniawski, in press). Thus, conodonts from GSC locality C-32487 appear to postdate Leonardian, lower Kungurian and lower Svalbardian.

The highest occurrence of *Gondolella idahoensis* in North America is in the upper Bone Spring Formation (Roadian). Similar forms are known from the lower and middle Kufeng Formation in southern China. The Meade Peak Member of the Phosphoria Formation in Idaho, Wyoming and the Cutoff Shale in Texas have yielded *Gondolella nankingensis*, an important index for the *Cancellina* Zone (Kurbergandinian). Pertinent discussion of correlation of the Cutoff Shale was provided by Furnish (1973). *Gondolella serrata* makes its first appearance at the same stratigraphic level as *G. nankingensis* but ranges considerably higher, through the Wordian.

In conclusion, the conodont fauna from locality C-32487 can be placed near the boundary between the *Misellina* and *Cancellina* Zones of Asia; which is now the Chihsian-Kubergandinian boundary in Tethyan regions. In terms of an important stratigraphic scheme proposed by Furnish (1973), this fauna can be placed in the Upper Roadian.

Conodonts from GSC locality C-31612 - Degerböls Formation

GSC locality C-31612 occurs 45 m above the base of the Degerböls Formation on the north side of Otto Fiord, northern Ellesmere Island (Lat. 81°17'N, Long. 85°58'W) (Fig. 28.1). In this area, the van Hauen Formation (shales, siltstones, chert) is absent from its typical position beneath the Degerböls Formation and the latter rests directly on shelf-deposited carbonate rocks of the Nansen Formation; uppermost Nansen rocks locally contain fusulinaceans of Sakmarian age and of early Artinskian age on northern Ellesmere Island. The Degerböls Formation has a thickness of 240 m north of Otto Fiord. The lowest 45 m of the formation, including beds containing conodonts in this report, are atypical for the formation and contain reddish-weathering crinoidal limestones and shales; these beds probably are equivalent to the van Hauen and Assistance formations as indicated by conodonts in the following paragraphs. The remainder of the formation contains thin bedded, grey weathering, cherty limestones, all replete with bryozoans, crinoid columnals and brachiopods. Although brachiopods have not been studied yet from the Degerböls Formation north of Otto Fiord, those that occur in the type-section of the formation, 22 km farther to the south of the locality discussed herein, are diverse and show a Kungurian-Kazanian range.

Several fragmentary conodont specimens from GSC locality C-31612 all belong to a single species *Gondolella* cf. *G. gracilis* Clark and Ethington, 1962. The platform is elongate and extremely narrow and tapers gradually toward the anterior end of the carina. The platform has a square posterior end; platform margins are slightly up-turned, and the upper surface is pitted but without transverse ridges. The carina has more than 12 low, somewhat elongated denticles and does not reach the posterior margin. The first 3 denticles at the anterior end of the carina are relatively shorter and broader than the others. The posterior denticle is markedly rounded and pointed. *Gondolella* cf. *G. gracilis* is morphologically intermediate between the more highly developed *G. gracilis* Clark and Ethington from the Meade Peak Member of the Phosphoria Formation in Idaho and Wyoming and the more primitive *G. idahoensis* Youngquist, Hawley and Miller. Whereas the slope of the platform of the Arctic species resembles *G. gracilis*, the denticulation of the carina as well as the form of the posterior end of the platform and the posterior end of the carina more closely resemble *G. idahoensis*. *Gondolella gracilis* characterizes the upper part of the Meade Peak Member of the Phosphoria Formation in Idaho and Wyoming where it is associated with *G. serrata*, and *G. nankingensis* as well as with representatives of *Neostreptognathodus*. As discussed previously, *G. idahoensis* occurs in the Assistance Formation on Ellesmere Island, the Bone Spring Formation in Texas and in the *Misellina* Zone (Chihsian Stage of Leven, 1975) in the Pamirs. No older or younger representatives of *Gondolella* resemble *G.* cf. *G. gracilis*. *Gondolella bitteri* Kozur, 1975 and other Middle and Late Permian species of *Gondolella* are clearly distinguished from *G.* cf. *G. gracilis* by the form or sculpture of the platform. *Gondolella bisselli* Clark and Behnken from Sakmarian, Artinskian and Leonardian strata (*sensu* Furnish, 1973) has a platform that is narrow and elongate compared to the Arctic species but which is more arched and has a different denticulation of the carina.

In conclusion, conodonts from the basal Degerböls Formation (C-31612) indicate correlation with the Meade Peak Member of the Phosphoria Formation. According to Behnken (1975), conodonts from the Meade Peak Member correspond to conodont faunas from the Cutoff Shale and from the basal Brushy Canyon Formation in Texas. In all of those American faunas, association of *Gondolella serrata serrata, G. nankingensis* and *Neostreptognathodus* spp. is characteristic. This fauna seems to be younger than the type Roadian (Road Canyon Formation) but older than Wordian *sensu stricto* (*Waagenoceras* Zone) faunas. According to Kozur (in press), this fauna indicates correlation with the Kubergandinian Stage (*Cancellina* Zone) of Central Asia. The upper part of the Meade Peak Member appears to be younger than the type Roadian (equivalent to the upper *Misellina* Zone or uppermost Lower Permian) and is certainly older than the Wordian. (*Waagenoceras* Zone) and should be correlated, therefore, with the

Cancellina Zone. The lower part of the Meade Peak Member, however, whose conodont fauna was investigated by Youngquist *et al.* (1951) is older than the association with *Gondollela serrata serrata, Gondolella nankingensis* and *Neostreptognathodus* of the *N. clinei* group and can be correlated with the type Roadian (association with *G. idahoensis* and *Neostreptognathodus* of the *N. sulcoplicatus* and *N. leonovae* groups). According to Leven (1975), the *Misellina* Zone and, therefore, also the Roadian should be regarded as Middle Permian. Kozur (in press), however, correlates the type Roadian (Road Canyon Formation) with the upper Chihsian (upper *Misellina* Zone) and regards the whole Chihsian (and, therefore, also the type Roadian) as uppermost Lower Permian. He regards the Kubergandinian (*Cancellina* fauna) as basal Middle Permian. *Gondolella* cf. *G. gracilis* seems to be somewhat more primitive than *Gondolella gracilis* from the conodont fauna with *G. serrata serrata, G. nankingensis* and highly developed *Neostreptognathodus,* but somewhat more highly developed than *G. idahoensis* from the lower Chihsian (Sul-Istyk fauna of Pamir). Therefore, the conodont fauna in the lower Degerböls should be placed near the Chihsian-Kubergandinian boundary and therefore near the Lower-Middle Permian boundary.

Additional conodont studies are required to determine whether the Degerböls fauna with *Gondolella* cf. *G. gracilis* belongs to the type Roadian (Upper Chihsian, uppermost Lower Permian) or to the Kubergandian (basal Middle Permian). At the present stage of our knowledge, it belongs to the upper type Roadian and, therefore, to the highest Lower Permian.

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29. OCCURRENCE OF *PALAEOAPLYSINA* (HYDROZOAN?) IN UPPER CARBONIFEROUS (LOWER MOSCOVIAN) CARBONATE ROCKS, NORTHEASTERN BRITISH COLUMBIA

Project 710022

R.W. Macqueen[1] and E.W. Bamber
Institute of Sedimentary and Petroleum Geology, Calgary

Palaeoaplysina Krotov is a distinctive calcareous fossil which is an important constituent of carbonate mounds within Upper Carboniferous and Lower Permian rocks of the west-central Soviet Union (Ural Mountains and Donetz Basin); northern Canada (Richardson Mountains of the Yukon Territory, and Ellesmere and Axel Heiberg islands of the Northwest Territories); and east-central Idaho of the United States of America (Davies, 1971; Davies and Nassichuk, 1973; Breuninger, 1976; and others). Although it resembles certain stromatoporoids and calcareous phylloid algae, *Palaeoaplysina* is referred tentatively to the Class Hydrozoa because of its growth form, internal canal system, and mamelon-like surface protuberences (e.g. Davies and Nassichuk, 1973).

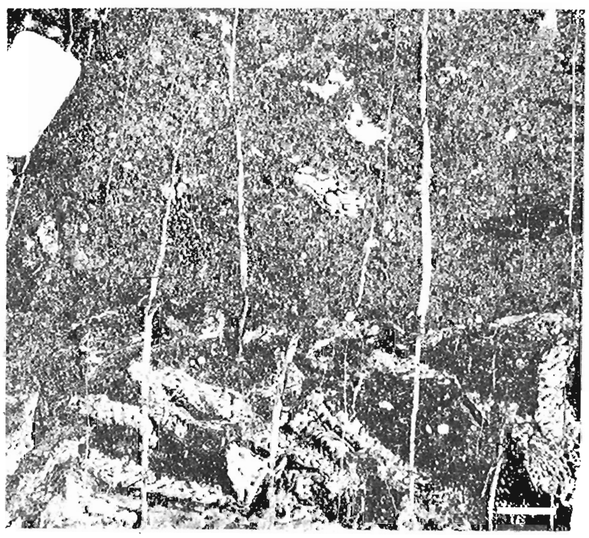

Figure 29.1

Cluster of subhorizontal plates or plate fragments of *Palaeoaplysina* sp. within skeletal packstone, overlain by fine grained skeletal packstone-wackestone with single *Palaeoaplysina* fragment. Bar scale = 0.5 cm. Thin section, plane light (x1.6), unnamed formation near Mount Hanington, northeastern British Columbia (Lat. 54°07'N; Long. 120°12.5'W). GSC loc C-44563.

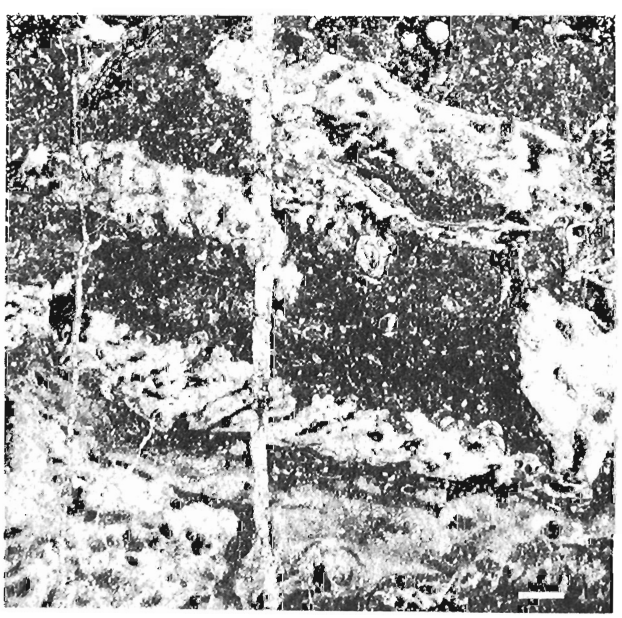

Figure 29.2

Enlargement from Figure 29.1, showing several plate fragments (x6). Note lime-mud filled canals (dark grey) enveloped by sparry calcite (white), which probably occupies space formerly filled by cellular tissue. Bar scale = 1 mm.

[1]Department of Earth Sciences, University of Waterloo, Waterloo, Ontario N2L 3G1.

From: Report of Activities, Part A;
Geol. Surv. Can., Paper 77-1A (1977).

In this paper, a new Canadian occurrence is reported in an unnamed Upper Carboniferous (lower Moscovian) carbonate unit in northeastern British Columbia, approximately 120 miles (192 km) northwest of Jasper, Alberta, in the Rocky Mountain Foothills within the southeastern part of the Monkman Pass, 1: 250 000 map-sheet (93 I), at Latitude 54°07'N, Longitude 120°12.5'W. This unnamed unit, which is 17.5 feet (≈5.3 m) thick, consists of partly dolomitized skeletal packstone and wackestone. It is separated by disconformities from underlying Lower Carboniferous (Viséan) carbonates of the Rundle Group and overlying Lower Permian (Upper Wolfcampian or Lower Leonardian) carbonates of the Belcourt Formation. Ages of the unnamed unit and the overlying Belcourt Formation are based mainly on identifications of fusulinacean foraminifers, including the genus *Profusulinella*, by C.A. Ross (Ross and Bamber, in press); the upper Paleozoic succession of the area is discussed by Bamber and Macqueen (in press). The unnamed unit occurs sporadically within the area as erosional remnants beneath a widespread sub-Permian unconformity (*ibid*).

Palaeoaplysina from the northeastern British Columbia locality occurs as individual plate fragments "floating" in a matrix of skeletal wackestone or less commonly packstone, or as clusters of subhorizontal plates or plate fragments (Figs. 29.1, 29.2). The host wackestone or packstone contains abundant lower Moscovian fusulinacean foraminifers, and has ostracodes, molluscan detritus, minor echinoderm detritus, scattered calcispheres, and abundant pelletoid grains (or peloids) and lime-mud. These sediments appear to represent shallow-water, muddy, lagoonal environments close to sea level, within which local skeletal sands, rugose corals, and *Palaeoaplysina*-bearing banks developed.

Palaeoaplysina does not appear to be a major biogenic component, nor does it appear to have built mounds in northeastern British Columbia. In other respects, however, it closely resembles *Palaeoaplysina* from elsewhere (*see* Davies and Nassichuk, 1973; Breuninger, 1976). It has a tabular form and occurs as broken fragments 1 mm to 6 mm thick and at least 3 cm long (plates are incomplete). There are sediment-filled internal canals enveloped in sparry calcite probably occupying space originally filled by soft tissue. One side of the plate is an irregular surface upon which encrusting foraminifers and bryozoans occur (probably the upper surface, by analogy with other localities). The surface pores and mamelon-like structures reported from other localities have not been observed in material from northeastern British Columbia.

The lower Moscovian occurrence in northeastern British Columbia is one of the oldest known occurrences of *Palaeoaplysina*. In Idaho, *Palaeoaplysina* occurs widely in rocks known to be Virgilian (latest Carboniferous) and Wolfcampian (earliest Permian) in age, although the earliest occurrences there could be Moscovian according to Breuninger (1976). The Richardson Mountains occurrence is Permian (Davies, 1971), and the Arctic Islands occurrences are mainly Permian although, there, *Palaeoaplysina* also occurs in sediments of Late Carboniferous age (late Moscovian; Davies and Nassichuk, 1973).

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30. DEVONIAN COAL IN THE SUBSURFACE OF GREAT SLAVE PLAIN: A GUIDE TO EXPLORATION FOR OIL AND GAS

Projects 750088 and 710011

P.R. Gunther and N.C. Meijer-Drees
Institute of Sedimentary and Petroleum Geology, Calgary

Introduction

A detailed study of beds transitional between Middle and Upper Devonian began under Project 710011. Several cores from exploratory wells drilled in the southern part of the Mackenzie District, Northwest Territories were examined and, in the four wells indicated on Figure 30.1, these beds contain very thin coal seams and coaly plant fragments. Figure 30.2 shows the stratigraphy of the coal seams; their complex relationships as interpreted by N.C. Meijer-Drees, are discussed in the first part of the paper.

Vitrinite reflectance study of coal samples was carried out by P.R. Gunther and the results are discussed in the second part of the paper.

Geology

In the southern part of the Mackenzie District, coal commonly is present in a dark-coloured unnamed shale unit that underlies the Upper Devonian Fort Simpson Formation, and the discovery of two coaly plant fragments in the Canol Formation of the CPOG Kugaluk N-02 well (Lat. 68°31'55"N, Long. 131°31'19"W) at depths of 882.63 and 883.61 m (2892.5 and 2899 ft.) indicates that coal may be distributed widely in the Devonian of the subsurface of the Northwest Territories.

In the southern part of the District, coal is present at two different stratigraphic horizons. The older coal analyzed occurs in a core of the IOE Providence K-45 well (Lat. 61°34'36.00"N, Long. 117°08'47.78"W) at

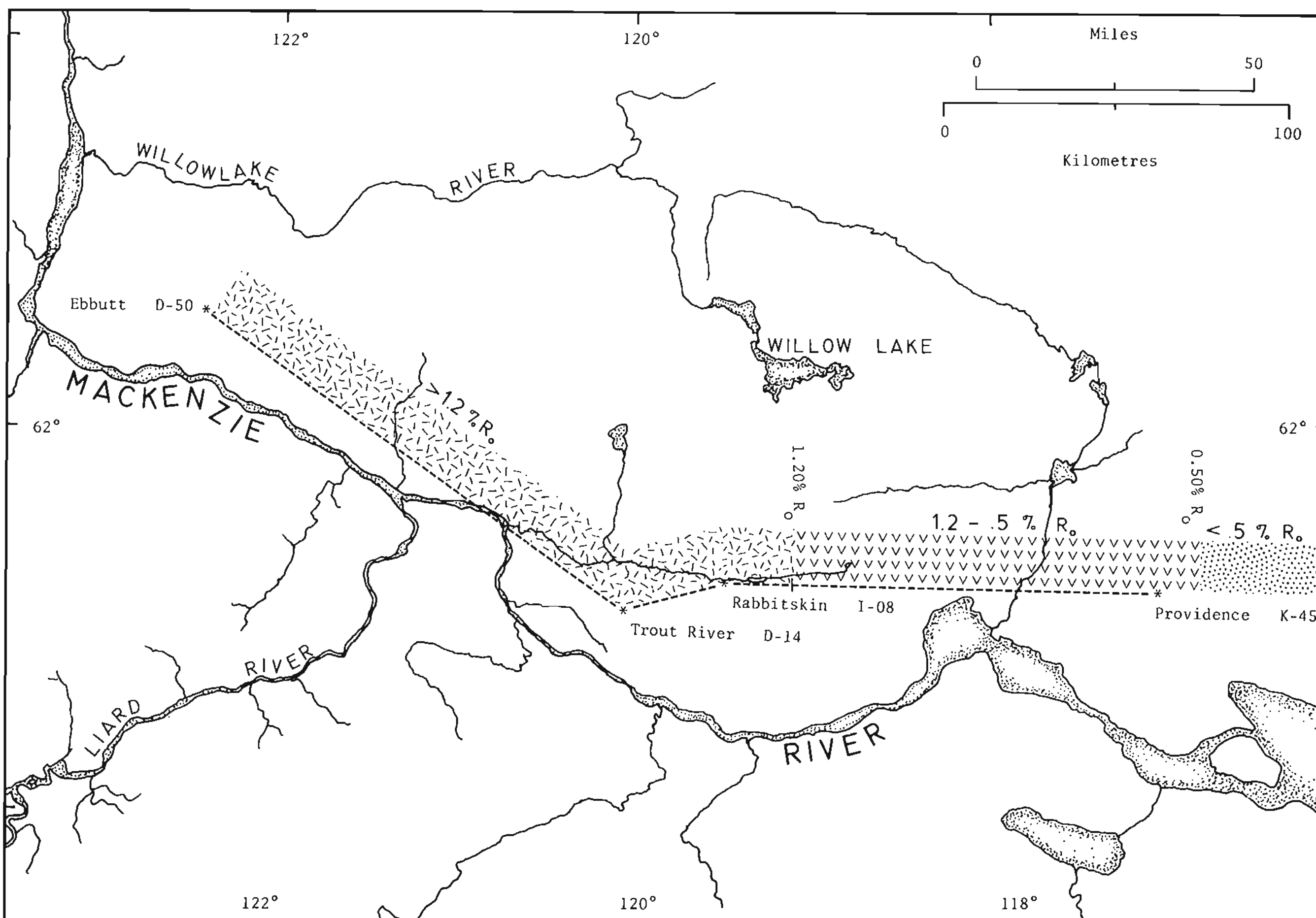

Figure 30.1. A location map indicating three zones of maturity for the study area — the southwestern part of the Northwest Territories.

From: Report of Activities, Part A;
Geol. Surv. Can., Paper 77-1A (1977).

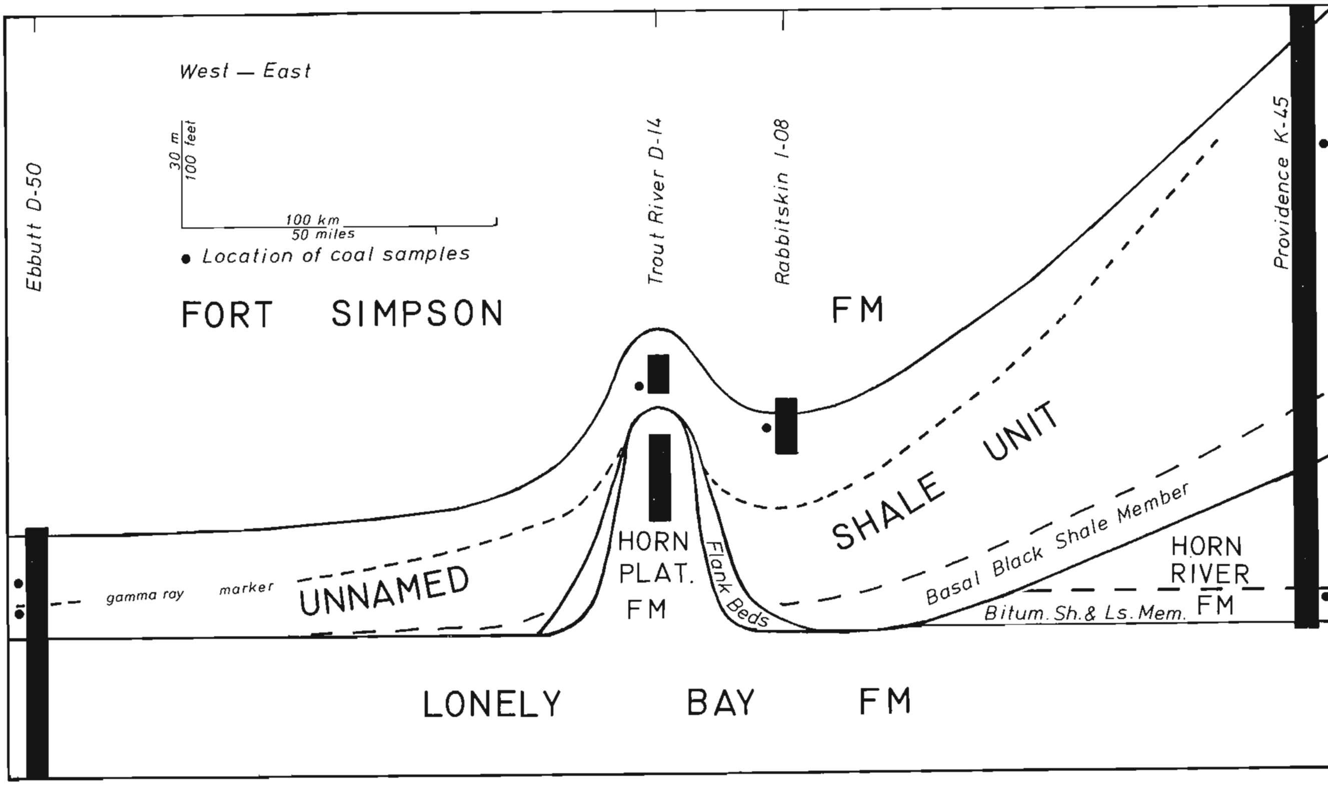

Figure 30.2. Stratigraphic cross-section showing the location of the coal samples and the cored intervals.

depths 251.92 and 252.22 m (826.5 and 827.5 ft.) as two very thin seams in a dark greyish brown succession of thinly interbedded shale, marl and argillaceous limestone (*see* Fig. 30.2). The shale contains, in places, and abundance of tentaculitids; some of the limestone beds are rich in brachiopods of the lower to mid-Givetian *Leiorhynchus castanea* zone (*see* Norris *in* Brideaux *et al.*, 1976, p. 12, 13). The dark coloured shale, marl and limestone succession has been referred to the Horn River Formation (Norris, 1965) and conformably overlies the limestone of the Lonely Bay Formation. Subsurface correlations in the Great Slave Lake area suggest that this shale, marl and limestone unit is continuous with the Bituminous Shale and Limestone Member of the Pine Point Formation.

Both seams are interbedded with laminated shale rich in tentaculitids suggesting a marine subtidal environment of deposition. It was not possible to determine whether the coal seams represent pieces of driftwood or if they are deposits of subaqueous vegetation.

The coal seam found in the IOE Providence K-45 well at a depth of 142.80 m (468.5 ft.) is 2.5 cm (1 in.) thick contains abundant calcite veins and some light grey, discontinuous shale laminae. The coal seam occurs in a light greenish grey, calcareous shale similar to the shale of the Fort Simpson Formation.

The coal seams of the IOE Triad Ebbutt D-50 well (Lat. 62°19'01"N, Long. 122°24'05"W), the Gulf *et al.* Trout River D-14 well (Lat. 61°33'10.18"N, Long. 120°03'39.96"W), and the Horn River Decalta *et al.* Rabbitskin I-08 well (Lat 61°37'32"N, Long. 119°30'17"W) occur in a very dark grey carbonaceous shale which contains scattered pyrite, ?sponge spicules and conodonts. The conodonts were examined briefly by T.T. Uyeno of the Institute of Sedimentary and Petroleum Geology and indicate an early Fransnian age (*see* also Braun *in* Norford *et al.*, 1970, p. 12; Fuller and Pollock, 1972, p. 153).

The dark grey carbonaceous shale (the unnamed unit on Fig. 2) containing the coal seams is correlative with the Spence River Formation as defined by Hunt (1954, p. 2300). In the Providence K-45 well, this unnamed shale unit overlies the Horn River Formation with an erosional contact and directly overlies the Lonely Bay Formation in the Rabbitskin I-08 and Ebbutt D-50 wells (*see* Fig. 30.2). In the Ebutt D-50 well, the section between the equivalent of the Lonely Bay Formation and the Fort Simpson Formation is greatly reduced in thickness. The uppermost part of the Lonely Bay Formation (the *Leiorhynchus castanea* Zone) is missing because of erosion. Subsurface correlations in the Great Slave Plain suggest that the lower part of the unnamed shale unit (the Basal Black Shale Member on Fig. 30.2) is missing because of non-deposition. The upper part of the unnamed shale unit is distributed widely and covers the reef deposits of the Horn Plateau Formation in the Trout River D-14 well where it contains conodonots of the lower Frasnian *Polygnathus asymmetricus* Zone (Fuller and Pollock, 1972).

The regularly laminated nature of the unnamed shale unit and the absence of fossils other than conodonts, ?sponge spicules and plant fragments suggest a subtidal, euxinic environment of deposition.

Organic Maturity

Vitrinite reflectance values were determined for seven samples in the four wells indicated in Figure 30.1. The values suggest a useful method by which to zone the area laterally, as far as organic maturation is concerned, and to indicate the most favourable areas for hydrocarbon exploration. It must be realized that data are few and this information is provided only as a guide. Further study is required to substantiate the trend described below.

Vassoevich *et al.* (1974) gave a zonation for hydrocarbon maturity as it is related to vitrinite reflectance. They indicate that sediments containing coals having reflectance (Ro) values less than 0.50 per cent Ro (immature) may produce and contain gas, but oil is unlikely to occur. Sediments containing coals having Ro values between 0.50 and 1.20 per cent Ro (mature) may produce and contain liquid hydrocarbons, and are the most favourable zones for oil exploration. Sediments containing coals having Ro values above 1.20 per cent Ro (overmature) may produce and contain diagenetic gas. Very few gas accumulations are found coincident with coals approaching the 3.0 per cent Ro value.

Reflectance data from the studied boreholes and Kugaluk N-02 are contained in Table 30.1. The Rabbitskin and Trout River wells are considered as one for the purpose of the study because of their proximity, while the averaged values from the other two wells are used in conjunction to scale the line of cross-section (Fig. 30.1). The map illustrates the overmature and mature areas along the line of section. The immature zone is interpreted to continue east of the area figured in the map.

Gas and Oil Potential

It appears Devonian coals may be used to zone the study area into regions of varying potential for oil and gas exploration. The reflectance values (increasing towards the west) agree with the generally accepted trend of higher temperature and deeper burial as the Disturbed Belt is approached (Hacquebard, 1975; Evans and Staplin, 1971). As far as organic maturity is concerned, the most favourable area for Devonian oil exploration is east of the 1.20 per cent Ro value located on Figure 30.1. West of this point along the line of section one could only anticipate gaseous hydrocarbons.

Of the four studied wells, only Trout River D-14 tested hydrocarbons; 73 mcf/d and 180 feet of gas cut mud at the top of the Horn Plateau Formation. A well, located just west of the Providence K-45 borehole, indicated oil bleeding from core no. 1. The well is Horn River-Placid-IOE Mink Lake I-38 (Lat. 61°37'31"N, Long. 117°35'51"W) and the depths of bleeding are 270.82 m (888.5 ft.) and 271.27 m (890 ft.). Both the gas show and the oil bleeding are evidence that substantiate the proposed zonation.

Dead oil seeps on the northwest side of Great Slave Lake (A.W. Norris, pers. comm.) would suggest migration of at least some liquid hydrocarbons in the vicinity of the study area.

Table 30.1

Borehole name	Depth sampled Metres	Feet	Vitrinite Reflectance	Ro values used in Figure 30.1	Organic Maturity
Providence K-45	142.80	468.5	0.57 ± .01		
	251.92	826.5	0.52 ± .02	0.54	Marginally mature
	252.22	827.5	0.54 ± .01		
Rabbitskin I-08	421.23	1382	1.32 ± .08	1.31	Overmature
Trout River D-14	459.33	1507	1.31 ± .04		
Ebbutt D-50	451.10	1480	3.09	2.96	Extreme overmature
	458.72	1505	2.83		
Kugaluk N-02	882.63	2892.5	2.24 ± .08		
	883.61	2899	2.48 ± .10		

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AN OCCURRENCE OF COAL IN THE BLAIRMORE GROUP ON WAIPAROUS CREEK, CENTRAL FOOTHILLS, ALBERTA

Project 760018

J. R. McLean
Institute of Sedimentary and Petroleum Geology, Calgary

Introduction

Coal seams of commercial thickness in the Blairmore Group generally are stated to be confined to an area north of the Clearwater River (Holter and Mellon, 1971; MacKay, 1949). A seam, 4.25 m (13 ft.) thick, has been found in outcrop on Waiparous Creek (SE21-28-8W5), 43 miles southeast of the Clearwater River (Fig. 31.1).

Stratigraphy and Depositional History

The Blairmore Group forms the basal unit of the Lower Cretaceous in the Foothills of western Alberta. Traditionally, the coal-bearing member of the group has been referred to as the Luscar Formation (MacKay, 1929a, b, 1930). More recently, Mellon (1967) has proposed a nomenclature based on a type section in southwestern Alberta (Table 31.1) and, in northeastern British Columbia, Stott (1960, 1968) has recognized several subdivisions of the Blairmore. The Moosebar Formation of Stott's nomenclature, a marine shale, now has been recognized south of Cadomin (Fig. 31.1), forming an important break in what was previously considered an entirely nonmarine section. In the area under consideration, this marine tongue has not been

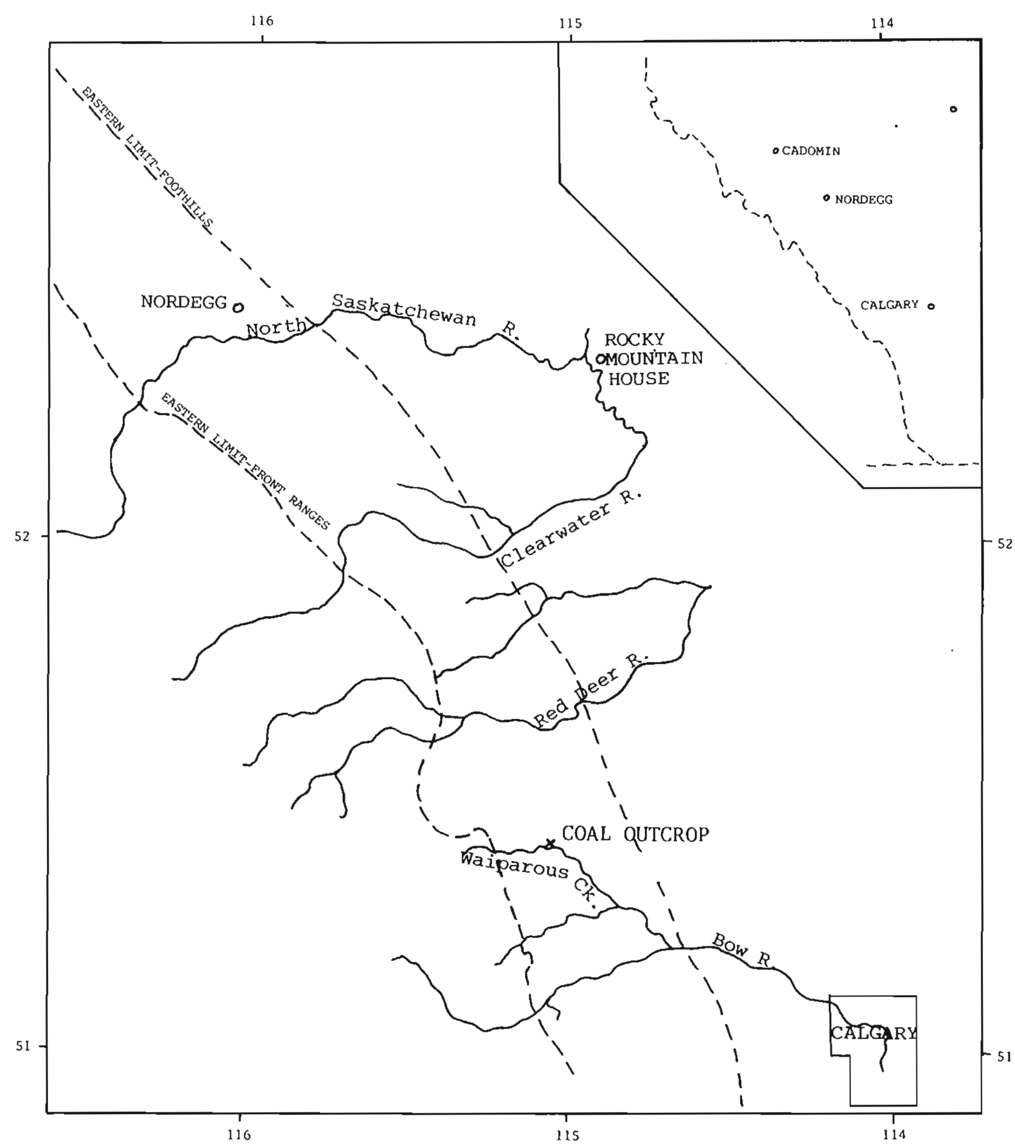


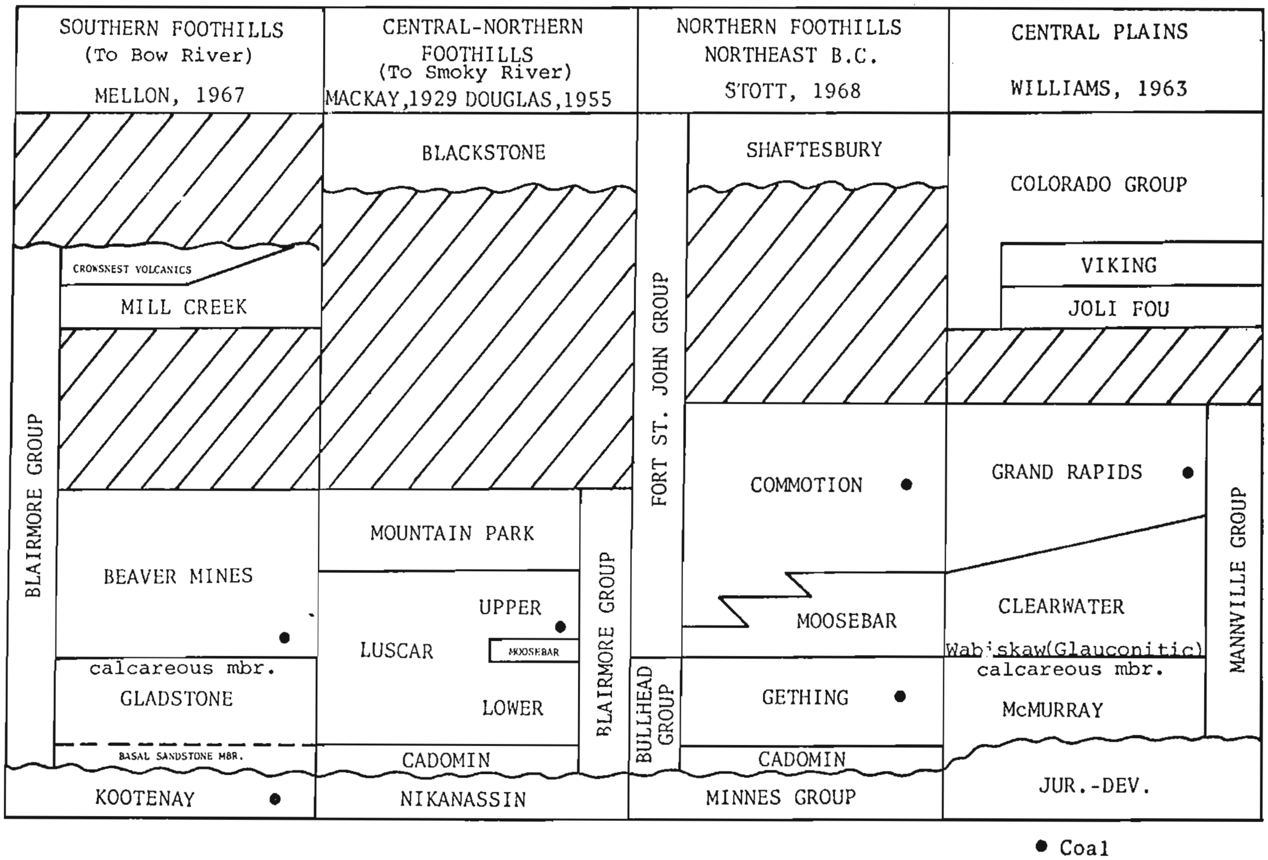
Figure 31.1

Location of Waiparous Creek section.

From: Report of Activities, Part A;
Geol. Surv. Can., Paper 77-1A (1977)

Table 31.1

Stratigraphic Chart



recognized but the "calcareous" member of Mellon's (1967) Gladstone Formation, which was found to immediately underlie the Moosebar Formation at Cadomin, is well represented.

A log of the stratigraphic section on Waiparous Creek (Fig. 31.2) shows an abundance of calcareous, terrigenous clastic sediments and two prominent limestone beds (units 1 to 13) below the coal seam, representing the "calcareous" member. Fossil evidence from this unit (Mellon, 1967, p. 74) indicates a fresh- to brackish-water environment of deposition. Aphanic limestone (unit 12) suggests precipitation accompanying evaporation in shallow lakes, while bioclastic limestone (unit 9) suggests wave action with reworking and concentration of shell debris, probably as chenier ridges. Associated terrigenous clastics may be shoreline and overbank deposits from streams entering lakes or large shallow bays and lagoons.

Units 14 to 16 occupy the same stratigraphic position as the Moosebar marine tongue at Cadomin, which Mellon (1967, p. 58) relegated to the base of his Beaver Mines Formation. No marine fauna have been identified yet and, although the sea may never have reached this area, the physical controls accompanying a major marine transgression would certainly have affected sedimentation well beyond the limits of the maximum transgression.

There is a distinct break in colouration from dark grey in unit 13 to olive-grey in unit 14 and greenish grey in higher units. This is related to a change in provenance, with volcanogenic detritus common in Mellon's (1967) Beaver Mines Formation but absent in his Gladstone Formation.

The coal seam (unit 17) represents a marsh or swamp environment. Seams at the same stratigraphic position are known from the Clearwater River (Holter and Mellon, 1971, p. 130) to northeastern British Columbia (Stott, 1974, p. 98), and are progressively younger to the north in the direction of regression of the sea. These seams represent coastal marshes and swamps, often separated from the underlying Moosebar shales by a sand barrier. The coal seam at Waiparous Creek may represent an extensive marsh-swamp complex which developed landward of the maximum extent of the Moosebar-Clearwater sea.

Coal deposition was terminated by the influx of coarser clastic fluvial sediments which gradually infilled the seaway, pushing the coastal environments ever farther northward and eastward.

Structure

The coal outcrop occurs in the Waiparous thrust sheet (Ollerenshaw, 1972), has a strike of 130 degrees, and a dip of 55 degrees southwest. This fault slice extends for 20 miles along strike, but outcrop is poor in most of this area, and the coal seam was not observed elsewhere.

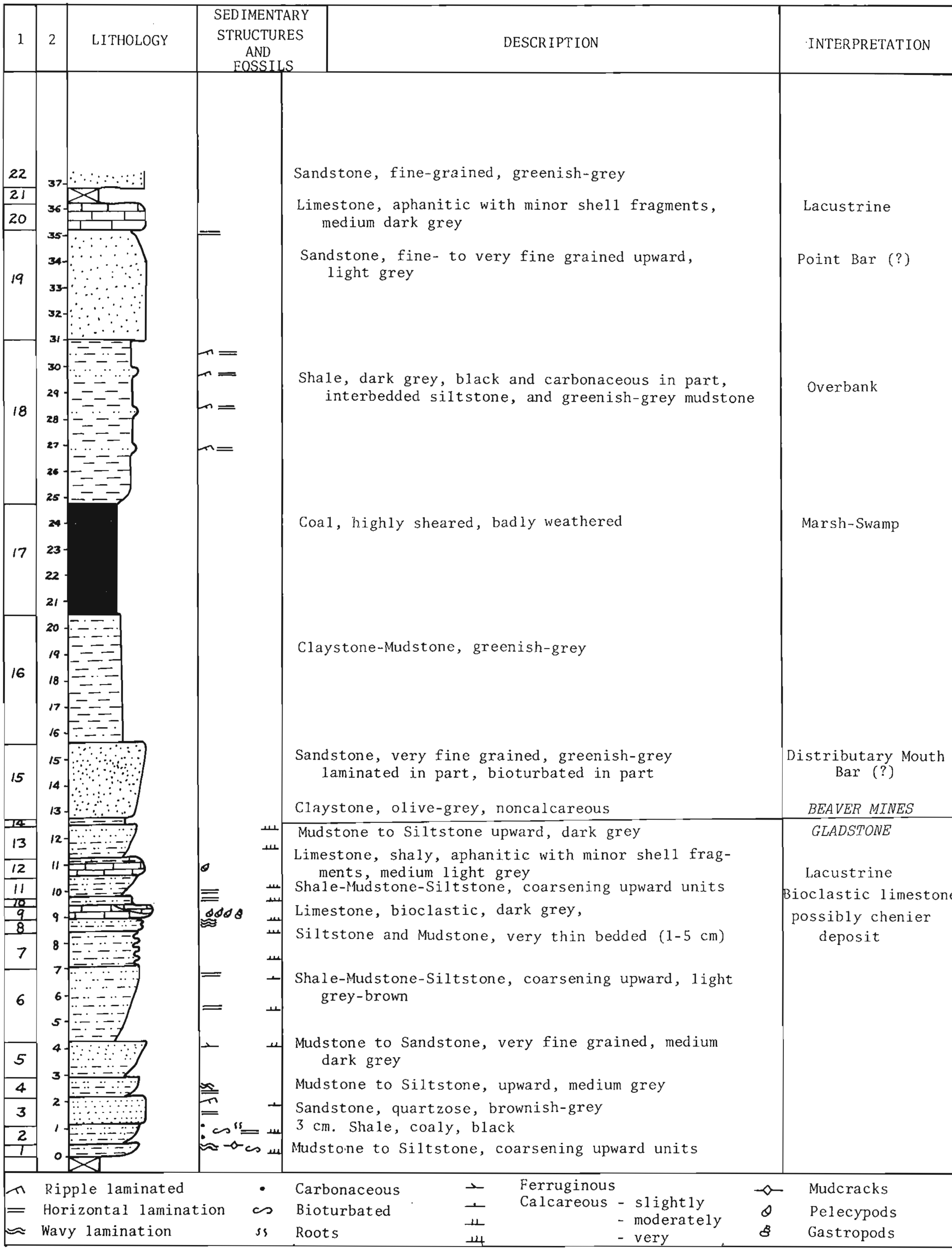


Figure 31.2. Stratigraphic section — Waiparous Creek. (1 — Unit number; 2 — Thickness in metres).

Discussion

Coal of commercial thickness has not been reported previously in this area, but its presence is not altogether unexpected. Yurko (1975, Fig. 10) in a study of the deep Cretaceous coal resources of the Alberta Plains, indicates coal seams in the Mannville Group with a cumulative thickness ranging from 1.5 to 9.1 m (5-30 feet) adjacent to the Foothills, 25 miles northeast of the Waiparous locality. Further extrapolation from Yurko's data suggests that coal might be found in the Foothills as far south as the latitude of Calgary (51°N Latitude), but probably no farther.

Examination of well logs in the Plains area adjacent to the Waiparous Creek location indicates that the best seam lies directly above, or only a short distance above, the "glauconitic" beds (Table 31.1) which overlie the "calcareous" member. Thus, it occurs in the same stratigraphic position as the seam on Waiparous Creek.

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Project 750082

N. C. Ollerenshaw
Institute of Sedimentary and Petroleum Geology, Calgary

Geological mapping and a detailed structural investigation of Dominion Coal Block, Parcel 73, to determine the quantity, distribution and geometry of the coal reserves, were begun in June 1975 and continued in the summer of 1976. It comprised a total of 15 weeks of field work as part of a collaborative project with the British Columbia Department of Mines and Petroleum Resources. Mapping was on a scale of one inch to four hundred feet.

Parcel 73 is situated at the south end of Sparwood Ridge and the northern end of Hosmer Ridge, about 6 miles south of the town of Sparwood, British Columbia and 1.5 miles east of the Elk River (Lat. 49°38'N, Long. 114°51'W). Natal fire lookout tower located 7390 feet above sea level, the highest point in the block, is about 3900 feet above the Elk River Valley. Within the area, slopes are very steep and thickly forested. Slopes of 40 degrees are not uncommon.

Parcel 73 is approximately 3.6 miles (E-W) by 2.1 miles (N-S) and covers a little more than 7.5 square miles. This parcel, together with Parcel 82, came into the possession of the Government of Canada in 1897 under an agreement with Canadian Pacific Railway.

Parcel 73 is separated into two sharply contrasting segments in terms of structure and stratigraphy, along a north-south surface of structural discontinuity, the Dominion thrust (Fig. 32.1). To the west of Dominion thrust, the surface structures are complex, with numerous imbricate fault slices involving the Fernie and Kootenay formations at the surface. Coal seams are exposed at the surface. East of Dominion thrust, the surface rocks belong mainly to the younger Blairmore Group and the structure consists of gentle, open folds. Coal seams in this sector must be at considerable depth.

Stratigraphy

The Fernie Formation of Jurassic age represents the oldest rocks in the area, but it is not well exposed. It occurs as small scattered outcrops of dark grey to greyish black shale with minor siltstone and sandstone beds and a unit of flaggy, brown weathering, limonitic sandstone at the top.

The overlying Jurassic-Lower Cretaceous Kootenay Formation commences with a conspicuous unit of hard, cherty, cliff-forming sandstone, the Moose Mountain Member (*see* Fig. 32.2) at the base, that is locally difficult to distinguish from the uppermost Fernie sandstone and other sandstone units in the lower Kootenay. Above the Moose Mountain sandstone in the Coal Bearing Member of the lower Kootenay, consisting of dark grey to black mudstones and shales, and limonitic siltstones and sandstones with local channels of pebble conglomerate. This part of the section contains all the commercial coal seams. Several examples of fossilized tree stumps, still in their growth position, were found, apparently inundated and buried by overbank deposits (*see* Figs. 32.3, 32.4).

Another conspicuous sandstone unit developed in and around Parcel 73 is designated informally the Lookout sandstone, because of its presence immediately beneath the fire lookout tower and, in fact, forming a cap to Lookout Hill. The sandstone recurs above Saddle thrust, producing a locally prominent cliff at the north end of Hosmer Ridge. This sandstone is petrographically similar to the Moose Mountain Member.

The interval between the Moose Mountain sandstone and the Lookout sandstone contains several thick, commercial coal seams, the largest of which, the Lookout seam, occurs some 30 m below the Lookout sandstone.

Several other thick coal seams occur above the Lookout sandstone. These have been removed by erosion on Lookout Hill but occur above Saddle thrust at the north end of Hosmer Ridge in the southwest corner of Parcel 73. Figure 32.1 shows the distribution of the Lookout sandstone in two plates, one in the southwest corner of the Parcel and the other slightly northeast of it. These plates can be used to determine the approximate position of the coal seams and the topographic highs. The northeastern plate outlines Lookout Hill, the highest point of which is on the northeasternmost tip at the fire lookout tower. The coal seams lie between the base of this sandstone plate and the Moose Mountain Member. The southwestern plate outlines the north end of Hosmer Ridge and has coal seams below it, as in the case of the Lookout Hill occurrence, but also has a thick section including at least two commercial coal seams above it.

Lookout Hill and Hosmer Ridge constitute the best areas for coal development, with several thick seams dipping at 15 to 25 degrees south to southwest and subject to only local structural complications. The best and thickest seams are the Lookout seam, some 30 m below the Lookout sandstone and the Wheeler seam some 70 m above the sandstone. Both seams are typically spit by a mudstone/siltstone unit and locally by several such units. It is unfortunate that the coal-bearing strata in the lower part of the Kootenay Formation are involved in the structural deformation and the coal is commonly sheared and locally disrupted. Figure 32.5 illustrates minor structural displacement in the roof of the Lookout seam. Small faults of this type are common in the area.

The upper, or Elk, Member of the Kootenay Formation is generally similar to the Coal Bearing Member, but lacks commercial coal seams. It contains numerous pebble conglomerate beds in this area, although conglomerates are much less abundant than in the section on Coal Creek several miles to the south (Newmarch, 1953). Sandstones in the lower part of the Elk resemble those of the Coal Bearing Member. Some

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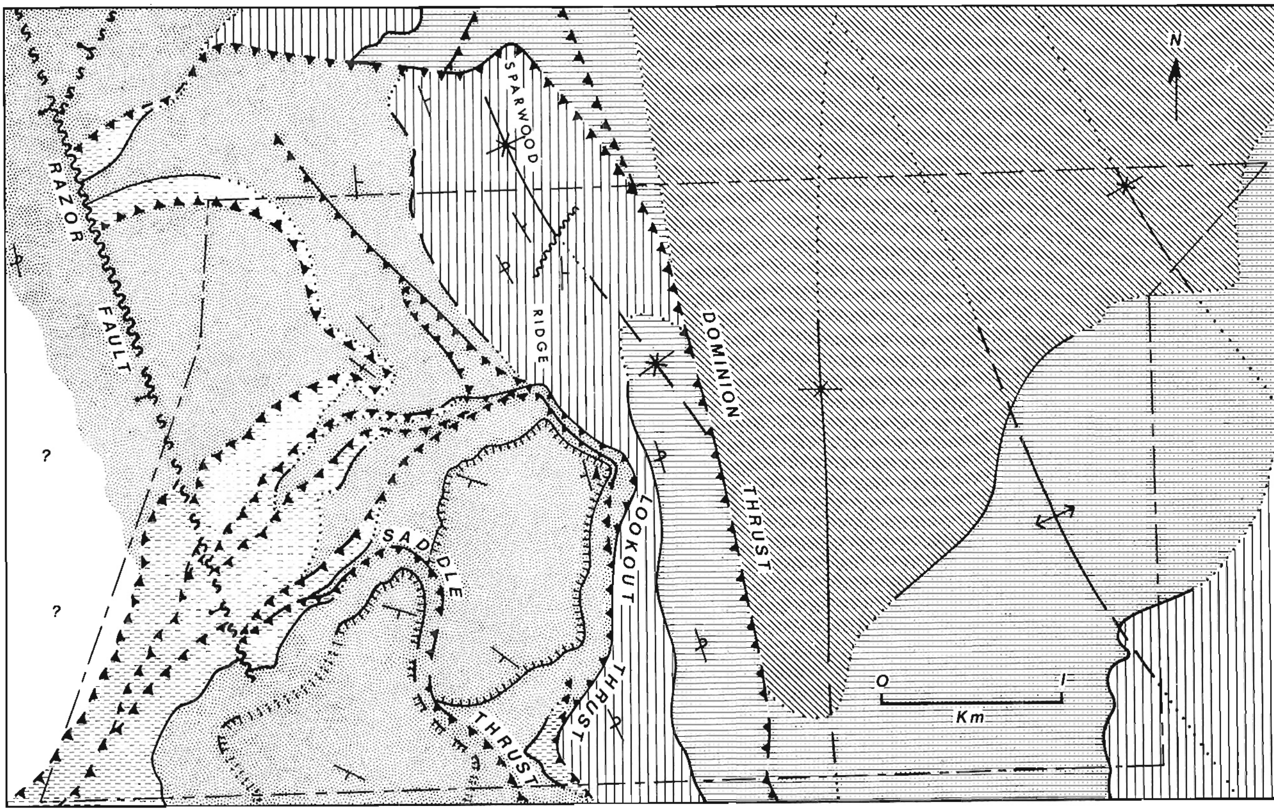
Geol. Surv. Can., Paper 77-1A (1977)



LEGEND

LOWER CRETACEOUS
BLAIRMORE GROUP
BEAVER MINES (and possible Mill Creek) FORMATION

LOWER BLAIRMORE GROUP AND CADOMIN FORMATION

JURASSIC-LOWER CRETACEOUS
KOOTENAY FORMATION
Elk Member

Coal Bearing Member

JURASSIC
FERNIE FORMATION

Geological contact (defined, approximate, assumed)............

Bedding (inclined, vertical, overturned)......................

Fault (defined, approximate, assumed) (bars on downthrown side)

Thrust fault (defined, approximate, assumed) (teeth on upthrown side)..

Anticline (defined, approximate, assumed).....................

Syncline (defined, approximate, assumed)......................

Base of Lookout Sandstone.....................................

Boundary of Parcel 73...

Natal Fire Lookout Tower......................................

Figure 32.1. Preliminary geological map of Dominion Coal Block, Parcel 73.

siltstones and sandstones in the upper part, however, resemble those of the lower part of the Blairmore Group in the Alberta Foothills. Thin seams of "Needle coals", composed of a felt of coalified needles, are a distinctive feature of the upper part of the Elk Member.

The Cadomin Conglomerate overlies the Kootenay Formation without any obvious discordance. The contact is not well exposed and the Cadomin occurs in an interval of conglomerate beds, rather than as an isolated individual conglomerate. Green, and minor red, mudstones and greenish grey sandstones indicate the presence of Lower Blairmore strata above the Cadomin.

The youngest bedrock in Parcel 73 consists of greenish grey mudstones and siltstones with brown weathering, feldspathic sandstones of the Beaver Mines and possibly Mill Creek formations of the Blairmore

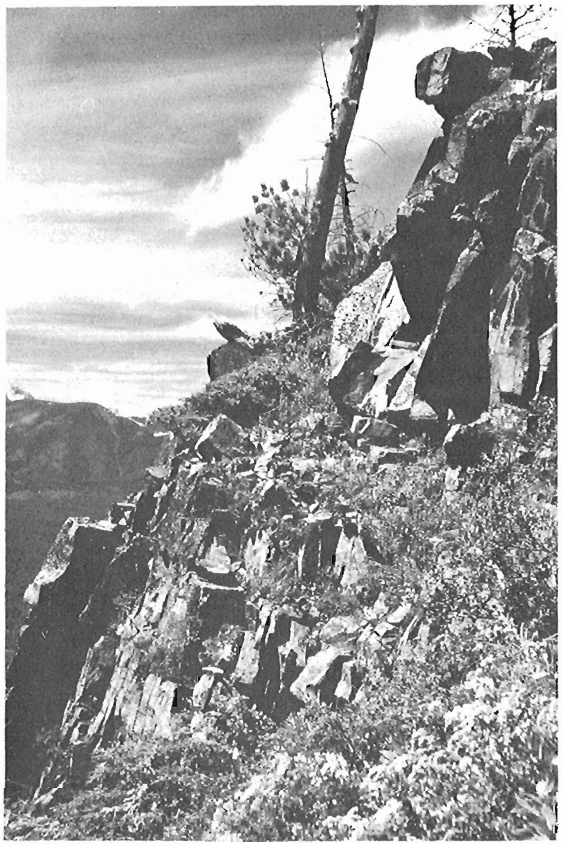


Figure 32. 2. Resistant sandstone unit, probably the Moose Mountain Member, above the window in the northeast corner of Parcel 73. GSC 199228

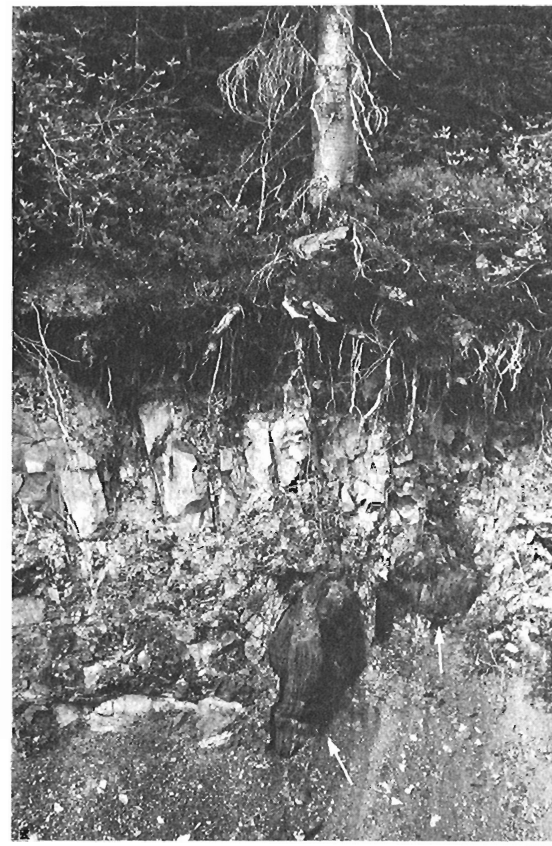


Figure 32. 3. Fossil tree trunks in growth position. GSC 199229

Figure 32. 4.

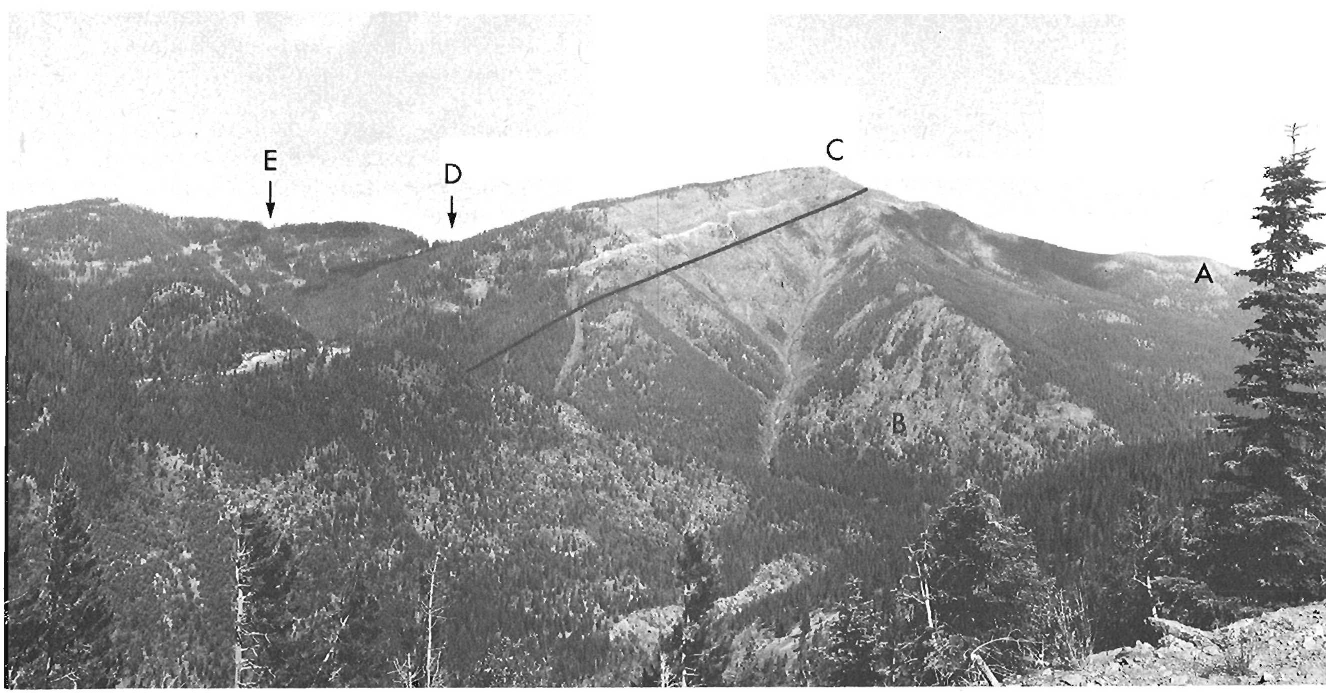
Fossil tree trunk in growth position and relict root zone. GSC 199230

Group. "Igneous pebble conglomerates" outcrop locally in the youngest strata preserved and are similar to those found in the Beaver Mines Formation in the Alberta Foothills north of Bow River. The Cadomin Formation and Blairmore Group are Early Cretaceous in age.

The Fernie Formation is of marine origin, the Moose Mountain Member probably marine to brackish, whereas the rest of the succession is apparently entirely nonmarine as far as has been established. Pearson and Duff's (1975) claim that there is evidence of marine beds in the Kootenay of this region has not been substantiated.

Figure 32.5.

Minor fault offsetting the roof of Lookout Coal Seam, west side of Hosmer Ridge. GSC 199227



A = Gently folded middle and upper Blairmore Group;
B = Overturned Cadomin and Lower Blairmore strata;
C = Fire Lookout tower and Lookout thrust (black line);
D = Saddle thrust;
E = Wheeler Seam.

Figure 32.6. Looking north from Marten Ridge across Wheeler Creek and Wheeler Ridge at Lookout Hill and a cross section view of Parcel 73. GSC 199225 and 199226.

Structure

Figure 32.1 illustrates the author's preliminary interpretation of the geology of Parcel 73. The Dominion thrust is the lowermost fault exposed and underlies the entire western half of the area. This thrust appears to have a southerly component of dip, so that its trace swings over and across Sparwood Ridge on the north. The plate is terminated in that direction by erosion, just north of Parcel 73. On the evidence of scattered and imperfect outcrop control, it is concluded that the Dominion thrust may be exposed in a window in the northwest corner of Parcel 73. This window is truncated on the western side by a steeply dipping normal fault, Razor Fault, downthrown on the west side.

Lookout thrust (Fig. 32.6), another southwest-dipping thrust fault, overlies the Dominion plate and probably is a large splay off the Dominion thrust. The panel between Dominion and Lookout thrusts consists of a very tight syncline of upper Kootenay Formation (Elk Member) and Lower Blairmore Group, with an overturned west limb, gradually losing the inversion northward away from the influence of lookout thrust. The overturning is very conspicuous just east of the fire lookout tower at the level of the Cadomin Conglomerate. The Lookout thrust dips to the southwest, so that its trace wraps around the west side of Lookout Hill and the north end of Hosmer Ridge. On its northwest side, the structure of the thrust is complicated and has to be interpreted from scattered outcrops in thick forest cover. It appears that the structure is imbricate, with several spalys enclosing lenticular wedges of Fernie shales and lower Kootenay strata, causing a piling up and thickening of the Lookout plate at this point. Scattered exposures of the Moose Mountain Member, the sandstone at the base of the Kootenay Formation, provide the evidence for this structure, although it has been noted above that other sandstones in the lower part of the Kootenay, within Parcel 73, strongly resemble this sandstone, thereby causing some confusion and making outcrop correlations a matter of interpretation. Once again there is some evidence to suggest that the northwest edge of the Lookout plate is displaced slightly by the southeastern extremity of the Razor Fault.

Although there are some local complications above the Lookout thrust on the southeastern side of Lookout Hill, it does not appear that the imbricate wedges continue under the ridges in this direction.

The Saddle thrust is the uppermost fault noted in Parcel 73. It is a relatively minor splay off the Lookout thrust system, but forms a distinctive and mappable feature which offsets the best coal-bearing strata of the lower Kootenay and runs through the saddle between Lookout Hill and Hosmer Ridge.

Razor Fault is a conspicuous structure on Razor Ridge just northwest of Parcel 73 where gently dipping Moose Mountain sandstones on the northeast side are truncated abruptly against steeply dipping, younger Kootenay strata on the southwest sides. Farther southwest, the Moose Mountain sandstone recurs, but overturned, implying another thrust fault between the Fernie of Elk River Valley and the Kootenay of Sparwood Ridge. This thrust has not previously been reported and could be a major fault along much of this segment of the Elk Valley.

The alignment of Razor Fault in relation to Wilson Fault, mapped by Price (1962) on the opposite side of Elk River Valley, suggests that Razor Fault may be the southeast extension of Wilson Fault.

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33. Rb/Sr ISOCHRONS FOR ARCHEAN GRANITOID PLUTONS WITHIN THE WABIGOON GREENSTONE BELT, NORTHWESTERN ONTARIO: A PRELIMINARY EVALUATION

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Dieter Birk[1] and Robert H. McNutt[1]
Regional and Economic Geology Division

Introduction

The Wabigoon Greenstone Belt is a region of metavolcanic-volcaniclastic rocks forming a lithostratigraphic subdivision to the Superior Province Characteristic of the Belt are internal high level granitoid plutons of either homogeneous granodiorite or multiphase granite-monzodiorite lithology. Recent mapping (Ayres, 1974) demonstrates that some granitoid batholiths considered continuous across large areas of the Shield are actually aggregations of numerous discrete plutons with dimensions comparable to the stocks that intrude the supracrustal rocks of the Wabigoon Belt. The role of small granitoid bodies in the genesis of Archean granitoid terrain therefore may have been underestimated.

Preliminary results of a Rb/Sr isotopic investigation of three such plutons the Burditt Lake Stock, Flora Lake Stock and Taylor Lake Stock intrusive into the Wabigoon Belt are presented. All three plutons are on the Kenora – Fort Frances Sheet of Davies and Pryslak (1965).

Blackburn (1976a) outlined the Burditt Lake Stock as a bilobed porphyritic to equigranular granodiorite roughly three miles in diameter and intrusive into lapilli metatuff. Detailed mapping (Fig. 33. 1) has established two mineralogical assemblages and four textural units within the borders of the pluton but all variations are sufficiently subtle and gradational to justify field classification as "homogeneous". The southern lobe is an equigranular muscovite-biotite granodiorite peripherally foliated concordant to the metavolcanic host; with abundant marginal aplitic zones and dykes. The north lobe is a biotite-hornblende granodiorite carrying microline-perthite megacrysts. All mineralogical and textural differences are attributed to endoblastesis and deuteric metasomatism (Birk and McNutt, 1976).

In contrast, at Flora Lake (Fig. 33. 2) a concentrically zoned granite-monzonite-monzodiorite pluton shows evidence of multiple intrusion (Heimlich, 1965). Autoliths of border monzodiorite and monzonite – occasionally on a mappable scale – have been rafted by pink granite occupying the core and dykes of this granite traverse both of the peripheral units.

The third stock, at Taylor Lake, 45 km south of Dryden, Ontario, was mapped recently by Blackburn (1976b). Pichette (1976) conducted a detailed study that uncovered a composite monzodiorite-monzonite-granodiorite pluton with at least six intrusive phases as well as an autometasomatic overprint.

Mineral modes for Burditt Lake reflect the subtle differentiation: average modal per cent of quartz ranges from a low of 21. 9% for megacryst bearing biotite-hornblende granodiorite to a high of 27. 5% for medium grained muscovite-biotite granodiorite. Flora Lake and Taylor Lake stocks show much greater diversity. The monzodiorite rim at Flora Lake carries only 0. 2% quartz while the granite core contains 22. 5% quartz (Heimlich, 1975). For Taylor Lake, Pichette (1976) found a range of 5. 6 to 24. 3 % quartz between major units. These three plutons therefore provide examples of homogeneous single stage as well as multistage intrusion into the Wabigoon supracrustals. These plutons were emplaced at high levels as magmatic diapirs, as evidenced by one or more of the following:

1. sharp contacts, intrusion breccia (Figs. 33. 1, 33. 2)
2. peripherally aligned abundant mafic enclaves, schlieren
3. metabasalt enclaves bearing pillow selvages (Pichette, 1976)
4. associated aplites, pegmatites
5. synneusis texture in plagioclase, sphene
6. microcline-perthite megacrysts, feldspathized enclaves.

Analytical Procedures

Laboratory chemical and mass spectrometric methods have been reviewed by Gibbins and McNutt (1975). Five replicate analyses of the Eimer and Amend $SrCO_3$ isotope standard averaged 0. 7080 ± . 0005 (1 sigma). Precision in the Rb/Sr ratio determination is estimated at 1 per cent (1 sigma). Rb/Sr ratios were calculated from unknown-versus-standard XRF count ratios by linear least squares regression (Marchand, 1973; Doering, 1968). This direct determination results in better precision (elemental Rb, Sr: 3%, 1 sigma).

All sets of whole-rock isochron data have been tested with the multiple regression treatment of Brooks *et al.* (1972) using blanket error estimates of 1% and . 065% for $^{87}Rb/^{86}Sr$ and $^{87}Sr/^{86}Sr$ respectively (1 sigma). Isochrons have been defined by the York I linear regression model with all isochron error parameters expressed at the 95% confidence interval (2 sigma). The acceptability of each linear regression as an isochron *sensu stricto* has been tested using the mean square of weighted deviates (MSWD of McIntyre I, Brooks *et al.*, 1972). Constants used in calculation include normalization to $^{86}Sr/^{88}Sr$ = 0. 1194 and

[1]Department of Geology, McMaster University, Hamilton, Ontario L8S 4M1

$\gamma = 1.39 \times 10^{-11} y^{-1}$. Ages expressed by the new ^{87}Rb decay constant ($1.42 \times 10^{-11} y^{-1}$) would be 2 per cent lower than quoted here.

Sample numbers for the Burditt Lake and Flora Lake isochrons correspond to the station numbers on Figures 33.1 and 33.2. Sample numbers for the Taylor Lake Stock are derived from Pichette (1976).

Discussion

The isotopic data for these three stocks (Table 33.1, Figs. 33.3, 33.4, 33.5) provide the first whole rock Rb/Sr isochron ages for the plutons within the Wabigoon Greenstone Belt of Northwestern Ontario.

The Burditt Lake isochron (Fig. 33.3) relies heavily on aplite samples of probable deuteric parentage; the age therefore is a minimum age. However, the authors suspect that the duration of magmatic consolidation was much shorter than the isochron error limits.

Some geological error was suspected in sample 40 of the Taylor Lake isochron, therefore this sample was not included in the regression treatment. All 6 points regressed give York I parameters of: 2654 ± 118 m.y. $R_o = 0.7007 \pm .0012$ and MSWD = 2.78. The five points used

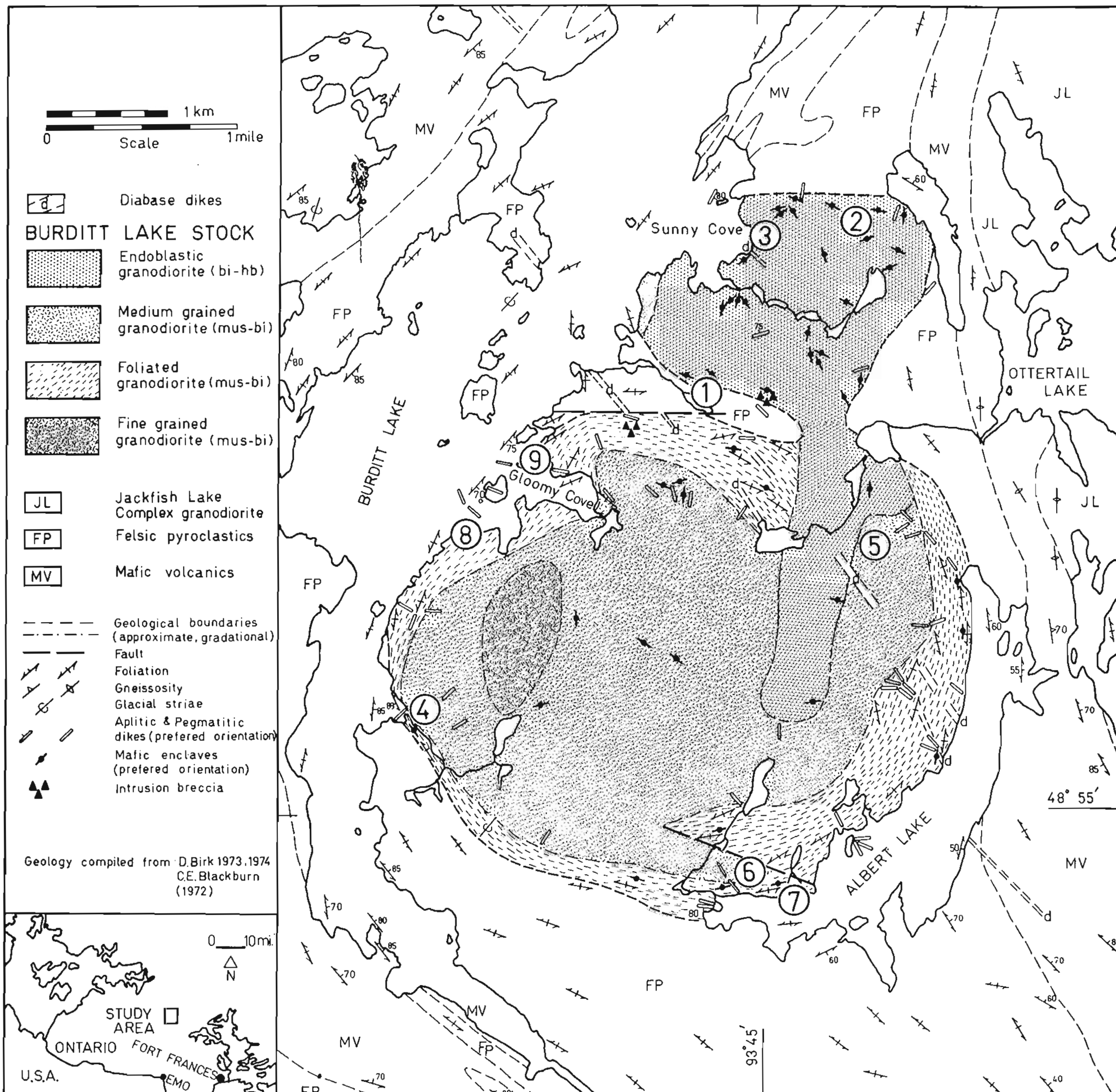

Figure 33.1. Geology of the Burditt Lake Stock, Northwestern Ontario.

in the accepted regression give: MSWD = 0.19 and much tighter errors. Sample 40 was collected from the west shore of Taylor Lake, straddling the Taylor Lake fault. The fault strikes north-northeast and causes a displacement of up to 1800 m, postdating all intrusive activity (Pichette, 1976). Fault induced disturbance of the Rb/Sr systematics is postulated.

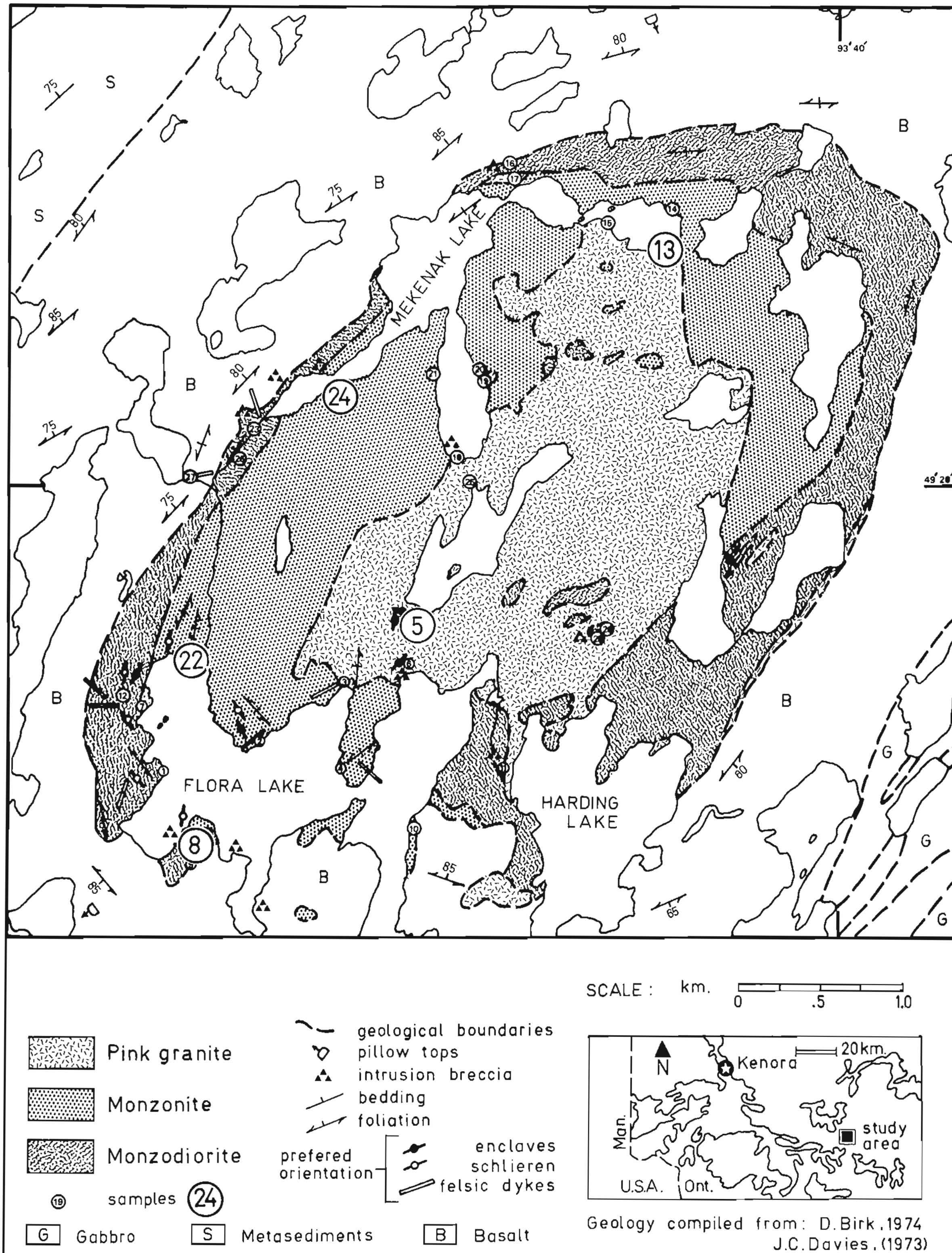

Figure 33.2. Geology of the Flora Lake Stock, Northwestern Ontario.

Table 33.1

Isotopic Data for the Wabigoon Belt plutons

Sample	Rb/Sr	Lithology	$^{87}Rb/^{86}Sr$	$^{87}Sr/^{86}Sr$ +
		BURDITT LAKE STOCK		
5	0.1583	granodiorite (musc-bi)	0.4587	0.7171
1	0.0635	granodiorite (bi-hb)	0.1839	0.7080
2	0.1049	granodiorite (bi-hb)	0.3039	0.7122
3	0.1089	granodiorite (bi-hb)	0.3153	0.7118
4	0.1756	quartz-eye aplite	0.3816	0.7154
6	0.4915	pink aplite	1.429	0.7549
7	0.6272	pink aplite	1.826	0.7677
8	1.255	pink aplite	3.679	0.8361
9	1.283	pink aplite	3.761	0.8383
		FLORA LAKE STOCK		
8	0.1006	foliated monzodiorite	0.2914	0.7124
22	0.2675	monzonite	0.7760	0.7311
24	0.3831	monzonite	1.113	0.7429
5	0.2175	granite	0.6306	0.7255
13	0.3121	granite	0.9057	0.7355
		TAYLOR LAKE STOCK		
42	0.0324	monzonite	0.0939	0.7041
36	0.1897	quartz monzonite	0.5497	0.7212
13	0.1280	granodiorite	0.3708	0.7141
57	0.2870	granodiorite	0.8325	0.7314
40*	0.2352	fine granodiorite	0.6821	0.7279
60	0.7370	microgranite	2.149	0.7809

* not included in regression treatment
\+ normalized

Mineral ages have previously been reported for the Flora Lake Stock by Heimlich (1963). Microcline Rb/Sr age limits (2840 ± 142 m.y., adjusted for γ = $1.39 \times 10^{-11}y^{-1}$) overlap with a biotite K/Ar age of 2660 ± 133 m.y. and when compared with the isochron age estimate suggest that the Flora Lake Stock has suffered no significant metamorphic re-equilibrations. The fact that both Taylor Lake and Flora Lake samples generate isochrons indicates that multiple intrusion occurred over a narrow time interval.

The low $^{87}Sr/^{86}Sr$ initial ratios for all three stocks indicate a source region having a low Rb/Sr ratio and further suggest that the isochron ages reflect the time of granitic emplacement rather than metamorphic ages (Jahn and Murthy, 1975).

Regional Geochronology

The foregoing age determinations are best compared with published Rb/Sr data from the Wawa Belt (Vermilion District) to the southeast and from

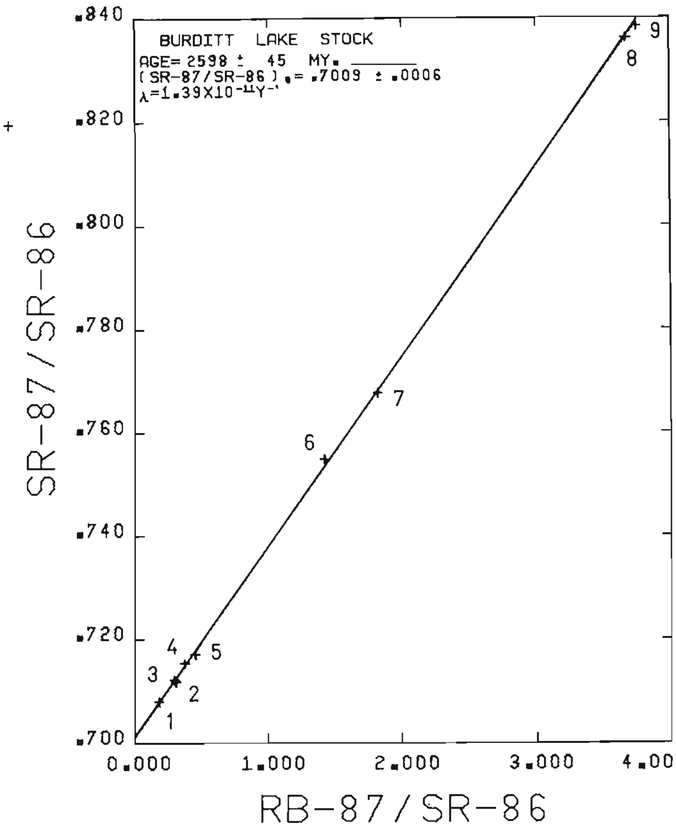

Figure 33.3. Whole-rock Rb/Sr isochron for the Burditt Lake Stock.

granitoids straddling the Wabigoon-Quetico Interface (Mackasey *et al.*, 1974, e.g. Rainy Lake area). All comparisons in this paper use Rb/Sr data that have been regressed by the York I method, using $\gamma = 1.39 \times 10^{-11}Y^{-1}$ and expressed with 95% confidence intervals, (Table 33.2).

Comprehensive Rb/Sr isotope data are available from the Vermilion District in the southwestern part of the Wawa Belt (Jahn and Murthy, 1975; Hanson *et al.*, 1971, Prince and Hanson, 1972). The Icarus Pluton, Giants Range Granite and the Vermilion Granite have age limits that do not overlap with the Burditt Lake isochron age limits. Field associations show that the Icarus Pluton intrudes the Saganaga Tonalite which in turn intrudes the Northern Light Gneiss (Jahn and Murthy, 1975). This suggests that the Wabigoon Belt late- to post-kinematic granitoid plutonism covered a duration that postdated the last granitoid emplacement in the Vermilion District.

Recent mapping (Harris, 1974) in the Rainy Lake region demonstrates that the "Algoman" samples of Peterman *et al.* (1972) come from four stocks whose geometric centres lie on a straight line trending north-northeast: the Rest Island stock, the Bears Passage stock, the Knuckle Island stock and the Ottertail Lake stock (Harris, 1974). If his isotopic samples are separated on the basis of cogenetic suites, two isochrons can be calculated, one each for the Rest Island stock

Table 33.2

Comparison of whole-rock isochrons

PLUTON	No. Pts.	AGE (±2σ)	R_o (±2σ)	SOURCE
Wabigoon-English River Gneiss Interface				
granodiorite gneiss	8	2664 ± 50 m.y.	0.7001 ± .0014	Farquharson and Clark (1971)
Caddy Lake Qtz Monzonite	7	2558 ± 12 m.y.	0.7087 ± .0015	" "
Whiteshell granodiorite	7	2610 ± 113 m.y.	0.7071 ± .0038	" "
" "	4	2594 ± 9 m.y.	0.7098 ± .0004	" "
Wabigoon Belt				
Burditt Lake Stock	9	2598 ± 45 m.y.	0.7009 ± .0006	this study
Flora Lake Stock	5	2640 ± 81 m.y.	0.7017 ± .0008	" "
Taylor Lake Stock	5	2640 ± 31 m.y.	0.7005 ± .0003	" "
Quetico-Wabigoon Interface (Rainy Lake)				
Ottertail Lake Stock	6	2471 ± 115 m.y.	0.7021 ± .0007	Peterman *et al.* (1972) b
Rest Island Stock	3	2455 ± 104 m.y.	0.7024 ± .0007	" " b
Bad Vermilion Tonalite	4	2520 ± 42 m.y.	0.7037 ± .0008	Hart and Davis (1969)
Wawa Belt (Vermilion)				
Northern Light Gneiss	7	2704 ± 123 m.y.	0.7007 ± .004	Hanson *et al.* (1971) a
Saganaga Tonalite	6	2721 ± 147 m.y.	0.7009 ± .0002	" " a
Icarus Pluton	4	2690 ± 21 m.y.	0.7008 ± .0001	" " a
Giants Range Granite	7	2670 ± 18 m.y.	0.7002 ± .0005	Prince and Hanson (1972) a
Vermilion Granite	11	2700 ± 50 m.y.	0.7004 ± .0003	Jahn and Murthy (1975)

a. Recalculated by Jahn and Murthy (1975)

b. Recalculated using blanket errors $^{87}Rb/^{86}Sr$ = 2% $^{87}Sr/^{86}Sr$ and omitting sample RL3766 = .05%

and the Ottertail Lake stock (Table 33.2). The resultant isochron parameters – although one is based on only three points – suggest that plutonism in the Rainy Lake area was significantly later than the last granitoid activity in Vermilion District. Error bars overlap with the Burditt Lake and Flora Lake isochrons – but the ages are significantly younger than the Taylor Lake isochron age. This suggests that Rainy Lake granitoids postdate the Wabigoon plutonism.

The Bad Vermilion Lake "Laurentian Tonalite" of Hart and Davies (1969) and Peterman *et al.* (1972) should more correctly be classed with the "Algoman" plutons. Tanton (1936) considered this pluton as "Laurentian" but was later corrected by local gold prospectors who conscientiously trenched clean an outcrop for him, showing unequivocally that this auriferous tonalite intrudes the Seine Conglomerate (unpublished data, Resident Geologist's Office, Ontario Division of Mines, Kenora). The isotopic data supports these prospectors since the isochron age (2520 ± 42 m.y.) corresponds with the age of Rainy Lake "Algoman" plutonism. The authors do not feel that it is necessary to evoke a concept of whole rock isochron resetting (Hart and Davis, 1969; Peterman *et al.*, 1972) for either the Rainy Lake or the Bad Vermilion plutons. The younger isochron ages for Rainy Lake and Bad Vermilion Lake are correlated with the location along the Wabigoon-Quetico fault interface. Granitoid plutonism may have been syntectonic to late-tectonic and roughly contemporaneous with the Quetico fault development. Mackasey *et al.* (1974) mention that several small mafic to ultramafic bodies lie marginal to the Quetico fault between Steep Rock Lake and Lac des Mille Lacs. The anorthosite unit at Bad Vermilion Lake therefore may also be contemporaneous with faulting which would explain concordancy on the pooled isochron of Hart and Davis (1969). If the above interpretation is correct, plutonism within the Wabigoon Belt and the Wawa Belt was followed by plutonism structurally controlled by the Wabigoon-Quetico interface.

The data from Farquharson and Clark (1971) for the Wabigoon-English River Belt interface (Table 33. 2) do not show such younger ages but, as Mackasey *et al.* (1974) point out – this boundary is only tenuously defined – whereas the Quetico Fault is a prominent structure visible even on ERTS imagery and airphotos.

Late-kinematic granitoid plutonism in the English River Gneiss Belt has yielded a U/Pb zircon age of 2660 ± 20 m.y. from a granite diapir intruding 3008 ± 5 m.y. old tonalitic gneisses (Krogh *et al.*, 1976). The authors suggest that late to post-kinematic granitoid plutonism for the Wawa Belt, the Wabigoon Belt and the English River Gneiss Belt spanned a period from 2550 to 2750 m.y.

It is interesting to note from Table 33. 2 that initial ratios are lower than 0. 702 for plutons intrusive into the Wabigoon and Wawa belts but are higher for plutons located at the belt interfaces. Nevertheless on a plot of initial ratios versus age as in Davies and Allsopp (1976), plutons from Wabigoon, Wawa and from the Wabigoon-Quetico Interface all plot within

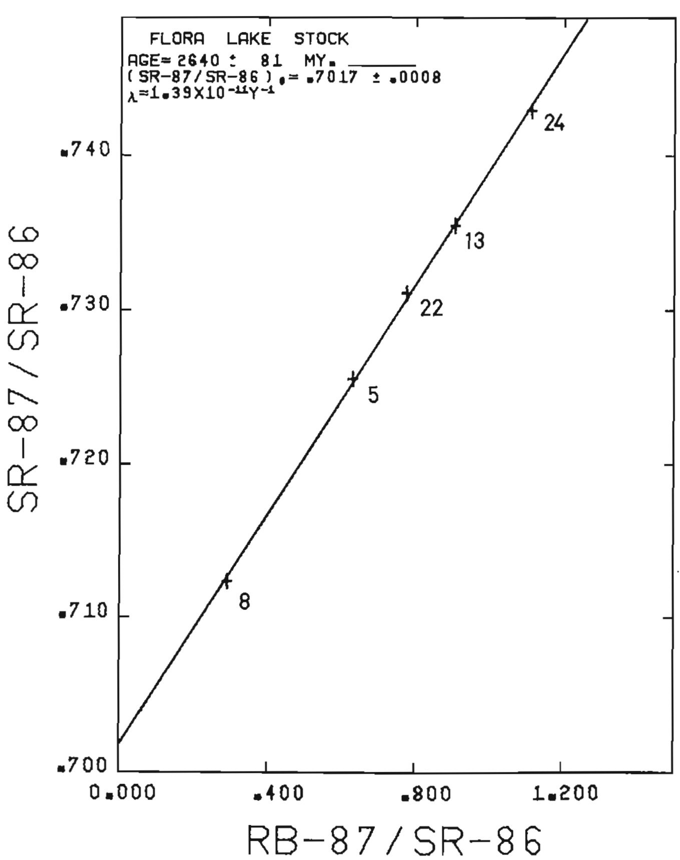

Figure 33. 4. Whole-rock Rb/Sr isochron for the Flora Lake Stock.

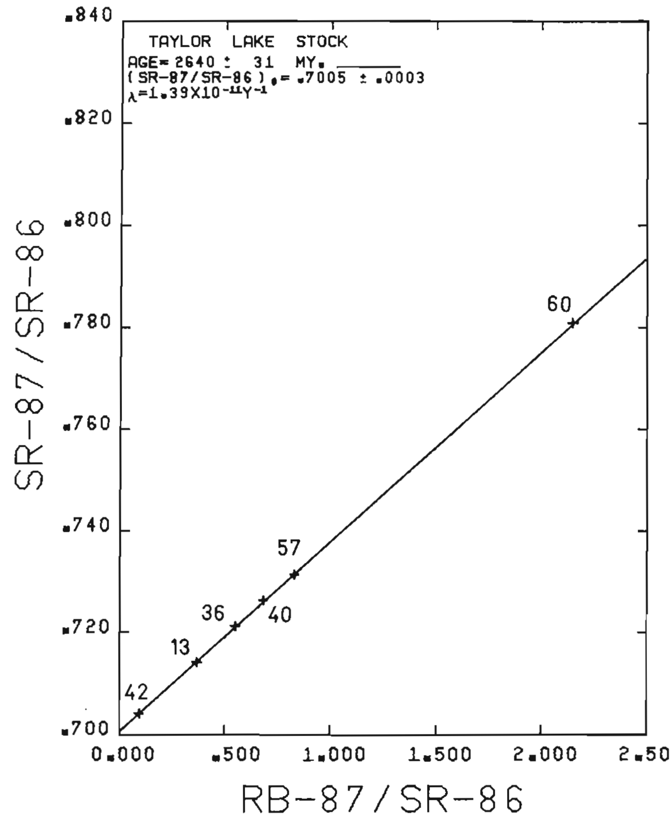

Figure 33. 5. Whole-rock Rb/Sr isochron for the Taylor Lake Stock.

or near the basalt source region. Therefore, late-tectonic granitoid plutonism was derived by partial melting of upper mantle material and not older sialic material. Arth and Hanson (1975) came to the same conclusion on the basis of trace element modelling.

Acknowledgments

Special thanks to R. Pichette for supplying Taylor Lake rock powders and field information.

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GEOCHEMISTRY OF ARCHEAN ROCKS FROM THE LAKE DESPAIR AREA, ONTARIO: A PRELIMINARY REPORT

E. M. R. Research Agreement 1135-D13-4-123

F. J. Longstaffe[1], R. H. McNutt[1] and H. P. Schwarcz[1]
Regional and Economic Geology Division

Introduction

This study forms part of a continuing project to determine the oxygen isotope distribution in Archean volcanic, sedimentary, granitic and gneissic terrains, to relate the oxygen isotopic data to other geochemical parameters, and to integrate the field, petrographic, geochemical and isotopic information into models of Archean crustal evolution. Some preliminary geochemical and isotopic results for the Lake Despair area are reported here.

Geological Background

The Lake Despair area occupies about 250 km^2 in the southern portion of the Wabigoon granite-greenstone belt (N48°55'; W93°40'; NTS ref.: 52C/12, 52C/13), and is about 40 km northwest of Fort Frances, Ontario.

Previous geological investigations include those by Lawson (1888) and Blackburn (1976), the latter work providing much of the geological background for this study.

Rocks in the study area have been divided into four major categories (Fig. 34.1):

1. interlayered mafic and felsic metavolcanic rocks
2. the Footprint gneiss
3. the Northwest Bay Complex
4. the Jackfish Lake Complex.

The mafic metavolcanic rocks increase in metamorphic grade from greenschist facies in the vicinity of Burditt Lake (about 10 km northwest of Lake Despair; not shown in Fig. 35.1; *see* Birk and McNutt, 1977) to amphibolite facies near the contact with the Jackfish Lake Complex. Ultramafic rocks, composed almost entirely of serpentine pseudomorphs after olivine, occur parallel to the regional foliation within the mafic metavolcanic pile.

Greenschist facies felsic pyroclastic rocks located in the vicinity of Burditt Lake grade along strike into quartz-feldspar-mica±garnet schists as both the Jackfish Lake Complex to the east and the Northwest Bay fault to the southeast are approached.

The Footprint gneiss is a relatively monotonous fine to medium grained, banded biotite trondhjemitic to granodioritic gneiss and migmatite (Streckeisen classification of rock names, 1967). Essential mineralogy is quartz (18-28%), oligoclase (51-71%), microcline (1-16%) and biotite (4-12%). Concordant and discordant pegmatoidal material is ubiquitous throughout. Numerous migmatized and strongly lineated amphibolite enclaves, many greater than 1000 m in length, as well as some enclaves of hornblendite have been observed within the gneiss.

The Northwest Bay Complex, a marginal phase of the Rainy Lake batholith, occupies the most southerly part of the study area and is separated from the other rock types by the Northwest Bay fault, a probable offshoot of the Quetico fault, located about 3 km to the south (Blackburn, 1976). The predominant lithology within the map-area is foliated to gneissic biotite granodiorite, but highly sheared biotite-muscovite granite is located near the fault zone.

The Jackfish Lake Complex is composed of two somewhat distinct rock suites, dioritic and granodioritic. Both are strongly foliated and contain numerous mafic enclaves; some enclaves within the dioritic rocks are up to a kilometre in length and contain remnant pillows (*see* Blackburn, 1976, p. 13).

The diorite suite contains meladiorite, diorite, monzodiorite, quartz diorite and leucodiorite (or oligoclasite). Volumetrically, diorite predominates. Essential mineralogy is hornblende (10-65%) and plagioclase (oligoclase to andesine; 35-90%). Minor phases include clinopyroxene, biotite, quartz, microcline, epidote, magnetite, chlorite and sphene.

The granodiorite suite contains quartz monzodiorite, granodiorite and microcline-albite syenite. Essential mineralogy is oligoclase, microcline, quartz, and hornblende in all but the syenite, which contains only a trace of quartz and has albite as the plagioclase feldspar. Abundant minor and accessory minerals include biotite, epidote, sphene, opaques, apatite, hematite, and white mica. Microcline occurs primarily as aligned patch and stringlet perthite megacrysts up to 20 mm in length. Petrographic observations suggesting late growth for the microcline megacrysts include:

1. aligned inclusions of altered plagioclase and unaltered plagioclase, each at different orientations within the same microcline megacryst;
2. abundant inclusions of plagioclase, quartz, hornblende, and sphene within the microcline megacrysts;
3. apparent embayment and pseudomorphic replacement of plagioclase by perthite megacrysts.

Other rock types found within the Jackfish Lake Complex are quartz-plagioclase dykes, aplites, pegmatites, and one occurrence of an albitite dyke.

The age relationships between the various rock units in the Lake Despair area are poorly known. The Burditt Lake granitic stock, which intrudes the Burditt

[1]Department of Geology, McMaster University, Hamilton, Ontario L8S 4M1.

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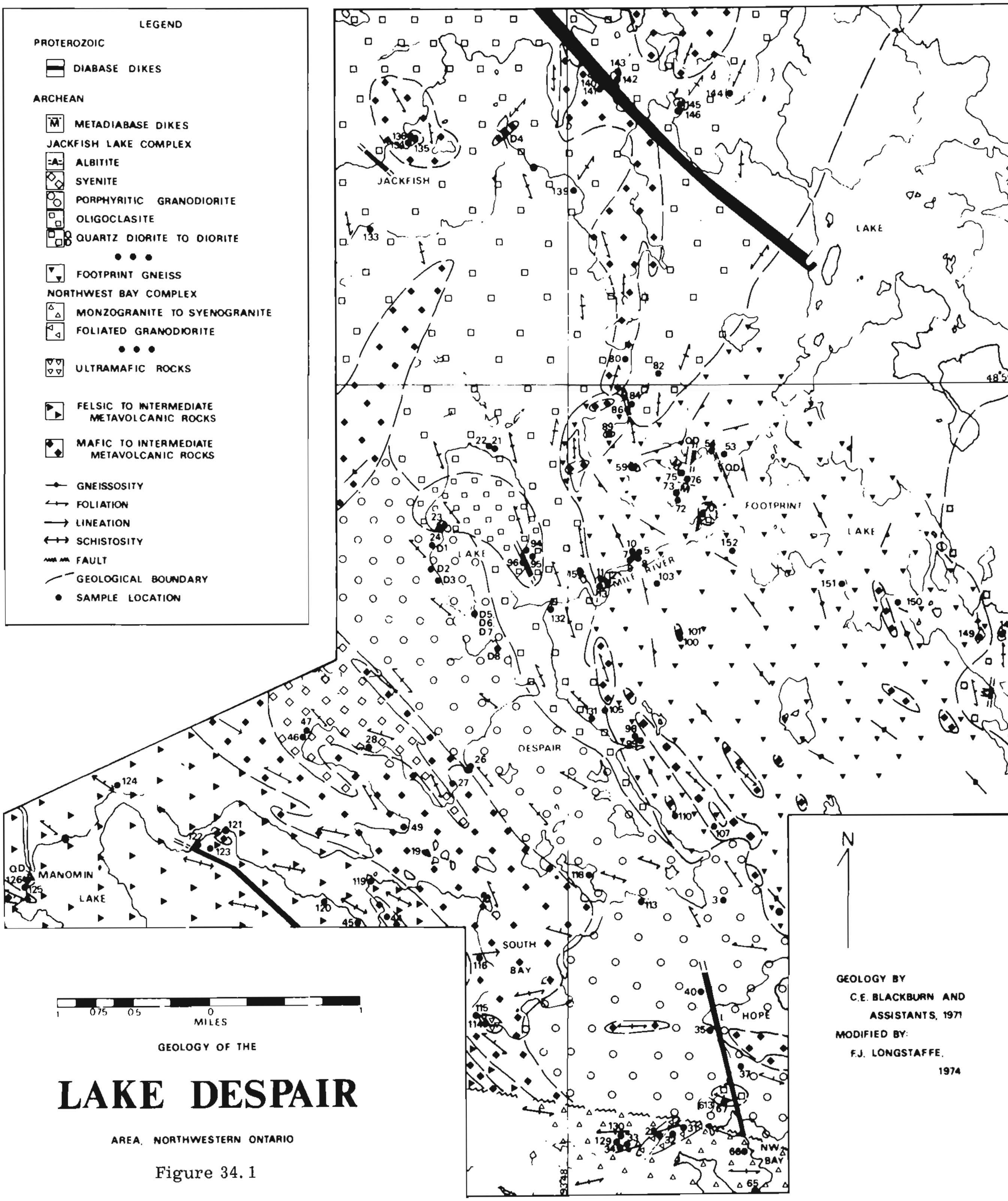

GEOLOGY OF THE

LAKE DESPAIR

AREA, NORTHWESTERN ONTARIO

Figure 34.1

Table 34.1

Average chemical results, tholeiites*

	1	2	3	4	5	6
SiO_2	52.95	52.74	49.50	53.75	51.40	51.07
TiO_2	.92	.56	1.01	1.03	1.41	1.02
Al_2O_3	14.27	14.85	15.20	14.99	14.03	14.99
FeO	11.56	9.47	12.78	11.17	12.62	12.44
MnO	.20	.19	.24	.21	.23	0.22
MgO	7.48	7.88	6.36	5.70	6.81	6.84
CaO	9.78	12.18	11.45	11.05	8.53	11.20
Na_2O	2.41	1.73	2.80	1.92	2.83	2.11
K_2O	.33	.26	.50	.12	.30	.11
P_2O_5	.10	.14	.16	.06	.14	
Rb	6	5	3† 5††		5	
Sr	111	81	209 139	98	150	
Ba	114	76	156 82		70	
K/Rb	448	427	1356 880		481	
Sr/Ba	.974	1.066	1.34 1.70		2.14	
Rb/Sr	.054	.062	.014 .036		.03	

1. Av. 5 greenschist facies metabasalts, Burditt Lake area (least altered).
2. Av. 3 amphibolite facies metabasalts, Lake Despair area (least altered).
3. Av. 7 amphibolite enclaves (over 1000 m in length) from within the Jackfish Lake Complex and the Footprint gneiss.
4. Av. 4 metabasalts, Off Lake-Burditt Lake area; C. Blackburn (1976).
5. Av. 5 tholeiites, Vermilion District; Arth and Hanson (1975).
6. Average mafic metavolcanic, Wabigoon greenstone belt; Goodwin (1976).

* All analyses recalculated to 100%, anhydrous, total Fe as FeO.

† Enclaves within Jackfish Lake Complex.

†† Enclaves within Footprint gneiss.

Major and minor element oxides in wt. %; Rb, Sr and Ba in ppm.

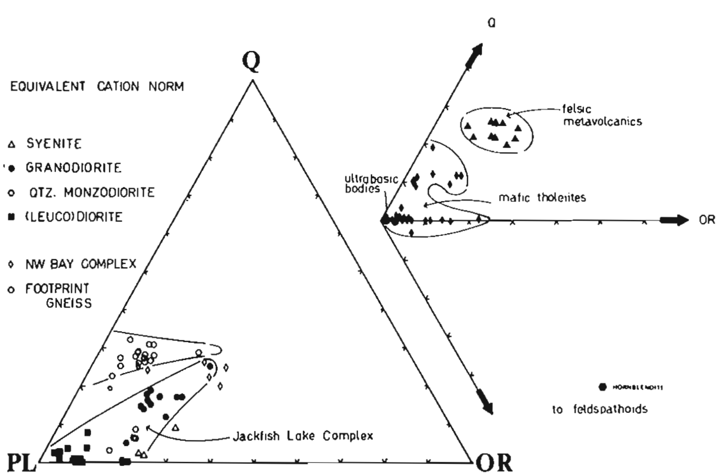

Figure 34.2. Normative quartz-plagioclase-orthoclase-feldspathoids diagram for various lithologies from the Lake Despair area.

Lake volcanics, has been assigned a Rb-Sr whole rock age of 2.598±0.045 billion years (2σ), (Birk and McNutt, 1977). Quartz diorite and leucodiorite dykes of Jackfish Lake Complex aspect cut both the mafic amphibolites and the Footprint gneiss. Within the study area, the Jackfish Lake Complex-Footprint gneiss contact is often marked by mafic breccias with a diorite infilling. Elsewhere the two units have gradational contacts. All units are cut by Proterozoic diabase dykes, which in turn have been fault offset. It appears that the Jackfish Lake Complex and the Burditt Lake stock are the youngest major Archean units within the map-area; the age relations between the Footprint gneiss, the Northwest Bay Complex and the metavolcanic rocks are uncertain.

Geochemistry

The compositions of all samples have been plotted as a function of normative quartz, plagioclase, orthoclase

Table 34.2

Comparison of $\delta^{18}O$ wholerock and LIL element abundances in the mafic tholeiites

Group	Sa. No.	$\delta^{18}O$	K(wt. %)	Rb(ppm)	Ba(ppm)	Sr(ppm)
greenschist tholeiites	B 1	7.0	0.10	2	44	135
	B 3	7.5	0.48	10	87	62
	B 5	9.0	0.29	9	142	121
	B 6	8.4	0.24	5	53	113
	B 8	9.2	0.23	4	244	124
least altered mafic amp.	F 19	5.5	0.20	3	74	76
	F 27	8.1	0.19	5	85	66
	F 115	7.5	0.25	7	68	102
altered mafic amp.	F 44	5.7	1.14	41	280	305
	F 49	6.0	0.76	41	139	182
	F 116	4.4	0.87	53	105	138
large mafic enclaves	*D 4	5.6	0.31	4	146	137
	*F 134		0.42	4	175	208
	*F 80	4.5	0.40	2	142	306
	*F 145		0.51	3	161	184
	**F 107	5.7	0.46	5	92	155
	**F 15	6.4	0.51	4	89	139
	**F 100	5.6	0.46	6	64	123
small mafic enclaves	*F 22		1.62	79	580	1376
	*D 6		2.67	71	2251	1435
	**F 105		0.49	11	92	158
ultrabasic rocks	B 2	5.3	0.02	2	16	5
	B 4	6.4	0.01	1	19	7
	B 7	6.8	0.01	<1	18	5
ultrabasic pods	*F 59		0.99	33	476	417
	*F 89		3.64	202	1007	305

*within Jackfish Lake Complex
**within Footprint gneiss.

and feldspathoids (Fig. 34.2), using the modified CIPW-Niggli equivalent cation norm of Shaw (1969). As mentioned, plutonic rock names have been assigned in accordance with I.U.G.S. nomenclature (Streckeisen, 1967) and aphanitic rocks after Irvine and Baragar (1971).

1. Mafic metavolcanic rocks and enclaves

The mafic metavolcanic rocks are tholeiitic (Fig. 34.3), and the average composition of least altered samples agrees well with results from other studies (Table 34.1). The mafic enclaves (Table 34.1, no. 3) are somewhat lower in SiO_2 and MgO and higher in Σ Fe (expressed as FeO) than the spatially related volcanics. The enclaves also have higher K/Rb (>880) than the volcanics (427-448). Enclaves from the Jackfish Lake Complex are also enriched in Sr and Ba, reflecting contamination by the Jackfish Lake Complex magma. For example, some small enclaves have Sr and Ba in excess of 1400 and 2250 ppm respectively (Table 34.2, F 22, D 6).

Oxygen isotope whole rock results for the mafic metavolcanics and enclaves range from $\delta^{18}O_{SMOW}$ = 4.4 to 9.2 ‰ (reported in the usual del (δ) notation relative to Standard Mean Ocean Water, SMOW; Table 34.2). The enclaves, including those enriched in Sr and Ba, have $\delta^{18}O$ values within the normal, unaltered basalt range (Taylor, 1968) despite their metamorphic nature. Some mafic amphibolites (F 44, F 49, F 116), also enriched in K, Rb, Sr and Ba, have seemingly undisturbed oxygen isotopic compositions as well. In these examples, any post-formation trace element migration and/or exchange has left the oxygen isotopic whole rock systems undisturbed. Other mafic amphibolites (F 27, F 115), considered least altered on the basis of their K, Rb, Sr, and Ba contents, have enriched $\delta^{18}O$ values. Systematic enrichment in heavy oxygen has been observed in the greenschist facies tholeiites from Burditt Lake (B 1, B 3, B 5, B 6, B8); $\delta^{18}O$ increases from 7.0‰ at the northern end of Burditt Lake to 9.2‰, near the Burditt Lake stock, approximately 3 km along

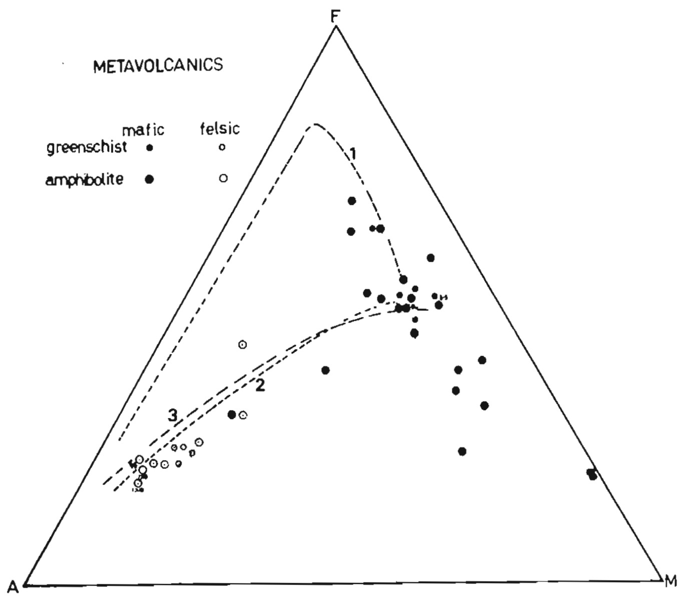

Figure 34. 3. AFM diagram for Lake Despair and Burditt Lake metavolcanic rocks; 1: Skaergaard tholeiitic trend; 2: calc-alkaline Cascade suite; 3: calc-alkaline lower California batholith; A = $Na_2O + K_2O$; Fe = FeO; Σ Fe as FeO; M = MgO.

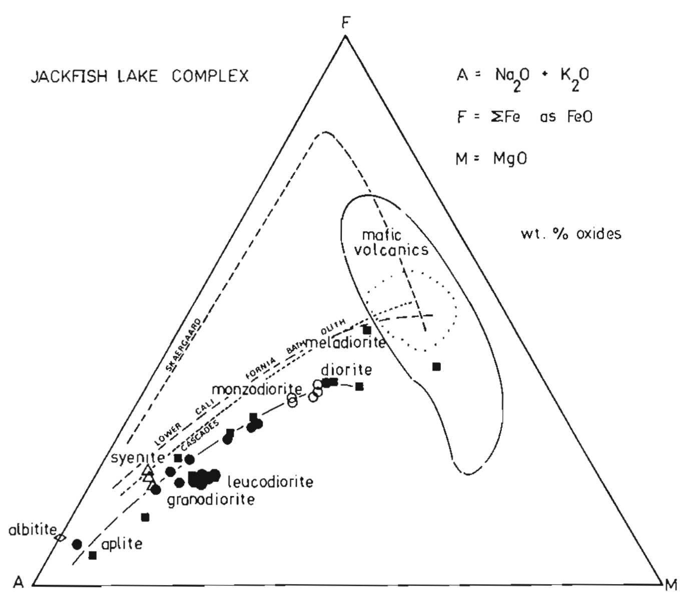

Figure 34. 4. AFM diagram for the Jackfish Lake Complex.

strike. The oxygen isotope results from the ultramafic bodies found within this sequence (B 2, B 4, B 7) have a parallel enrichment pattern. Further investigations are required before these oxygen isotope and trace element anomalies can be fully explained. It is pertinent to note, however, that low temperature alteration of modern ocean floor basalts results in ^{18}O enrichment, due to the formation of authigenic clay minerals. Basalts recovered from DSDP legs 34 and 37, for example, have all been affected by submarine weathering ($\delta^{18}O$ = 6 to 9‰), and are enriched relative to unaltered ocean floor basalts (5.7‰) (Muehlenbachs and Clayton, 1976).

2. Felsic metavolcanic rocks

The dacitic metavolcanic rocks, often interstratified with the tholeiites, are chemically similar to other felsic metavolcanic rocks from the Wabigoon greenstone belt (Table 34.3). The compositional gap between the mafic and felsic metavolcanics appears to be real (Fig. 34.2), suggesting bimodal basaltic-dacitic volcanism within the study area.

Whole rock oxygen isotope values for the amphibolite facies felsic metavolcanics are 7.3 to 9.4‰ (Table 34.4), and lie within the range observed for younger unaltered felsic metavolcanics (Taylor, 1968), as well as other Archean felsic tuffs (Longstaffe and Schwarcz, unpubl. Ms). Anomalously heavy values have been measured in some greenschist facies felsic metavolcanics near the Burditt Lake stock (B 10, B 11, B 12; Table 34.4); whole rock results for the granitic stock itself range from 7.8 to 8.2‰ in the south lobe to 8.6 to 9.2‰ in the endoblastic north lobe and associated microgranitic phases. Some relation between the intrusion of the stock and the enrichment of the metavolcanics is likely.

3. Footprint gneiss

Geochemically, there is little difference between the Footprint gneiss and the felsic metavolcanic rocks (Table 34.3, nos. 1, 4, 5, 6, 7). The gneiss does contain more normative quartz and less normative orthoclase (Fig. 34.2). Geologically, however, the location of the gneiss at the outer margin of the Rainy Lake batholith suggests a plutonic origin for this body. Chemical comparisons can be made equally well with the nearby foliated to gneissic granodiorite of the Northwest Bay Complex (Table 34.4, no. 3). Any possibility of a sedimentary origin for the gneiss can probably be ruled out by the oxygen isotope data (*see* below).

The oxygen isotope whole rock results range from 5.9‰ in migmatized grey gneiss to 8.9‰ in associated pegmatoidal material (Table 34.4). Much of the gneiss ranges from 7.5 to 7.8‰, isotopically too light for direct derivation from Archean clastic metasediments, which range from 8 to 13‰ (Longstaffe *et al.*, 1976a). Lighter values (5.9 to 7.1‰) have been measured in some portions of the gneiss; such areas also exhibit disequilibrium isotope mineral fractionations. Depletion in heavy oxygen has been documented in other high grade metamorphic terrains (Shieh and Schwarcz, 1976; Fourcade and Javoy, 1973; Longstaffe and Schwarcz, unpubl. Ms) and apparently accompanies upper amphibolite grade metamorphism. The possibility of isotopic depletion within the 'normal' (7.5 to 7.8‰) gneiss cannot be entirely ruled out, but isotopic mineral fractionations in these samples indicate an approach to equilibrium.

Table 34. 3

Average chemical results, tonalitic rock types*

	1	2	3	4	5	6	7	8
SiO_2	71.65	73.76	70.71	69.37	69.01	70.00	70.01	71.60
TiO_2	.27	.10	.27	.31	.37	.30	.37	.33
Al_2O_3	15.61	15.89	15.85	16.20	16.20	16.17	16.46	15.87
*FeO	1.91	.68	2.03	2.33	2.71	2.51	2.77	2.60
MnO	.03	.03	.03	.03	.05	.07	.04	0.06
MgO	.82	.30	.89	1.12	1.28	1.20	1.70	1.33
CaO	2.94	1.93	2.88	3.18	3.04	2.50	2.47	2.47
Na_2O	5.12	4.27	5.22	4.91	5.06	5.34	4.43	4.49
K_2O	1.53	3.78	1.99	2.39	2.18	1.82	1.68	1.26
P_2O_5	.12	.06	.14	.15	.10	.09	.07	
Rb	42	90	55	47	45			
Sr	555	461	708	571	506		313	
Ba	602	1173	617	513	451		367	
K/Rb	302	348	300	420	400			
Sr/Ba	.922	.393	1.15	1.11	1.12		1.17	
Rb/Sr	.076	.195	.078	.082	.089			

1. Av. 14 Footprint gneiss.
2. Representative mobilisate segregation, Footprint gneiss (F-110).
3. Av. 2 analyses, Northwest Bay foliated to gneissic granodiorite.
4. Av. 6 amphibolite facies felsic metavolcanic schists.
5. Av. 4 Burditt Lake greenschist facies felsic tuffs.
6. Av. 8 Burditt Lake greenschist facies felsic tuffs (D. Birk, pers. comm.).
7. Av. 4 Off Lake-Burditt Lake felsic tuffs; (Blackburn, 1976).
8. Av. Wabigoon felsic metavolcanic rock; (Goodwin, 1976).

*All analyses recalculated to 100% anhydrous, total Fe as FeO.
Major and minor element oxides in wt. %; Rb, Sr and Ba in ppm.

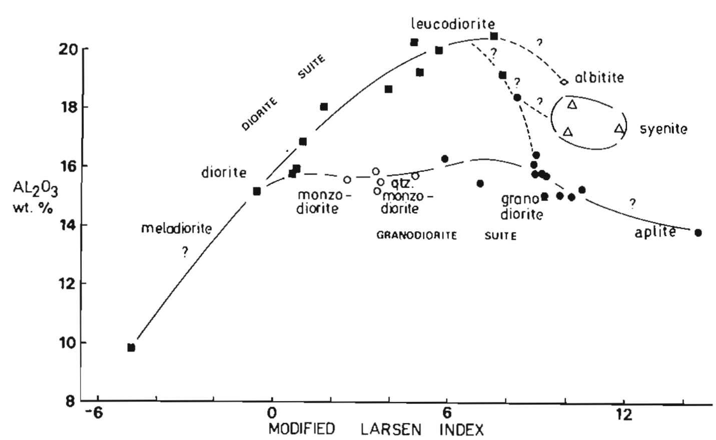

Figure 34. 5. Al_2O_3 versus Modified Larsen Index; variation diagram illustrating the diorite-leucodiorite suite and the quartz monzo-diorite-granodiorite suite.

3a. Origin of rocks of tonalitic composition

Even if the felsic metavolcanic rocks and the Footprint gneiss are not direct protolith equivalents, a common petrogenetic origin is reasonable in light of recent investigations on Archean rocks of tonalitic composition. Partial melting of rocks of basaltic composition, such as quartz eclogite, has been proposed by many (Hanson and Goldich, 1972; Arth and Hanson, 1972, 1975). Calculations similar to those made by these authors indicate that both the Footprint gneiss and the felsic metavolcanic rocks are compatible with formation by a maximum of 12 to 20 per cent partial melting of quartz eclogite. This calculation is based on the partial melting model of Shaw (1970) and enrichment factors for K, Rb, Ba, and Sr calculated against least altered tholeiites present in the study area. For low degrees of partial melting of quartz eclogite, Sr/Ba of both the melt fraction and the parent should be similar (Hanson and

Table 34.4

Some whole rock oxygen isotope results for the felsic rock types, Lake Despair area

Rock type	Sa. No.	$\delta^{18}O$	Rock type	Sa. No.	$\delta^{18}O$
felsic metavolcanics,	F 123	7.27	Jackfish Lake		
amphibolite facies	F 124	8.73, 8.93	Complex		
	F 45	9.21			
	F 119	9.37	i. meladiorite	F 139	6.37
				F 146	6.68
felsic metavolcanics,	424-2	9.42	ii. diorite	F 84	6.99
greenschist facies	434-1	7.87		F 21	6.69, 7.03
	G 89	9.59		D 8	6.96
	B 9	7.92		F 131	7.19
	B 10	10.01		F 144	7.26, 7.39
	B 11	11.39	iii. leucodiorite	F 94	7.54
	B 12	11.32		F 24	7.91, 8.24
			iv. qtz.-plag.	F 23	9.23
Footprint gneiss	F 103	7.79	dyke		
	F 10	7.59			
	F 73	7.72	v. qtz. monzo-	F 135	7.54
	F 152	7.45	diorite	F 82	7.72
	F 98	7.57		F 133	7.26
	F 99	6.83, 7.16	vi. granodiorite	F 113	7.61
	F 150	6.31		F 118	7.77
	F 151	6.05		F 26	7.48
	F 149	5.93		D 1	8.17
	F 7	6.09		F 3	8.24
	*F 110	8.85		F 40	6.58
				F 37	5.41
Northwest Bay Complex	F 129a	7.49, 7.95	vii. Na syenite	F 46	7.10
				F 28	6.49
granodiorite			viii. albitite		7.21
Burditt Lake	G 53	8.23			
granite	G 56	8.34			
	442-10	7.76			
	441-2	8.04			
	G 21	8.55			
	G 22	8.91			
	G 78	8.84			
	G 51	9.27			
	G 8	9.17			

*mobilisate within migmatized grey gneiss.

Goldich, 1972); values for the least altered tholeiites (1.0), the felsic metavolcanics (0.9), the Footprint gneiss (1.2) and the Northwest Bay Complex granodiorite (1.1) are consistent with such a requirement.

4. The Jackfish Lake Complex

The chemical variation within the Jackfish Lake Complex is illustrated in Figures 34.2 and 34.4; a summary of some chemical data is given in Table 34.5. The Complex shows similar trends to calc-alkaline suites such as the lower California batholith and the Cascades, but is somewhat poor in Σ Fe, expressed as FeO (Fig. 34.4). All samples range from alkali-calcic to calc-alkali, relative to the Peacock alkali-lime index.

Two samples of meladiorite do not fit the 'differentiation' curve; these are located only metres from mafic metavolcanic rocks, within a zone of mechanical mixing of volcanic fragments and dioritic magma, and thus are probably hybridized.

Chemical variation diagrams serve to distinguish the dioritic and granodioritic rock suites; Al_2O_3 versus the Modified Larsen Index, [(1/3 Si+K)-(Ca+Mg)], is one example (Fig. 34.5).

The Jackfish Lake Complex is of special interest as an example of Na-alkali enrichment similar to that recognized in a few other late Archean plutonic bodies within northwestern Ontario and northeastern Minnesota (Arth and Hanson, 1975; Longstaffe *et al.*, 1976b).

Table 34.5

Average chemical results*, Jackfish Lake Complex and similar rock types from the Vermilion District, northeastern Minnesota

	1	2	3	4	5	6	7	8	9	10	11	12
SiO_2	56.16	56.09	53.19	59.15	68.98	59.09	71.60	63.61	59.83	57.25	59.54	61.09
Al_2O_3	16.35	13.94	16.10	16.49	15.50	17.83	14.90	17.73	19.54	19.17	15.58	16.67
TiO_2	.68	.76	.83	.66	.24	.54	.28	.36	.38	.65	.62	.60
*FeO	7.09	6.14	8.79	5.50	2.20	5.26	2.02	3.02	3.38	5.79	6.17	5.46
MnO	.14	.10	.17	.14	.06	.10	.03	.07	.07	.09	.12	.10
MgO	5.37	6.08	6.22	3.94	1.40	4.60	.56	1.39	2.65	3.22	4.69	3.39
CaO	7.56	7.63	7.86	4.67	2.62	4.95	2.03	2.77	5.07	5.00	6.16	5.18
Na_2O	4.78	4.58	3.15	6.35	5.92	4.22	4.35	7.17	7.07	5.99	4.40	4.84
K_2O	1.54	4.08	3.69	2.77	2.97	3.11	4.13	3.65	1.11	2.64	2.45	2.39
P_2O_5	.33	.60		.33	.11	.31	.10	.23	.27	.21	.27	.28
Rb	31	102	101	55	72	74		51	14	40	71	76
Sr	1355	1871	886	501	1243	991		1941	2316	1465	1089	753
Ba	972	1725	1280	1082	1693	1071		1856	748	1928	1279	992
K/Rb	412	332	303	418	342	349		594	658	548	286	261
Sr/Ba	1.39	1.08	.692	.463	.734	.925		1.05	3.10	.760	.851	.759
Rb/Sr	.023	.055	.114	.110	.058	.075		.026	.006	.027	.065	.101

*All analyses recalculated to 100%, anhydrous; total Fe as FeO.

1. av. 7 monzodiorites and diorites, Jackfish Lake Complex (F 146, F 126, F 84, F 132, F 131, F 144, F 82).
2. syenodiorite, Icarus Pluton; Arth and Hanson, 1975.
3. syenodiorite, Vermilion Batholith; Arth and Hanson, 1975.
4. quartz poor granodiorite, Jackfish Lake Complex (F 35)
5. quartz rich granodiorite, Jackfish Lake Complex (F 118).
6. quartz poor granodiorite, Icarus Pluton; Arth and Hanson, 1975.
7. granodiorite, Vermilion Batholith; Arth and Hanson, 1975.
8. av. 2 Jackfish Lake Complex syenites (F 46, F 28).
9. av. 3 Jackfish Lake Complex leucodiorites (F 24, F 94, F 21).
10. av. 2 Farm Lake Na syenodiorites; Arth and Hanson (1975).
11. av. 3 quartz monzodiorites, Jackfish Lake Complex (F 133, F 135, F 136).
12. quartz diorite Giants Range; Arth and Hanson (1975).

Major and minor element oxides in wt. %; Rb, Sr and Ba in ppm.

Such enrichment is characterized best by the end member Na syenites and leucodiorites (*see* Table 34.5):

SiO_2	54-65%
Al_2O_3	17-21%
Na_2O	5-9%
Na_2O/K_2O	2-12%
Sr	>1500 ppm
Ba	>700 ppm; often >1500 ppm in syenites
K/Rb	>500
Rb/Sr	<0.03.

Combined modal data, whole rock trace and major element geochemistry, as well as mineral chemistry provide mass balance calculations indicating that the Na and alkali enrichment is controlled by the type and abundance of feldspar crystallized. Similar Archean rocks occur in the Vermilion District, northeastern Minnesota (Arth and Hanson, 1975); *see* Table 34.5. However, the Jackfish Lake Complex rock types, when compared against reasonably similar representatives from the Icarus Pluton, the Vermilion Batholith, and the Giants Range Batholith of northeastern Minnesota, are even higher in Na/K and lower in K.

Whole rock oxygen isotope results for a wide range of compositions within the diorite suite range from 6.4 to 8.2 ‰ for major lithologies, and have a positive correlation with increasing values of the Modified Larsen Index (Table 34.4). A late stage plagioclase-quartz dyke (F 23) is the most enriched member of the suite (9.2‰). The albitite dyke, on the other hand, has a $\delta^{18}O$ whole rock value of 7.2‰, similar to unaltered microcline-albite syenite (F 46; 7.1‰).

Samples of the megacryst-bearing granodiorite, which are very similar in chemical composition, range in $\delta^{18}O$ from 5.4 to 8.2 ‰. The isotopically lightest sample (F 37) has been highly altered (development of chlorite, saussuritization of plagioclase), and the quartz, microcline, plagioclase, hornblende and brown mica phases are completely out of oxygen isotopic equilibrium. The remainder of the granodiorite samples analyzed show at least an approach to isotopic equilibrium as the mineral phases (with the exception of microcline), are progressively enriched in ^{18}O in the normal order (Taylor, 1968), and show mineral isotope partitioning consistent with a final isotope equilibration temperature of 415 to 450°C (quartz-biotite and quartz-magnetite temperatures). Microcline does occur out of equilibrium in some samples; microcline-oligoclase fractionations vary from -0.6 to +1.8 ‰.

The somewhat low oxygen isotope temperatures may reflect final isotopic re-equilibration of the granodiorite during regional metamorphism. Another explanation would be the presence of a volatile phase during the later stages of crystallization (autometasomatism). Petrographic evidence suggesting late formation of the microcline perthites has already been presented; progressive crystallization of the megacrysts can be inferred from the behaviour of Ba, which decreases from about 13 000 ppm in the first formed megacrysts in the quartz monzodiorites to about 3000 to 7000 ppm in the much more abundant megacrysts of the granodiorite.

Such evidence, coupled with the oxygen isotope geothermometry, as well as the presence of microcline megacrysts within altered mafic enclaves, is consistent with the possibility of late magmatic or autometasomatic activity, resulting in the geochemical alteration and modification of the granodiorite suite.

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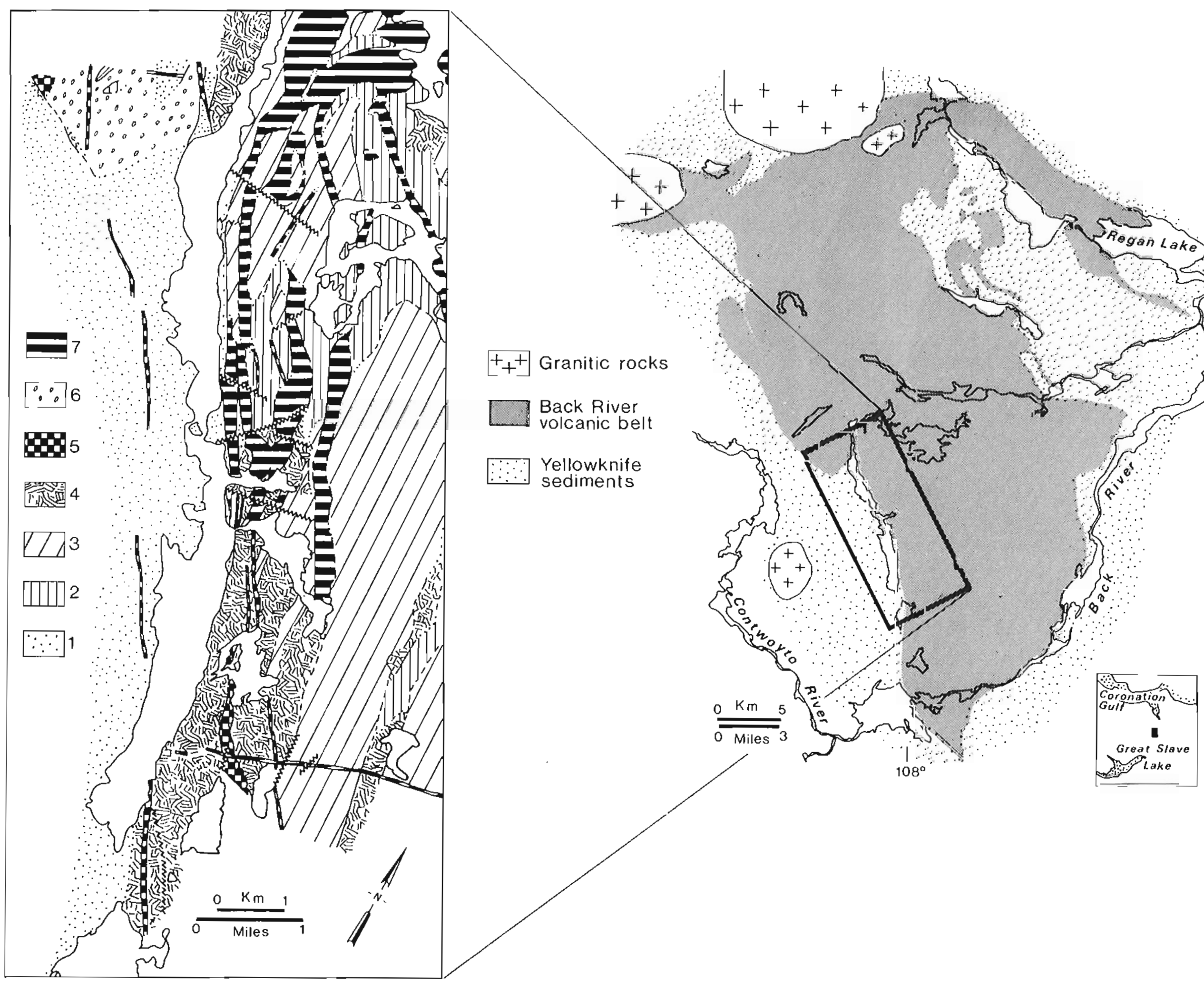
Project 740019

M.B. Lambert
Regional and Economic Geology Division

Introduction

The Back River Complex (Lambert, 1975) lies about 480 km northeast of Yellowknife, N.W.T. It comprises felsic to intermediate volcanic rocks, including tuffs, breccias and lavas, that generally lie conformably on greywackes and mudstones typical of the Yellowknife Supergroup of Archean age (Henderson, 1970). Basaltic rocks, almost invariably forming pillow lavas, make up a minor proportion of the complex. Plutons of granodiorite and adamellite and swarms of gabbro dykes intrude both sedimentary and volcanic successions. The dykes postdate folding of the sediments.

Two weeks of the 1976 field season were devoted to detailed mapping of the Complex along its southwestern margin (Fig. 35.1) with the purpose of (1) sampling felsic volcanic units for chemical analysis, (2) examining the nature of sediments and the boundary between sediments and the volcanics, and (3) determining the nature of large rhyolitic masses. Only the sediments (units 1 and 6, Fig. 35.1) and rhyolites (unit 4) are discussed in this report. A lake 15 km long provided easy access along the margin of the complex. W.N. Houston provided superb assistance during field operations.



1 - greywacke, shale, siltstone;
2 - dacite, minor rhyolite;
3 - andesite, minor dacite;
4 - rhyolite;
5 - basalt pillow lavas;
6 - rhyolite boulder conglomerate
7 - gabbro dykes

Figure 35.1. The Back River Volcanic Complex showing detail of mapping in 1976.

From: Report of Activities, Part A;
Geol. Surv. Can., Paper 77-1A (1977).

Sedimentary Rocks

The sedimentary succession (unit 1, Fig. 35.1) is dominantly greywacke with interbeds of shale and minor tuffaceous units. It is folded and metamorphosed to the greenschist facies. Near the contact with the volcanic complex and where sediments interfinger with volcanics, shales are graphite- and pyrite-bearing and locally associated with chert-jasper-hematite iron-formation. Locally pyrite forms nodules up to 3 cm across. Primary structures, such as graded bedding, cross-bedding, load and current structures, are well preserved in greywackes.

Shales appear to have absorbed most of the strain during three episodes of deformation. They display complex fold patterns that are contained between planar or gently undulating beds of greywacke. Generally, greywacke beds along the western side of the volcanic complex trend northwesterly and dip 70 to 85 degrees toward the southwest. Similarly, in sporadic outcrops along the eastern side of the lake, thin slivers of shale are intensely deformed against essentially undeformed, massive volcanics. Neither the fold style nor the penetrative structural features in the pelites are found in the volcanics. In this area, the drastic difference in deformation patterns obscures the precise nature of the contact between the Yellowknife sediments and the volcanic complex.

At the northwestern end of the area mapped in detail, rhyolite boulder conglomerate (unit 6) forms a thick wedge 2300 m across. This extremely massive unit contains pebbles, cobbles and boulders of rhyolite (and locally andesite) in a matrix of grey westhering grit. Subangular to well-rounded rhyolite boulders, ranging from 1 to 60 cm (averaging 5 to 10 cm) across, make up from 30 to 80 per cent of the rock. The unit is characteristically poorly sorted and non-bedded. Generally, fragments in the conglomerate are elongate and have a strong preferred orientation at 140 to 145 degrees. This orientation is regarded as a tectonic structure rather than a primary feature. Sandstones near the contact with the conglomerate contain rhyolite pebbles that decrease in size and abundance away from the conglomerate contact. Fragments in the conglomerate and the pebbly wacke are lithologically similar to the adjacent rhyolite mass (unit 4) from which they were derived.

Rhyolitic Masses

White to grey weathering felsic rocks (unit 4), mainly of rhyolitic composition but possibly including some dacitic rocks, form a dome northwest of the lake, a complex of flows about 1500 m wide by at least 8 km long along the southeastern side of the lake, and a narrow zone of breccia in the southeastern part of the area. The body northwest of the lake is massive except for a zone of brecciation along its western margin. In this rhyolite, albite phenocrysts (making up 10 to 15 per cent of the rock) lie in a matrix dominantly of albite microlites and quartz. Quartz occurs mainly in the matrix, and rarely as phenocrysts. Grain size (averaging 0.1 to 0.15 mm) of the matrix is coarser than the microfelsitic matrix of other rhyolites in the area. This body is interpreted as a dome that rose through sediments along the western margin of the volcanic complex and shed an apron of coarse detritus (rhyolite conglomerate) along its flanks.

Rhyolite southeast of the lake contains primary flow layering near the western margin and is variably massive, flow layered, brecciated and sheared in other parts. Autobrecciated rhyolite dominates this mass. This grey to white weathering rhyolite contains 5 per cent phenocrysts of quartz and albite up to 1 mm across, in a microfelsitic matrix. The rhyolite weathers pink adjacent to large gabbro dykes — apparently an effect of thermal alteration. Numerous zones of intense shearing mark faults in the unit. The shear zones generally contain abundant brown to grey weathering carbonate, disseminated pyrite, and locally galena, sphalerite and chalcopyrite mineralization. This mass may be a complex of lava flows and exogenous domes.

The rhyolite band, in the southeastern corner of the area mapped in detail, is a continuation of a prominent arcuate feature mapped to the north in 1975. The rock is mainly a tectonic breccia comprising rhyolite fragments in a grey to orange-brown weathering carbonate-rich matrix. The breccia forms almost vertical zones in massive autobrecciated rhyolite. In these zones, sheared breccia grades into brecciated and massive rhyolite bearing no carbonate. The breccia varies from an array of closely packed elongated blocks (ranging up to 10 by 60 cm) with carbonate in interstices to a carbonate-rich grit. Fragments are generally angular to subrounded, elongate slabs and slivers in the vertical plane, and rounded ovoid forms in the horizontal plane. The elongation defines a steeply plunging lineation.

Although in most places this rock is interpreted as a tectonic breccia, in at least one area well-bedded rhyolite pebble conglomerate and carbonate-rich sandstone indicate that it is partially of epiclastic origin. Movements along a system of vertical arcuate faults in the rhyolite body may have produced intense brecciation, differential grinding, and partial rounding of broken fragments. Carbonate, possibly related to volcanism, filled fractures and invaded crushed and brecciated areas. The epiclastic part of the unit may represent erosion from unstable fault scarps where this fracture zone reached the surface.

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36. STRATIGRAPHY AND SEDIMENTOLOGY OF THE ARCHEAN MANITOU GROUP, NORTHWESTERN ONTARIO

E.M.R. Research Agreement 1135-D13-4-123/76

P. Rae Teal[1] and Roger G. Walker[1]
Regional and Economic Geology Division

Introduction

The Manitou Group is a well-exposed sequence of volcanic and sedimentary rocks with few structural complications, lying along the southeastern edge of the Manitou Straits (Thomson, 1933) about 35 miles south of Dryden in Northwestern Ontario. Reports of abundant large-scale cross-bedding (Pettijohn, 1937, and pers. comm., 1971) suggested terrestrial sedimentation in some of these rocks, so it was decided to map them in detail to further the understanding of the Continental Association of Archean sedimentation (Turner and Walker, 1973). From mapping completed in 1972, 1973, and 1974 it has been possible to define stratigraphy, delineate several major environments of deposition, and establish paleocurrent directions. These will be used as a basis for determining paleogeography and basin evolution.

Most Archean sediments appear to belong to the Resedimented Association (Turner and Walker, 1973), mainly turbidites. Studies by Donaldson and Jackson (1965), Campbell (1971), Walker and Pettijohn (1971), McGlynn and Henderson (1970), Ojakangas (1972), Peeling (1974), and Henderson (1972) indicate the prevalence of turbidites in the Archean.

On the other hand, modern descriptions of shallow, agitated-water on terrestrial sediments are relatively rare. To date, the only well-documented terrestrial sediments are those of the alluvial fan deposits in the Sioux Lookout area (Turner, 1972; Turner and Walker, 1973) and the braided stream deposits near Kirkland Lake (Hyde, 1975). Because of the apparent rarity of occurrences of terrestrial sediments, and the paucity of descriptions of such occurrences, their role in Archean basin development is poorly understood.

It will be shown that a large portion of the Manitou Group is of terrestrial origin. Detailed mapping has clearly established the stratigraphic relationships of these terrestrial deposits, providing an excellent opportunity to assess their role in the development of the basin.

A corollary to the sedimentary interpretations is the problem which may be termed 'volcanic interpretations'; i.e., determining the environments of deposition of Archean pyroclastic and lava-flow rocks. In the Manitou area, both pyroclastic and flow rocks underlie and are in part laterally equivalent to the sedimentary pile. This close association permits an integrated approach to environmental interpretation, using the characteristics of both the volcanic and the sedimentary deposits to arrive at an interpretation compatible with both rock types. It also provides an opportunity to study the relationship between Archean volcanism and sedimentation.

[1]Department of Geology, McMaster University, Hamilton, Ontario.

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Stratigraphy

From the mapping, it has been possible to divide the volcanic and sedimentary rocks along the southwestern edge of the Manitou Straits into four units (Fig. 36.1). These units are referred to collectively as the Manitou Group (replacing "Manitou Series" of Thomson, 1933). A brief description of the stratigraphic units is presented here. Much fuller descriptions, with formal definitions of formations will be forthcoming (Teal, in prep.).

Underlying Rocks

Blackburn (1976) has divided the massive and pillowed basalts and their associated sediments below the Manitou Group into three informal units: the Beck Lake basalts, Etta Lake metasediments and basalts, and the Glass Bay basalts. They are not actually part of the Manitou Group, but rather form a base upon which the group rests. The Etta Lake metasediments are composed of volcanogenic tubidites and massive tuffs. In both the Beck Lake and Glass Bay basalts, pillowed lavas are nearly ubiquitous. They represent a large platform of submarine basalts, possibly 10 km thick, and at least 55 km long, upon which the Manitou Group was desposited.

Units of Manitou Group

Unit 1

The lowermost unit of the Manitou Group is the felsic pyroclastic Unit 1. It extends from the western end of the field area at Lower Manitou Lake, eastward as far as Sunshine Lake (Fig. 36.1). It ranges in thickness from 800 to 1600 m, being thickest at Cane Lake. The base of the unit appears to be generally conformable upon the Glass Bay basalts. At the eastern end of the area, Unit 1 intertongues with, and is conformably overlain by Unit 2. The southwestern part of the formation is conformably overlain by Unit 3b.

Unit 1 is dominated by thick, massive, heterolithic tuff-breccias, with minor tuffs. In a few places the tuff-breccias have only one or two clast types, but most of the unit is composed of pyroclastics with a wide assortment of clast types. At the southwestern end of the area, a small percentage of clasts of 'granite', quartz, iron formation, and chert occur among the otherwise volcanogenic clasts.

The tuff-breccias are composed of large, rounded to subangular clasts generally dispersed in a coarse tuff matrix. Sorting is poor. The maximum clast size is not uniform throughout the unit, being coarsest near Cane Lake and decreasing in size both east and west

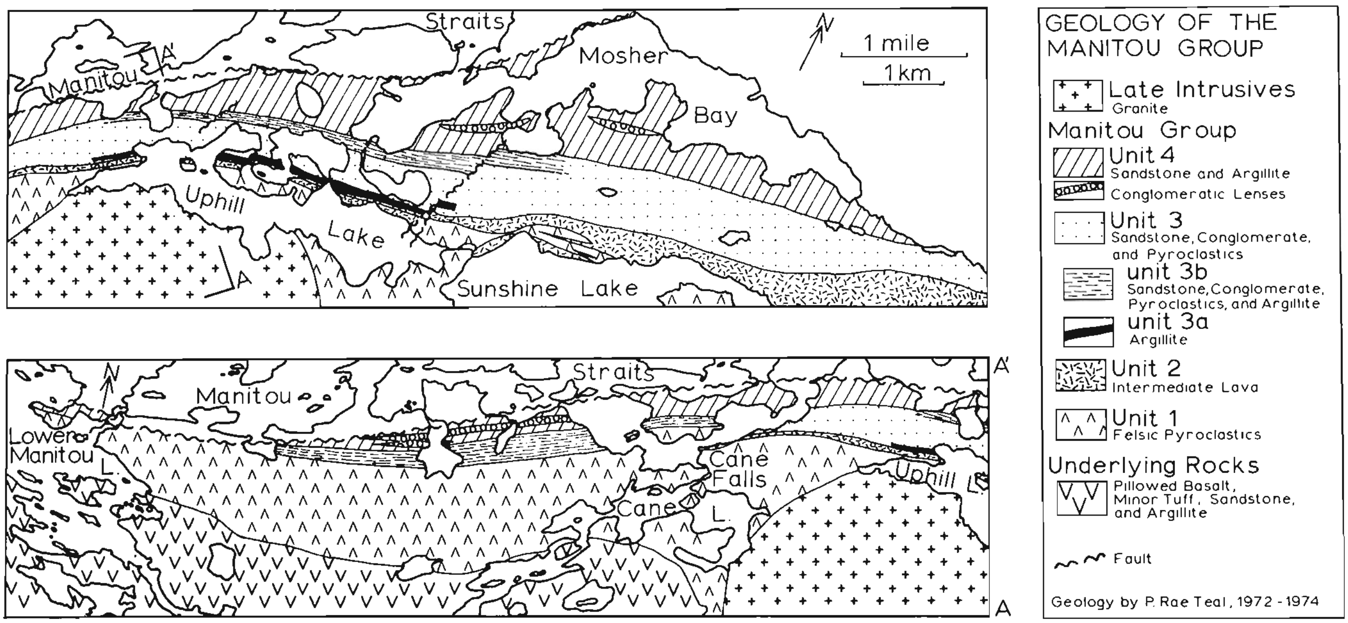

Figure 36.1. Generalized geology of the Manitou Group at Manitou Lake, Northwestern Ontario. Cane Falls is at 49°20'N, 92°49'W.

from there. This indicates that the volcanic pile is centred on Cane Lake, but no rocks indicative of a vent area are exposed at the present erosional level.

The tuff-breccia beds are thick (2 to 10 m) and most commonly show an irregular normal grading of the larger clasts; a few others are graded only in their top few centimetres. Many of the tuff beds are as thick as the tuff-breccias and equally massive. The thinner tuff beds (10 to 50 cm) are not graded and often faintly to distinctly laminated.

Unit 1 is believed to be subaerial mainly on the basis of negative evidence: there are no fine-grained 'background' marine sediments, and none of the sand-sized materials develop features like turbidites. Using Parsons' (1969) key for the recognition of volcanic breccias, it appears that many of the coarser pyroclastics are lahars: nonstratified, unsorted, heterolithologic, subangular to subrounded clasts, no recognizable volcanic cone structure, and some crude bedding. The essentially mono or dilithologic beds with much pumice are pyroclastic-flow breccias. The thinner, bedded tuffs are probably air-fall. Unit 1 is a subaerial volcanic pile built up by explosive volcanic activity and modified by associated volcanic and sedimentary processes.

Unit 2

Unit 2 is a group of intermediate, somewhat alkalic lava flows. The bulk of the unit lies directly north of Sunshine Lake (Fig. 36.1) where it is at least 550 m thick. A tongue of the unit tapering from 60 m at Surprise Lake to only a few metres at Cane Falls serves as a marker horizon separating Unit 1 and Unit 3. The top of Unit 2 shows no angular unconformity with the overlying Unit 3, but the base of Unit 3 is crowded with clasts derived from Unit 2, indicating reworking of loose debris from Unit 2 during initial Unit 3 deposition.

The unit is composed of generally massive, porphyritic, amygdaloidal lava. Both the phenocrysts of feldspar and amphibole and the quartz or calcite-filled amygdules are 2 to 5 mm in diameter. In many places flow breccias are developed: the porphyritic and amygdaloidal lava is broken into angular to subangular blocks 10 cm to several metres in diameter, and the spaces between blocks filled with a very fine grained massive lava apparently of similar composition to that of the blocks, but without amygdules or phenocrysts. A few sandstone and conglomerate beds of material derived from the flows are intercalated with the lava flows.

Unit 2 is interpreted to comprise a series of subaerial lava flows. None of the intercalated sediments could be construed as deep marine. There are certainly no fine grained "background" sediments. None of the lavas is pillowed, and there are no pillow-breccias. The lavas are intercalated with the subaerial pyroclastics of Unit 1 and conformably overlain by the terrestrial Unit 3. Some of the intercalated sediments are cross-bedded. Flow breccias which are "cemented" by the solidification of lava squeezed between the blocks are more likely to have formed on land.

Unit 3

Unit 3 extends virtually the entire length of the sedimentary belt, although southwest of Cane Falls it is represented only by Unit 3b (Fig. 36.1). The unit varies in thickness from 500 m to 1200 m, the thickest portion being east of Surprise Lake.

It has been possible to recognize two distinct subunits within Unit 3: Unit 3a, entirely enclosed within the formation, and Unit 3b, which forms the top of the main unit over most of its length. On the basis of bedding style, sedimentary structures, and composition, the remaining bulk of the unit can be further divided into three parts, each laterally equivalent to and transitional into those beside it: the Western, Central, and Eastern Areas.

The Western Area of Unit 3 is pyroclastic material identical in all respects to Unit 1. The distinction between these two units from about the western end of Uphill Lake to Cane Falls depends solely upon the presence of the tongue of Unit 2 lying between them. West of where this tongue ends, there is no longer any justification for two units. Since it is all pyroclastic material it is referred to Unit 1. In effect, the west end of Unit 3 is completely transitional into Unit 1 at Cane Falls.

The Central Area is composed entirely of volcanic debris, like the Western Area, but in contrast to it, clasts are more rounded, bedding thickness is only about 30 cm, bedding contacts are distinct but commonly not sharp, and there are a few occurrences of cross-bedding (15 cm sets). The Central Area is laterally equivalent to and transitional into both the subaerial pyroclastics of the Western Area and the braided fluvial deposits of the Eastern Area (*see* below), so it is constrained to a subaerial environment. Its compositional similarity to the Western Area, the general lack of fine grained sediments, the poorly sorted, dispersed pebble conglomerates, diffuse bedding contacts, and the rare cross-bedding all strongly suggest an alluvial fan origin for this area (following the detailed arguments of Turner and Walker, 1973).

Preliminary grain orientation and maximum clast size data indicate that the sediments of this area were deposited by eastward moving currents coming directly from the Unit 1 pyroclastic pile. Additional paleocurrent work is underway to further delineate the paleogeography of this volcanic terrane and its relationship with the associated sedimentation.

The Eastern Area differs considerably from the other two. It is not composed exclusively of volcanogenic material, but also contains granite, iron formation, chert, and quartz clasts in both the conglomerates and sandstones. The conglomerates are commonly finer grained than those to the west, and the sandstones are in the fine to medium sand range, as opposed to the coarse to very coarse sandstones to the west. The sandstones appear sorted and the conglomerates well rounded and close-packed. Large-scale cross-bedding with sets from 10 cm to 1 m thick, is abundant in both the sandstones and the conglomerates. Some outcrops are entirely trough cross-bedded. This almost ubiquitous occurrence of large-scale trough cross-bedding, especially in conglomerates, coupled with an almost total lack of fine grained material, suggests that the Eastern Area represents the deposits of a braided fluvial system.

Grain orientation studies indicate that flow in this area was approximately perpendicular to that of the adjacent alluvial fan.

Unit 3a occurs near the base of Unit 3 in the Uphill Lake area, and is a small, thin unit of laminated siltstone and argillite which is completely enclosed in coarser sediments of terrestrial environments. It is considered to represent a lacustrine deposit.

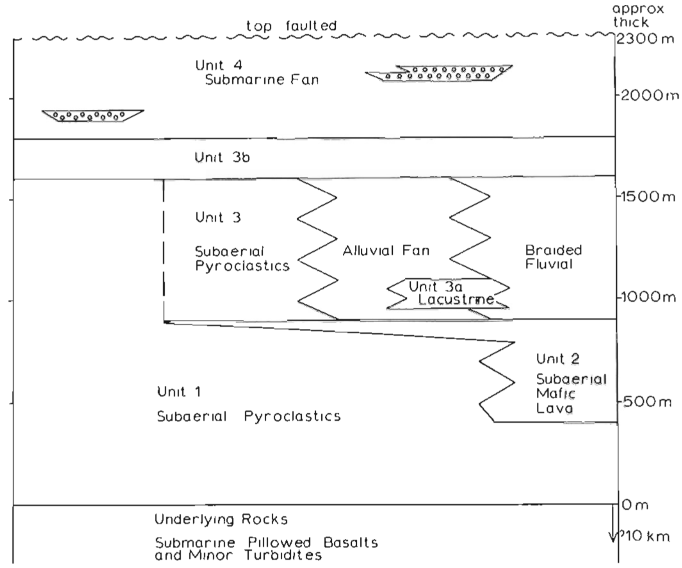

Figure 36. 2. A schematic representation of the relationship between stratigraphic units of the Manitou Group. Section runs parallel with strike of units in Figure 36. 1, from SW to NE.

Unit 3b is a highly variable group of siltstones, argillites, volcanogenic and quartzose sandstones, conglomerates, tuffs, and tuff-breccias occupying the top of the Uphill Lake Formation over most of its length. Unfortunately, the unit apparently is entirely without any diagnostic features in composition, stratigraphic organization, or sedimentary structures.

Unit 4

Unit 4 is a group of argillites, sandstones and conglomerates of the Resedimented Association (Turner and Walker, 1973) at the top of the Manitou Group (Fig. 36. 1). It is a minimum of 500 m thick, but its true thickness cannot be determined due to faulting and isoclinal folding in the upper part of the unit. The conglomerates contain quartz, chert, magnetic iron formation, 'granite' and mafic and felsic volcanic clasts. Compositionally, the sandstones are fine grained equivalents of the conglomerates. The argillite and sandstone form thick, monotonous sequences of graded beds which are clearly turbidites deposited in quiet water, below storm wave base. Associated with these turbidites are fairly thick accumulations of laminated argillite, indicating background sedimentation in a relatively deep, quiet marine environment. There are also a few thin magnetic iron formation occurrences.. The conglomerate, together with interbedded coarse, massive sandstone, forms a number of lenses within this general background of argillite and graded sandstones. The largest is 2. 8 km long and 200 m thick. They probably represent submarine channels cut into the finer sediments, probably on a submarine fan.

Interpretation

Initially, a thick and widespread base of submarine mafic lavas, with associated minor volcanogenic sediments was developed throughout and beyond the Manitou Lake area (Fig. 36.2).

At the western end of the area a felsic pyroclastic pile (Unit 1) began to grow and appears to have become emergent almost immediately. At about the same time, subaerial lava flows of intermediate composition (Unit 2) erupted at the eastern end of the area and these intertongued with the pyroclastics near the centre of the area.

Cessation of lava effusion suggests a break in the volcanic activity, but the pyroclastic pile in the west resumed its growth. As it did so, an alluvial fan of volcanic debris (the Central Area of Unit 3) developed on its flank (Fig. 36.2). This alluvial fan developed simultaneously with a braided fluvial system (the Eastern Area of Unit 3) which was flowing approximately perpendicular to the alluvial fan, and bringing some non-volcanic debris into the area. A relatively small, short-lived lake received some fine grained sediment (Unit 3a) early in the development of the alluvial fan and fluvial systems.

During, or immediately following deposition of the assorted facies of Unit 3b, the entire area was submerged, probably rather rapidly, and was covered by a submarine fan (Unit. 4).

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37. SEDIMENTARY ENVIRONMENTS AND THE EVOLUTION OF THE ARCHEAN GREENSTONE BELT IN THE KIRKLAND LAKE AREA, ONTARIO

E.M.R. Research Agreement 1135-D13-4-123/76

Richard S. Hyde[1] and Roger G. Walker[1]
Regional and Economic Geology Division

Introduction

Up to 3500 m of Archean sedimentary and alkaline volcanic rocks occur stratigraphically above a thick sequence of dominantly mafic volcanic rock in the Ontario portion of the Abitibi greenstone belt. These sedimentary and alkaline volcanic rocks were long known as the Timiskaming "Series" (Miller, 1911), although the name Timiskaming Group is more correct.

As well as geological mapping by the Ontario Division of Mines, important earlier studies on the Timiskaming and surrounding volcanic rocks are those of Thomson (1946), Bass (1961), Hewitt (1963), Cooke and Moorhouse (1969), Ridler (1970, 1976), Jolly (1974, 1975), and Jensen (mapping, summer 1976).

In contrast to this previous work, our efforts are directed towards determining depositional environments, their vertical and lateral arrangement, and establishing paleocurrent patterns. We also wish better to define the provenance of the Timiskaming Group.

The study area lies near Kirkland Lake and Larder Lake, Ontario, and, although the work focuses on this area, reconnaissance work for comparative purposes was carried out near Lake Timiskaming, Timmins, Matachewan, and northwestern Quebec (Fig. 37.1). Figure 37.2 is a simplified geological map of the area largely taken from Ridler (1970).

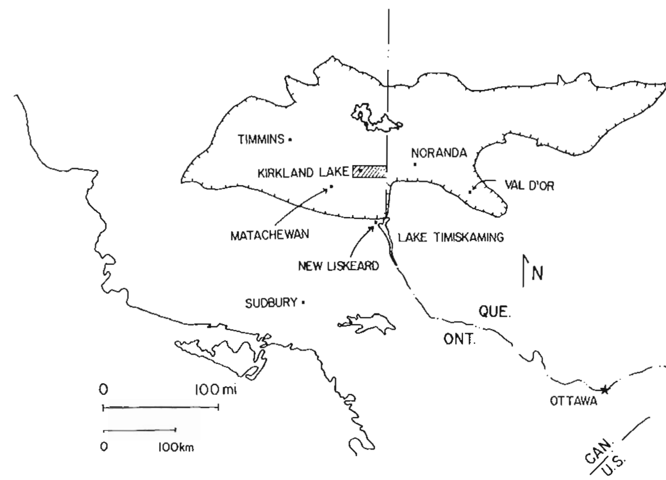

Figure 37.1. Abitibi greenstone belt. The study area is shown by the hatchured rectangle.

Sedimentary and Pyroclastic Facies

Sedimentary and pyroclastic rocks are here subdivided into 12 facies based on grain size, nature and thickness of bedding, type and size of sedimentary structures, and vertical sequence of sedimentary structures. Four of the sedimentary facies can be grouped into a Nonmarine Association, which is identical to the Continental Association of Turner and Walker (1973), but for which the informal name Nonmarine Association is proposed to avoid the implication that these sediments were deposited on continents.

Nonmarine Association

Facies A: This facies is characterized by polymictic conglomerates and medium- to coarse-grained sandstones with very small amounts of argillite. Large scale (> 10 cm in thickness) trough and planar cross-stratification and parallel lamination are common in the sandstones, and cross-stratification also occurs in the conglomerate. Contacts between conglomerates and sandstones commonly show evidence of scouring with grooves up to half a metre deep. Graded bedding is very rare in both rock types, and when present is associated with filling of scours. One section shows a 5-m-thick fining-upward sequence beginning with pebbly sandstone and terminating in laminated siltstones and argillites.

Clasts in the conglomerate range from pebbles to boulders in size, are generally ellipsoidal in primary shape, and are closely packed such that the clasts support themselves. Imbrication is common.

Comparison of these features with observations on modern day braided river systems (Williams and Rust, 1969; Collinson, 1970; Smith, 1970; Boothroyd and Ashley, 1975; amongst others) shows a close correspondence. In particular, the copious amount of conglomerate, lenticular bedding, scouring, cross-bedding in both sandstones and conglomerates, imbrication, and fining-upward sequence (channel fill, not point bar origin) are most consistent with a braided river origin. The braided rivers may have been developed in the distal portion of alluvial fans, but there is no compelling evidence that alluvial fans were present.

A glacial till origin for the conglomerates as favoured by McLean (1956) is discounted on the basis of cross-stratification in the conglomerates, scouring, cross-stratified and parallel laminated sandy lenses within the conglomerates, and lack of a muddy matrix and associated varves.

Facies B, C, D: These facies are much less extensive than facies A, but are all environmentally related to facies A, and consist of flood plain, lacustrine, eolian, and sandy channel deposits in braided rivers. They are discussed more fully in Hyde, (in prep.).

[1]Department of Geology, McMaster University, Hamilton, Ontario

From: Report of Activities, Part A; Geol. Surv. Can., Paper 77-1A (1977).

Resedimented Association

Facies E: This facies consists of interbedded sandstone, siltstone, and argillite, in which sandstone beds range in thickness from less than a centimetre to greater than one metre. Sandstones show sharp basal contacts with underlying argillites, are commonly graded, and together with the finer sediments display partial Bouma sequences. For these reasons, facies E beds are interpreted to be turbidites. High sandstone/argillite ratios (variable, but not less than 7:1), and high percentages of partial Bouma sequences beginning with the A division indicate near-source depositional environments which, because of associated conglomerates, were probably on submarine fans.

Facies F: Conglomerates interbedded with turbidites are either 1) in the granule-pebble range, closely packed, commonly graded, and well-imbricated, or 2) are in the pebble-boulder range, more dispersed in the matrix, poorly graded, and poorly imbricated. Argillite is present only in the form of clasts. The first type of conglomerate was probably emplaced by turbidity currents, and the second type by mass flow processes, probably within submarine fan channels.

Facies G: Facies G comprises very fine to medium grained sandstone, siltstone, and argillite, typically laminated and thinly bedded. Graded bedding occurs in approximately one third of the sandstone beds, which are also extensively cross-laminated. Association with facies E and F permits the inference that these are also relatively deep-water sediments, with deposition by turbidity currents (forming Bouma CDE or CE sequences) followed by varying degrees of reworking. Extensive soft-sediment deformation in some areas suggests deposition on levees or interchannel areas of submarine fans. Other examples of facies G may have been deposited in abyssal plains distal to the submarine fans.

Facies H: Facies H is not resedimented in a strict sense, and includes oxide iron-formations with or without chert, and graphitic horizons with associated sulphide minerals. Magnesitic carbonate rocks (L.A. Tihor, pers. comm.) also occur stratigraphically enclosed in the Resedimented Association, and if they are exhalative in origin as Ridler (1970, 1976) believes, they too would be included in facies H. The fact that facies H is closely related stratigraphically to other facies in the Resedimented Association argues that these sediments formed below storm wave base (Hyde, 1975) during periods of reduced clastic input or, perhaps, in areas where clastic input was minimal due to topographic barriers.

Pyroclastic Facies

Timiskaming volcanic flows are not treated as a facies *per se*, but are typically massive or flow banded and, in places, porphyritic with phenocrysts of feldspar or pseudoleucite. Emphasis is here placed on pyroclastic rocks, which have been subdivided into four facies (I-L).

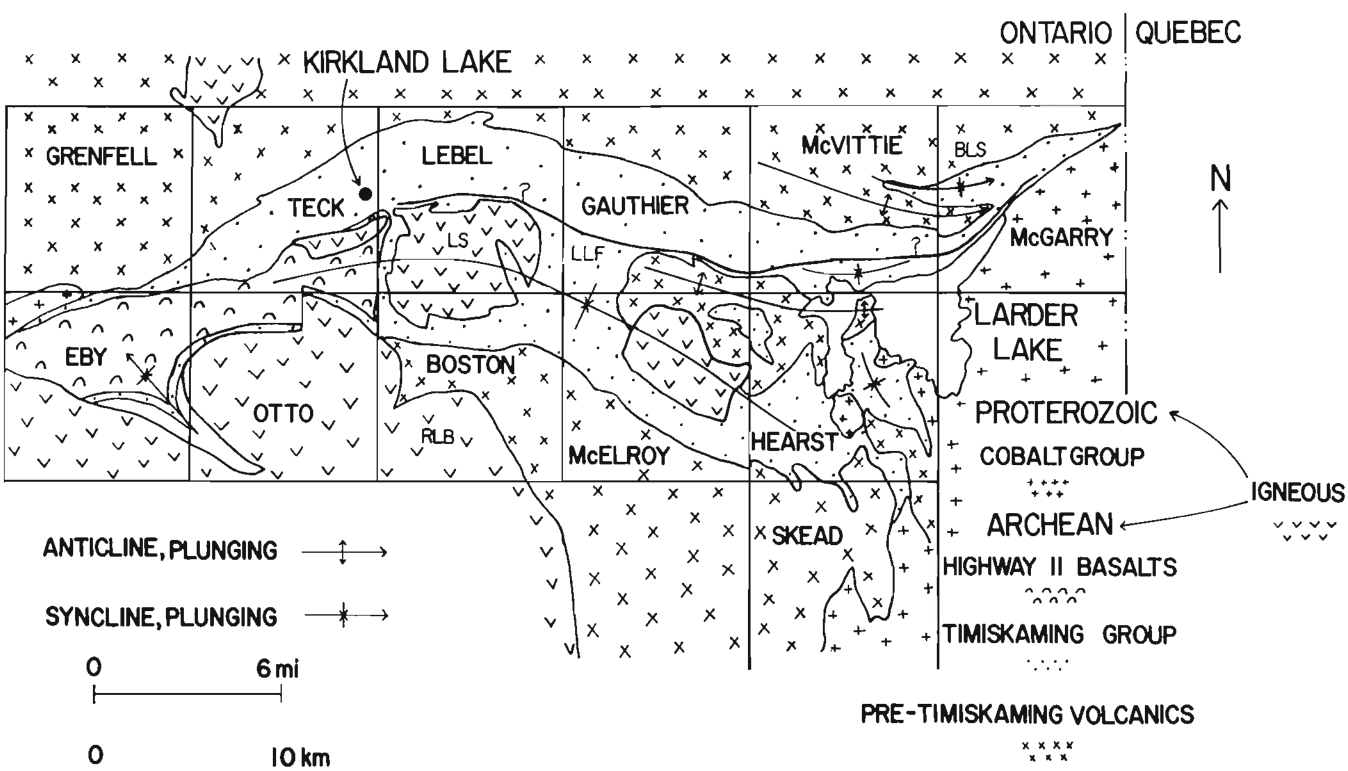

Figure 37.2. Simplified geological map of the Kirkland Lake-Larder Lake area largely from Ridler (1970).

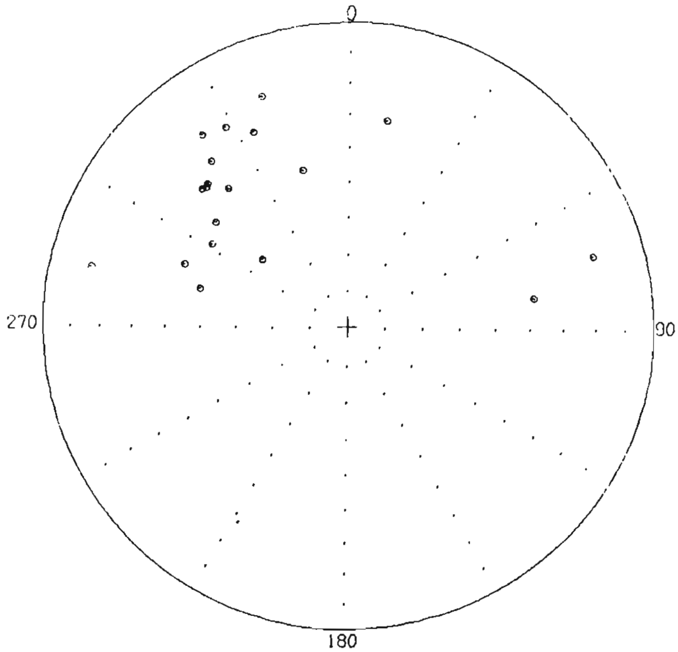

Figure 37.3. Stereoplot of the strikes and dips of the AB plane (plane formed by the intersection of long and intermediate axis) of individual clasts from a conglomerate bed after correction for folding. Points are projected onto the lower hemisphere and are not poles to the AB plane. In this example, (taken from a conglomerate bed in the Chaput-Hughes roadcut 1.6 km west of Kirkland Lake) paleoflow was towards the southeast.

Facies I: Thin bedded silty to lapilli tuffs and lapillistones, which are sometimes graded or inversely graded make up facies I. In some occurrences the thin beds are composed almost entirely of feldspar crystals which show graded bedding, and in which some of the crystals at the base of the bed appear to have sunk into the underlying layer. Facies I tuffs are interpreted to be of airfall origin.

Facies J: This facies comprises poorly sorted, thick bedded, massive tuffs, lapilli tuffs, tuff breccias, and agglomerates. Most of these beds are thought to represent non-welded ash flows, but some of them may be of lahar or airfall origin.

Facies K: Welded tuffs are thought to be present where dark, chloritic and flattened fragments are present in otherwise dense, homogeneous rock. Flattening is parallel to regional bedding, and the chloritic fragments are interpreted as fiamme. These rocks are similar to inferred Archean welded tuffs described and illustrated by Henderson (1975).

Facies L: Facies L is volumetrically the most important pyroclastic facies in the Timiskaming. These beds are extensively cross-stratified, parallel laminated, and scoured. Facies L forms stratigraphic units interbedded with the fluvial facies A, and have apparently formed by aqueous reworking of tephra, most likely by fluvial processes.

Paleocurrents

In the turbidites of McGarry Township (Fig. 37.2) paleocurrent flow towards the southeast is indicated by sole marks and oriented argillite clasts viewed on the base of a bed, and also to the northeast based on ripple marks. Near Kirkland Lake an easterly component of paleoflow is indicated by 2-dimensional views of cross-stratification. Other localized paleocurrent results based on sedimentary structures represent only a few measurements per outcrop, and are widely scattered throughout the study area.

Dimensional grain orientation and clast orientation in conglomerates have provided additional data. Clast orientation indicates flow towards the northwest in Skead Township. By combining all lines of evidence (cross-stratification, clast orientation, dimensional grain orientation, and facies distribution) paleoflow appears to be generally towards the northeast and southeast in the braided river facies near Kirkland Lake.

Sedimentary fabrics have rarely been measured in Archean rocks, in part because the primary fabrics have often been extensively modified by later tectonic processes. However this study indicates that where primary fabrics have been preserved, measurement of the fabric provides paleocurrent information that is otherwise unavailable. As an example of field clast orientation data, Figure 37.3 shows a stereoplot of the orientation of clasts in one of the fluvial conglomerate beds.

Provenance

Newly recognized clast types in Timiskaming conglomerates are ultramafic clasts with spinifex texture in McElroy Township, and one occurrence of a conglomerate clast that was noted near Kirkland Lake. Point counting of conglomerates at seven localities gave results that generally agree with Bass (1961) and Hewitt's (1963) conclusions that a predominantly volcanic terrane was being eroded, but it would appear that Hewitt may have underestimated the content of phaneritic igneous rocks. This is thought to be due to the fact that he numerically counted the clasts rather than point counting, an approach that results in the smaller volcanic clasts being overemphasized.

Boutcher, *et al.* (1966) studied the petrography of Timiskaming sandstones from the Late Timiskaming area, and Cooke (1966) did likewise for Timiskaming Tuffs in the Kirkland Lake-Larder Lake area. The present study reports the first petrographic study of the sandstones from the latter area. Most samples are from fluvial rocks, because the turbidites are much more altered, and yield less petrographic information.

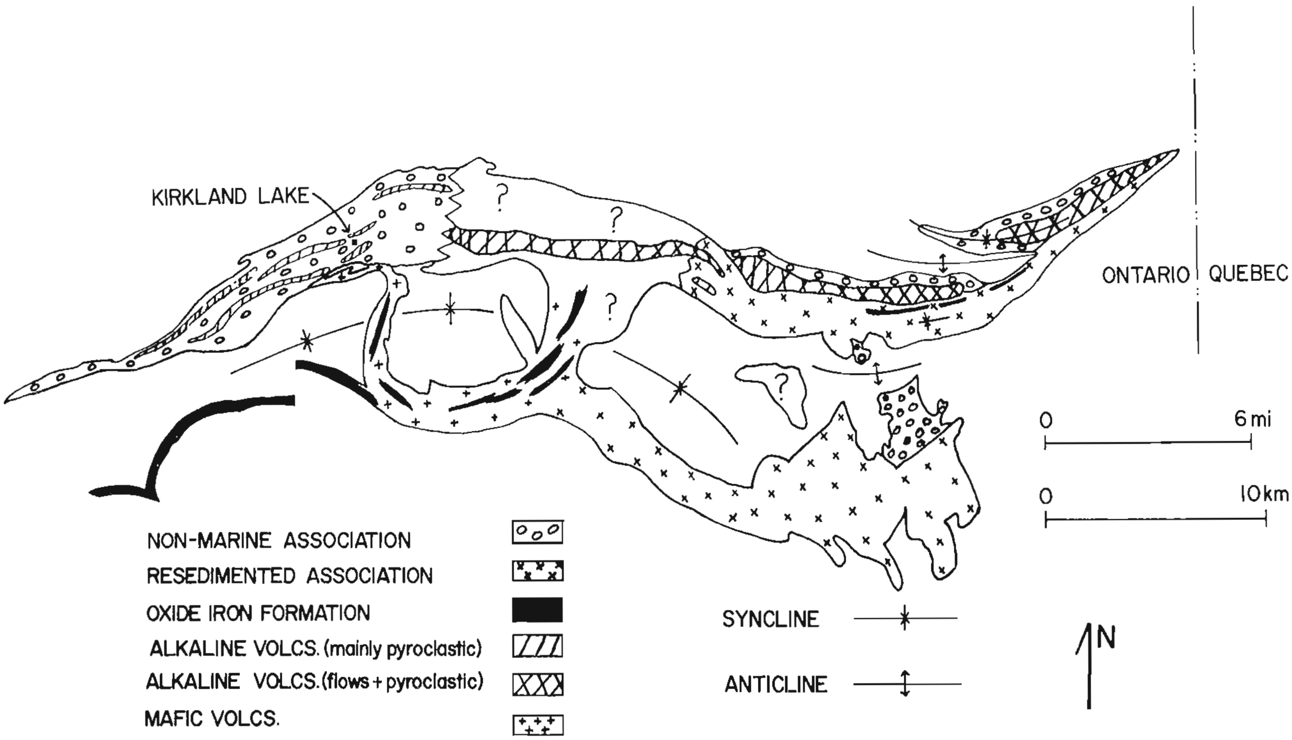

Figure 37.4. Distribution of Timiskaming sedimentary facies associations plus Timiskaming volcanic rocks. The majority of volcanic rocks (> 95%) near Kirkland Lake are pyroclastic. Question marks show areas of such poor exposure that depositional environments cannot be defined. Prominent volcanic unit below the two question marks serves as a marker "bed", and permits correlation of the Nonmarine Association with the Resedimented Association.

The most abundant constituent in all the samples was volcanic lithic fragments (average 56.8%). Quartz averaged 9.9 per cent and feldspar, overwhelmingly albitic plagioclase, averaged 12.8 per cent. Table 37.1 shows a typical modal analysis. Fluvial rocks at the base of the Timiskaming succession in McVittie and McGarry townships contain much more quartz (average 23.3%) than other fluvial Timiskaming rocks. Turbidites show little difference in quartz content from the Timiskaming fluvial facies.

Jolly (1974) has documented the widespread occurrence of prehnite and pumpellyite in volcanic rocks older than the Timiskaming. Possible examples of these minerals have been found only in detrital lithic fragments — none has been found in the sandstone matrix which is largely chloritic.

Facies Distribution and Paleogeography

Figure 37.4 shows the regional distribution of Timiskaming sedimentary facies associations. Pyroclastic facies are developed on too small a scale to be shown, but flows are virtually absent in the western third of the map-area. Question marks indicate areas of poor exposure.

Fluvial sediments do not appear to form continuous rock bodies along strike. Complex structural geology in the Larder Lake area makes stratigraphic relationships

Table 37.1

Modal analysis of Timiskaming lithic arenite

Component	Volume percentage
Quartz	9.75
Potash feldspar	0.75
Plagioclase	13.75
Mafic volc. fragments	28.75
Felsic volc. fragments	19.75
Phaneritic fragments	0.50
Sedimentary fragments	0.00
Chert	0.25
Other non-opaques	0.00
Opaques	2.50
Matrix	18.25
Cement	5.75
	100.00

perpendicular to strike uncertain, and the fluvial sediments may be repeated by folding. Along strike, however, in the northern belt the fluvial rock bodies in Teck and McVittie townships are separated by turbidites near the base of the Timiskaming in Gauthier Township. Braided river sediments near Kirkland Lake are time correlated with these turbidites by means of an alkaline volcanic unit, which intertongues with the two sedimentary facies, and forms a link between them. The braided river sediments near Kirkland Lake are laterally separated from the correlative turbidites by at most 9 km. Such a short distance between a nonmarine environment and a submarine fan implies a very narrow shelf, or no shelf at all — the fluvial rock body probably represents a remnant of a volcanic island, with a submarine fan developed offshore in deeper water. This inference is in part based upon the extreme narrowness of "shelves" around modern volcanic islands.

If the Timiskaming rocks in the southern belt are correlative with those in the northern belt, the fluvial strata must pinch out southwards as no fluvial sediments have been identified in the southern belt of sediments.

Tectonic Evolution

Most workers would probably agree that the underlying McVittie Basalts of Ridler (1970) are probably submarine, inasmuch as they are extensively pillowed. Braided river deposits at the base of the Timiskaming define points of uplift, and this is offered as independent evidence of an unconformity between the Timiskaming Group and McVittie Basalts deduced on structural grounds (Thomson, 1946). It is uncertain whether adjacent areas receiving turbidity current deposition were likewise uplifted, because there is no way of obtaining absolute or even relative water depths between the pre-Timiskaming volcanics and the turbidites. If these turbidite areas were tectonically stable or subsided, then the uplift that did occur was selective and did not occur on a basin-wide scale.

In the western third of the northern belt, fluvial deposition and emplacement of pyroclastic material continued until near the end of Timiskaming deposition, when cherts and exhalative deposits accumulated in a quiet water environment. In contrast, the eastern third of the northern belt experienced stronger subsidence such that following a period of fluvial deposition and trachytic volcanism, turbidity current deposition ensued. The intervening volcanism is suggested as a cause of subsidence, whereby the crust loses support for itself as the magma is drawn out during volcanic activity.

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Sample distribution and partial metamorphic zonation, 1976

Wayne T. Jolly[1]
Regional and Economic Geology Division

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 section study of the Abitibi greenstone belt and the associated intrusives is in its final stages. The aim of the project is to gather closely spaced specimens from traverses along all available access routes and to compile a metamorphic zonation of the belt. The purpose of this report is to summarize the distribution of samples and to outline a preliminary zonation of metamorphic minerals in part of the Abitibi Belt.

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 billion years (Krogh and Davis, 1971), of great extent and thickness are usually folded into narrow synclines between large granitic intrusions. The lavas are normally arranged in piles that become progressively more felsic upward (Goodwin, 1976; Jolly, 1975; Gelinas and Brooks, 1974). The oldest lavas are dominantly mafic and ultramafic flows of Mg-rich komatiitic type, but these are mixed with moderately Fe-enriched tholeiitic types (Arndt, 1974). A thick sequence of strongly Fe-enriched tholeiites that displays progressively increasing breccia fractions concentrated at flow tops (Wilson and Morrice, 1976) follows. The upper third of the typical stratigraphic section is composed largely of pyroclastic volcanic cones.

2) Sediments, mostly post-volcanic but also apparently pre- or synvolcanic (Dimroth, *et al.*, 1973), are normally exposed in synclinal troughs within the volcanic exposures. Most of the sediments are deep water greywackes. Iron-formation is 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 deposited under subaerial conditions. Jolly (1974) reported that numerous clasts from the conglomerates, derived by erosion of the adjacent volcanic pile, carry a prehnite-pumpellyite assemblage. These relations lead naturally to the conclusion that some of the sedimentary basins were formed above the lava pile after or during its initial episode of folding.

3) Intrusive rocks, broadly circular in plan, are abundant at all margins of and within the lava-sediment pile. These are divided into four broad types:

a) Gabbroic, dioritic, tonalitic, and other intermediate bodies of layered or massive character, are identical in composition to the enclosing lavas (Jolly, 1976a and b). Commonly, the bodies contain anorthositic phases.

b) Granitic plutons (Fig. 38.1) of large size are commonly present within the lavas and the surrounding terrain; many such plutons are closely associated with abundant pegmatitic dykes.

c) Granitic gneiss plutons (Fig. 38.1), commonly with igneous interiors, are present at many places on the margin of the volcanic belts and in places intrude the volcanics. The granitic gneissic margins have commonly experienced retrograde metamorphism and now exhibit assemblages characteristic of the greenschist facies (Ridler, 1972; Jolly, 1974).

d) Late Syenitic stocks and accompanying epizonal dykes and sills are widely distributed. Chemical data suggest that these plutons are co-magmatic with the alkaline volcanism of the primarily sedimentary Timiskaming Group of the Kirkland Lake area (Cooke and Moorhouse, 1969; Ridler, 1970).

Tectonic history. 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. The major synclinal features within the belt trend broadly east-west. Ages of the granitic materials average 2.4 billion years (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 Lake and Larder Lake, are not intruded by massive plutons, and behaved during Kenoran deformation as independent tectonic blocks of broadly synformal character.

The interrelations between the plutons and synclinoria may be readily used to subdivide the belt into tectonic blocks (Fig. 38.2, c.f. Kalliokoski, 1968). Each such block is delineated by structural breaks or major synclinal features, commonly containing sediment, which surround uplifted volcanic 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

[1]Brock University, St. Catherines, Ontario.

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Geol. Surv. Can., Paper 77-1A (1977)

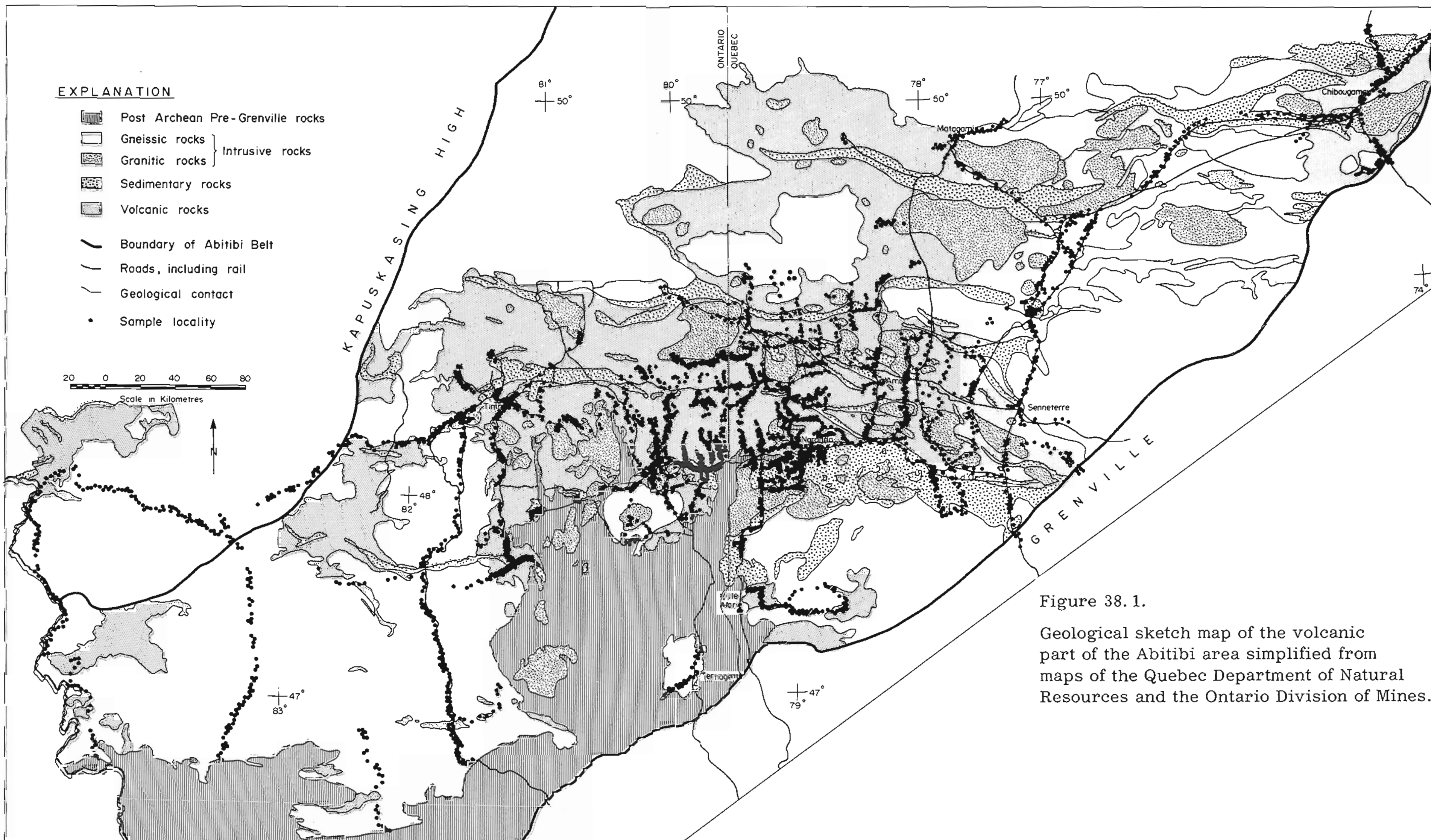

Figure 38. 1.

Geological sketch map of the volcanic part of the Abitibi area simplified from maps of the Quebec Department of Natural Resources and the Ontario Division of Mines.

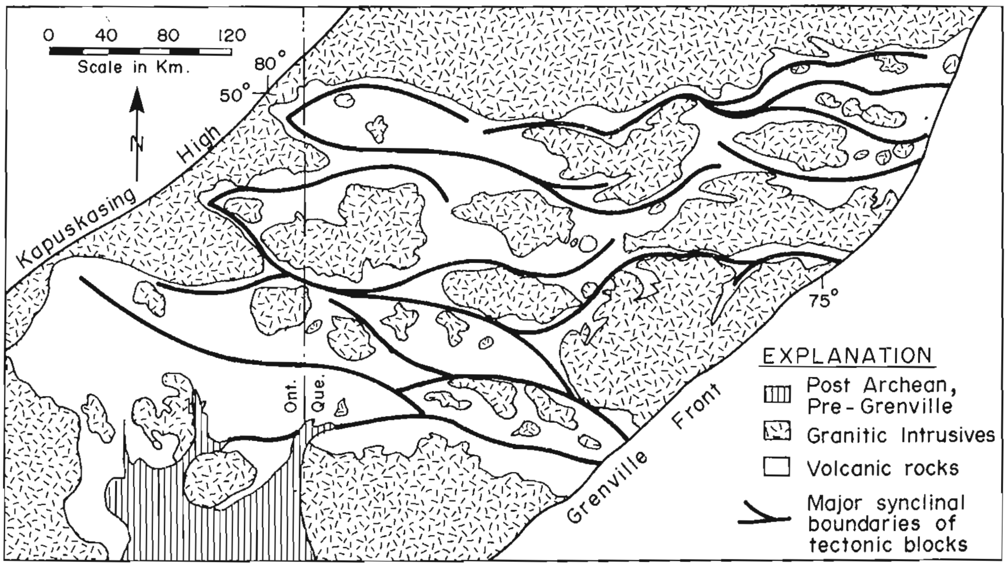

Figure 38.2.

Tentative tectonic subdivision of part of the Abitibi Belt, indicating approximate positions of major synclinal belts or structural breaks discussed in the text (c.f. Kalliokoski, 1968).

plutonic core. The tectonic blocks are viewed as centres in which the granitic cores rose to form anticlinal structures while the denser surrounding lavas sank to form deep, steep-sided troughs.

Metamorphic Zonation

The extrusive rocks exposed in the region between Chibougamau, Quebec and Timmins, Ontario have been divided, for purposes of metamorphic classification, into five zones and three metamorphic facies (Fig. 38.3) as follows:

Zone 1 – Diagenesis

Subgreenschist facies

Zone 2 – Prehnite-pumpellyite zone

Greenschist facies

Zone 3 – Albite-epidote-actinolite-chlorite zone

Amphibolite facies

Zone 4 – Hornblende zone

Zone 5 – Pyroxene zone

Each of the five subdivisions is based on microscopic secondary mineralogy. However, most specimens may be classified in the field from textural features.

Zone 1. Rocks exhibiting simple diagensis are not common in the Abitibi Belt. They are present only in sedimentary belts composed largely of greywacke or conglomerate. The best known example is the western part of the Timiskaming Group near Kirkland Lake, Ontario. The matrices, clasts and rock fragments of these sediments commonly carry white mica and clay minerals as the only secondary phases. A few volcanic clasts have been reported to contain detrital epidote, prehnite, and pumpellyite derived from nearby volcanic rocks (Jolly, 1974). These sediments, in part fluvial in origin, overlie unconformably all other Archean rocks in the area except a few plutonic intrusions of syenitic composition. The latter have developed metamorphic aureoles, so that certain parts of the sedimentary deposits contain actinolite or even hornblende.

Subgreenschist facies – Zone 2. Because no zeolites have been observed in any of the Archean terranes studied, the zeolite facies as defined by Coombs and others (1959) is not present in the Abitibi Belt. Rocks containing prehnite, prehnite plus pumpellyite, or pumpellyite plus epidote are widespread (Baragar, 1968; Goodwin, 1972; Jenson, 1972; Gelinas and Brooks, 1974; Jolly, 1974) within the volcanic rocks (Fig. 38.3). Such rocks are most common in the Lake Abitibi region near the Ontario-Quebec boundary, where the width of the greenstone belt is greatest. Near intrusive bodies, such prehnite-pumpellyite rocks give way to actinolite-epidote-bearing types which originated as a result of relatively low pressure contact metamorphism. Even in areas relatively close to lobate intrusions at the boundary of the greenstone belt, the metamorphic effects on the volcanics are commonly superficial and suggest little deep-seated activity:

1) The pumpellyite-prehnite mineralogy has been overprinted by actinolite-epidote-bearing assemblages in all contact aureoles, suggesting that the former was originally present throughout the lavas before intrusion occurred.

2) Higher grade terranes commonly contain irregular patches of over one kilometre in width in which pumpellyite is preserved, a common feature of hornfelses derived at low pressures in association with intrusion (*see* Turner, 1968).

3) Prehnite-pumpellyite-bearing rocks are neither foliated nor penetratively deformed, and therefore still display original textures; plagioclase laths, olivine euhedra, and matrix components may still be recognized because the secondary minerals did not grow across grain boundaries.

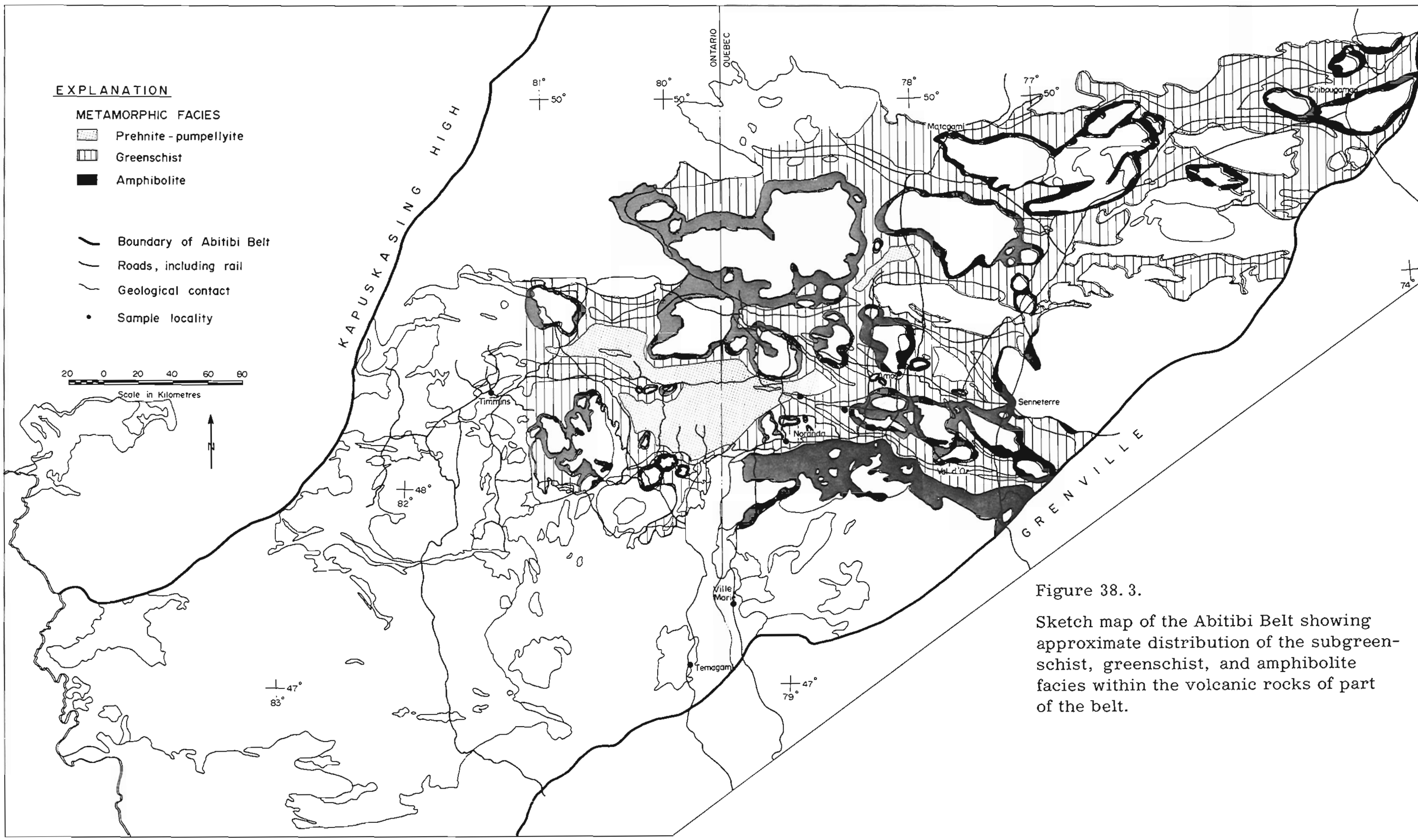

Figure 38. 3.

Sketch map of the Abitibi Belt showing approximate distribution of the subgreenschist, greenschist, and amphibolite facies within the volcanic rocks of part of the belt.

4) Most of the pumpellyite or prehnite-bearing rocks contain 4 to 6 per cent H_2O, suggesting that they were hydrated during alteration. In addition, rock hydration occurred predominantly in porous rocks (amygdaloidal zones, flow breccias, pyroclastic rocks, and along fractures, pillow margins, flow tops, or other channelways. Impermeable rocks (flow interiors) contain few secondary alteration products. Such relations suggest that both fluid and load pressures were low during metamorphism.

The above observations are interpreted to reflect a low pressure regional metamorphic event in the volcanic pile before it was intruded and deformed during the Kenoran Orogeny. The thickness of the lava pile was probably no greater than 10 kilometres, for even in more recent metamorphic terrains with low geothermal gradients (New Zealand, Coombs, *et al.*, 1959; Seki, 1969) prehnite pumpellyite is supplanted by lawsonite as the dominant Ca-Al silicate at depths greater than 10 000 m.

Greenschist facies – Zone 3. This group of rocks consists of those carrying albite, actinolite, epidote, chlorite, and stilpnomelane as the dominant secondary phases. Actinolite needles are normally visible in hand specimens, especially when associated with light coloured veins of calcite, prehnite, or quartz. Most of these rocks display a crude foliation, with actinolite growing in preferred orientations across primary grains; original textures are blurred and, with progressively higher grade, finally blotted out by complete recrystallization. All such rocks in the Abitibi Belt are broadly associated witb intrusive bodies. Thus, the metamorphic isograds are related to proximity to intrusions rather than to depth of burial. The common occurrence of actinolite-bearing rocks at margins of the volcanic belts suggests that the hornblende-bearing granitic gneisses disrupting the pile were cool at the time of intrusion. An east-northeast-trending belt of hornblende-bearing sedimentary gneiss, probably correlative with the Wabigoon Gneiss Belt of the western Ontario-Minnesota region, forms the northern boundary of the Abitibi greenstone belt. The volcanics near the contact contain a predominantly actinolite-bearing assemblage. It seems possible, therefore, that the gneiss belt represents an early pre-Kenoran and pre-volcanic orogenic event greater in age than 2. 75 billion years, the zircon age of the Abitibi volcanics.

Amphibolite facies. The rocks of the amphibolite facies are invariably schistose. For simplicity, hornblende has been distinguished in thin sections from actinolite on the basis of pleochroic formulae. All brownish amphiboles are considered aluminous and therefore to be hornblende. Electron microprobe studies in progress confirm the accuracy of such a generalization. Rocks of the amphibolite facies are found only in contact margins near intrusive bodies of various sizes. Many hornblende-bearing rocks are closely associated with pegmatites. Widths of hornblende zones are proportional to pluton size. In certain areas anthophyllite, cordierite, and garnet appear within hornblende zones. Many of the largest intrusions are surrounded by thin zones containing diopsidic pyroxene along with garnet, plagioclase, anthophyllite and less commonly cummingtonite if bulk rock compositions are appropriate.

Current Studies

Over 1000 localities in the granitic gneiss terranes intervening between the greenstone belts have been visited (Fig. 38.1) and sampled during the past year; these are currently being petrographically subdivided into metamorphic types. Included in the area studies are the granulites and migmatites of the Kapuskasing High, which trends northeast across the Superior Province. In addition, volcanic exposures not yet completed are being metamorphically zoned.

Acknowledgments

The author wishes to express appreciation to R.H. Ridler for discussion in the field. This project is supported by the Geological Survey of Canada and the National Research Council of Canada.

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REGIONAL METALLOGENY AND VOLCANIC STRATIGRAPHY OF THE SUPERIOR PROVINCE

Project 740097

R.H. Ridler
Regional and Economic Geology Division

Introduction

Two brief field excursions were undertaken in summer 1976. These were: (1) to the Kirkland Lake area with R.K. Wanless, to follow up and extrapolate by further field examination and sampling, previously established isotopic age data; and (2) to northwestern Ontario, guided by J.M. Franklin; to examine and contrast metallogenic features of greenstone belts of the western Superior Province with those of the central Superior Province. The generous assistance provided by geologists of both the Provincial Government and industry is gratefully acknowledged.

Jolly (1977) and Tihor and Crocket (1977) have discussed regional metamorphism of the Wawa-Abitibi, and the distribution of gold in exhalites and associated rocks of the Kirkland Lake — Larder Lakes area, respectively.

Age of the Otto Stock

In the attempt to define the regional stratigraphy of the Abitibi supergroup by isotopic means, attention has been focused on zircon-bearing units, specifically felsic plutonic and volcanic rocks. The Timiskaming Group is a major stratigraphically relatively young, volcano-sedimentary complex of the south margin of the Abitibi Belt. Its age is of great interest, but previous attempts to determine this age have failed because the intermediate, potassic composition of the volcanics of the group is incompatible with the occurrence of zircons, and the group is generally characterized by penetrative deformation and pervasive greenschist metamorphism.

Associated with the Timiskaming volcanics is a series of differentiated, mesozonal to epizonal alkaline plutons which includes the Otto Stock (Lovell, 1972). Recently published whole-rock isochron (Bell and Blenkinsop, 1976) and paleomagnetic (Pullaiah and Irving, 1975) data have been interpreted as suggesting approximate correlation of the Stock with the Nipissing Diabase at about 2200 m.y. Because the Stock may be co-magmatic with the Timiskaming volcanics and thus Archean in age, the massive, medium grained granitic core of the Stock and a known occurrence of diabase cutting the Stock's coarsely trachytoid annulus were sampled for isotopic age determination.

The dyke is a northerly-trending, plagioclase glomeroporphyritic diabase with a markedly chilled margin. It cuts and brecciates coarse grained syenite possessing an easterly trachytoid fabric. The dyke is therefore a member of the pre-Cobalt Matachewan swarm and by usual criteria, the Stock is Archean. Supporting isotopic data and a review of the pertinent geology and controversy, will be published later.

Northwestern Ontario

Areas visited included: (1) Red Lake (Campbell Red Lake and Dickenson mines); (2) Confederation Lake (South Bay mine); (3) Savant Lake; (4) Sturgeon Lake (Matabi and Sturgeon Lake mines); (5) Atikokan (Steep Rock mine); and (6) Finlayson Lake. The following impressions gained at Red Lake are based on the essentially cursory examination afforded by a field trip format.

The geology of the Red Lake belt has some striking similarities to that of the Abitibi Belt of the Timmins area. Within a series of veined and altered intermediate to mafic flows (including a variolitic member) reminiscent of the Vipond Series, the mineralization consists of, (a) auriferous discordant, carbonate-rich quartz veins occupying a prominent, dilatant, second(?) fracture cleavage (Campbell), and (b) auriferous, laminated, carbonate-sulphide-chert interflow exhalite (Dickenson) (cf. Fryer and Hutchinson, 1976). Stratigraphically, the mineralization lies close to, but below a locally enigmatic but exhalite-bearing felsic volcanic unit analagous to the Krist fragmental. Epizonal plutons associated with the sequence contain gold deposits of probable "porphyry" affinities, completing the picture.

An occurrence of carbonate-rich, chloritic schist reminiscent of Larder Lake-type exhalite occurs just below the oxide exhalite-bearing portion of the felsic unit mentioned above, as another facies of the unit. The outcrop examined, to which our attention was directed by Jim Pirie of the Ontario Division of Mines, lies on the shoreline of Red Lake immediately west of the main shaft of the Cochenour-Willans mine. In this outcrop, lenses of brown-weathering, beige carbonate with gash quartz veins, and bearing a few flecks of a green mica, are folded in various degrees of transposition into the westerly-trending, prominent second foliation. In spite of its "sheared" appearance, the rock has not been interpreted as the locus of a major fault, as have its cousins in the Abitibi Belt, perhaps because the pronounced foliation is about perpendicular to bedding, and therefore to the northerly trend of the zone.

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LAC DE MORHIBAN AND NATASHQUAN RIVER MAP-AREAS, QUEBEC

Project 760020

James H. Bourne, G. Stott[1], B. Borduas[2], and A. Lalonde[3]
Regional and Economic Geology Division

Introduction

The Lac de Morhiban (12 M/E½) and Natashquan River (12 N/W½) areas (approximately 6000 square miles) were mapped by helicopter-supported reconnaissance during the summer of 1976. The centre of the area is 170 km northeast of the town of Havre St. Pierre (population 3000). There are no permanent settlements or habitations within the map-area.

There are few outcrops in the area and the bush extremely thick; however there is considerable local variation.

All rocks encountered are of Precambrian age. A table of the units mapped, in the order of their inferred age, is presented in Table 40.1. Approximately 30 per cent of the area is underlain by low-grade metamorphosed sedimentary (Wakeham Group) and volcanic rocks. Unmetamorphosed intrusive rocks (units 8 and 9) postdate these sedimentary and volcanic rocks and underlie approximately 20 per cent of the region. The remaining 50 per cent is composed of gneisses, foliated plutonites, and an anorthosite complex; all of higher metamorphic grade than the above.

Unit 1 — Gneiss

This unit is restricted to the northern third of the map-area, and is best exposed around Lac de Morhiban in the northwest corner of the map-sheet.

This unit comprises a varied assemblage of pinkish gneisses of granitic composition to whitish grey gneisses of tonalitic to dioritic composition. Compositional layering (gneissosity) is well developed throughout. The presence of sillimanite at Lac de Morhiban suggests that at least some of this unit was derived from sedimentary rocks. Of all map-units encountered, this is the only one that has been subjected to partial melting.

Detailed study will be required before a meaningful subdivision of this unit can be worked out. The poor exposure prevalent in the northern portions of the map-area will make this difficult.

Unit 2 — Foliated quartz monzonite to granite

This unit consists of varieties of pink, foliated, medium grained homogeneous rocks which are interpreted as metamorphosed plutonic rocks. They are cross-cut by pegmatites and mafic dykes. Compositionally they range from adamellite to monzonite. The grain size of these rocks is 2 mm; however it was originally much larger. Primary phenocrysts of plagiocalse have been rounded and necked off at the ends and fractured internally to produce an augen-like texture which is composed of fine grained monomineralic aggregates instead of large single crystals. This feature is particularly common from the northwest corner of the the map-area southeast to the vicinity of Lac la Galissonière. In the southeast corner of the map-area, this unit consists of medium grained foliated granitoids and corresponds to the "foliated and gneissic augen granite" of Sharma (1973).

Unit 3 — Anorthosite

About 70 km^2 of the 1170 km^2 Romaine River anorthosite complex (Sharma and Franconi, 1975) outcrops in the western margin of the map-area. The Romaine River mass itself is only a lobe of a still larger body, which is one of the largest anorthosite massifs known.

The rock is composed of an interlocking aggregate of plagioclase grains and is white. It is almost devoid of mafic minerals and has been completely recrystallized. It lacks the large purplish to greenish coloured grains of plagioclase typical of the Morin and other anorthosite masses.

Table 40.1

Table of units encountered in the Morhiban-Natashquan map-areas. The units are listed in the order of inferred geological age, with unit 1 being the oldest

MAP UNIT	LITHOLOGY
10	diabase dykes
9	hornblende adamellite
8	biotite adamellite
7	gabbro
6v	intermediate volcanic rocks
6h	hypabyssal quartz monzonite
5	sandstone/quartzite
4a	foliated adamellite
4d	foliated diorite
3	anorthosite
2	foliated quartz monzonite to granite
1	gneiss complex

[1]University of Waterloo, Waterloo, Ontario.

[2]Ecole Polytechnique, Montreal, Quebec.

[3]University of Ottawa, Ottawa, Ontario.

From: Report of Activities, Part A;
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A relict igneous layering, defined by 4 mm plagioclase crystals in mafic-poor layers alternating with 2 mm plagioclase grains in more biotite-rich layers (between 3 and 15 mm thick), was found at one locality. Trains of fine grained magnetite, present in the 4 mm plagioclase-rich layers, locally emphasize the layering.

The contact of the anorthosite with the surrounding foliated diorite correlates well with the 56 500 gamma contour on the aeromagnetic map.

Unit 4d — Foliated diorite

The diorite is restricted to the perimeter of the anorthosite mass. It is a fine to medium grained, rusty brown coloured rock. It is a very deeply weathered, strongly foliated, homogeneous unit, with the foliation being defined by trains of mafic minerals.

The foliated diorite is clearly younger than the anorthosite since blocks of anorthosite several feet across are found within the diorite near the contact between the two rock types. The composition of the diorite changes gradually, becoming more silica-rich with increasing distance from the anorthosite mass. The boundary with the enveloping foliated adamellite unit has arbitrarily been placed where "augen" of potash feldspar make their appearance. The unit is therefore considered as the metamorphosed equivalent of a non-porphyritic igneous rock.

Unit 4a — Foliated adamellite

The foliated adamellite is the outermost of the three rock types which together comprise the Romaine River anorthosite complex. In this map-area, it is volumetrically the most important of the three units, although on a larger scale it appears as a "rind" wrapping about the anorthosite mass, (Sharma and Franconi, 1975).

It closely resembles some of the foliated quartz monzonite to granites of unit 2. It is a medium grained, foliated, augen granitoid rock which is pink on both the fresh and weathered surfaces. Distinctive features of this unit are:

- presence of augen of potash feldspar (not mineral aggregates as in unit 2).
- distinctive northwest-trending aeromagnetic trend. (Unit 2 rocks are not associated with any recognizeable aeromagnetic pattern.)
- spatial association of this rock type with the other rocks of the anorthosite complex.

As has been mentioned, this foliated adamellite unit grades into the diorite unit, so the two are undoubtedly genetically related. The precise nature of the relationship between these two units and the anorthosite mass itself is not clear, but they are known to be younger (probably only slightly younger) than the anorthosite.

A large northeast-trending lineament seen on the LANDSAT imagery (presumably a fault) corresponds to the mapped contact between this unit and the meta-sandstones of unit 5.

Unit 5 — Sandstone

The metasedimentary rocks in this region are part of the Wakeham Group (McPhee, 1959). They are chiefly feldspathic quartzites consisting of more than 90 per cent quartz. Large scale cross-bedding is a common feature throughout the area. These are often sufficiently undeformed to enable top determinations to be made. Minor components of these rocks are shale and a distinctive muscovite-chlorite schist (meta-arkose).

The sandstone occurs as two "tongues" which narrow northward and are separated from one another by volcanic rock. The rocks of the western tongue are white, well cemented with a conchoidal fracture, and in most places lack visible bedding. The rocks of the eastern tongue are light to medium grey. In hand specimen the original outlines of the component sand grains are easily seen. Compositional layering (bedding) is clearly visible. It appears that the sandstone to the east has been less intensely deformed and may be of lower metamorphic grade than that to the west.

Unit 6v — Intermediate volcanic rocks

Volcanic flows and pyroclastics, predominantly of intermediate composition, form a northward extension of the volcanic rocks originally described by Sharma and Jacoby (1973). They are porphyritic, the phenocrysts consisting of salmon pink potash feldspar and blue quartz eyes although many rocks have potash feldspar phenocrysts only. The composition ranges from quartz latite porphyry to latite porphyry (Travis, 1955).

Mafic volcanic rocks considered to be a part of this unit are exposed along the shore of Ruffin Lake (12 M/east half) and pink rhyolitic volcanics near the southern boundary of the map-area, but at least 95 per cent of the volcanic rocks are of intermediate composition.

Unit 6h — Hypabyssal quartz monzonite

In several places the volcanic rocks of unit 6v may be traced into coarser grained rocks of quartz monzonite composition. These quartz monzonites are finer grained and more saussuritized than are similar rocks of unit 7 and 8. A good example of the relationship between the volcanics of unit 6v and the hypabyssal quartz monzonite is exposed on the shore of Lac Barrin — a small lake 6 miles southwest of Lac la Galissonnière. The lake trends northeast and is approximately 1 mile wide. Along the southeast shore and striking parallel to it, is exposed a suite of very fine grained nonporphyritic volcanic rocks. On islands in the centre of the lake the volcanic rocks are porphyritic and resemble typical members of unit 6v. Further northwest the groundmass of these volcanics becomes relatively more coarse grained. Finally, in the northwest corner of Lac Barrin, there is a sudden increase in grain size and the rock becomes a hypabyssal quartz monzonite. There are no intrusive relationships visible between this hypabyssal intrusive unit and volcanic rocks, rather the relationship appears to be gradational. The large grain size difference between this unit and the volcanic rocks enables the definition of the contact between the two to be made with considerable confidence. A similar situation apparently exists in the southeastern corner of the map-area where a suite of volcanic rocks grades southwards into a hypabyssal intrusive equivalent. In the case of the Lac Barrin region, it is suggested that the exposures define a cross-section through a volcanic pile, with the coarse grained rocks of the northwest corner of Lac Barrin corresponding to the deeper section of the pile, and the aphanitic volcanic rocks of the southeast corner of the lake corresponding to the top of the pile. Preliminary results suggest the presence of ferro-hastingsite as a mafic mineral, which may indicate that these rocks have alkaline affinities.

Unit 7 — Gabbro

Dykes and sills of gabbroic composition are widespread throughout the area underlain by the sandstone (unit 5) although only one of them (the Lillian Lake Gabbro), is sufficiently large to be shown on Figure 40.1. All the gabbroic rocks have been deformed and metamorphosed to the same metamorphic grade as the sandstone and volcanic rocks, and although these rocks are clearly younger than the sandstone and volcanics, there is no evidence of a great time span between these units. In the southwest, the gabbro is a northward extension of a gabbroic suite of rocks mapped and petrographically described by Claveau (1949) and Grenier (1957).

The largest mass of gabbro encountered during the mapping was called the Lillian Lake Gabbro and is exposed in the southeastern corner of the map-sheet. The outcrop pattern suggests that it is a large sill which has been tilted and deformed during a period of folding with a northeast trend. It is not associated with an aeromagnetic anomaly, nor is it associated with sandstone — a fact which casts some doubt on the correlation of this mass with the large number of aforementioned gabbro dykes and sills.

Unit 8 — Biotite adamellite

The previously described units are cut by a number of discrete plutons of adamellite to quartz monzonite composition. They are coarse grained, pink porphyritic ($\sim$ 25 x 15 mm) rocks which contain blue quartz phenocrysts and biotite as the major mafic mineral. They are massive throughout, with the exception of the margin, which shows a faint to strong foliation defined by streaky mafic minerals. Several of the larger intrusions have been assigned letters on the sketch map for purposes of discussion. All are associated with distinctive highs on the aeromagnetic map. The aeromagnetic expression appears to be related to a retrograde reaction which has resulted in the breakdown of the primary mafic mineral(s) into aggregates of biotite, sphene and magnetite. Since these rocks do not show any other signs of metamorphism, it is assumed that this breakdown is of autometamorphic origin.

The plutons were divided into two groups (group 1 consists of plutons a, c, e, f, and group 2 of plutons b and d). The distinctive features of each group are:

1) The aeromagnetic anomaly pattern associated with group 1 plutons is circular, as are the plutons themselves. The aeromagnetic pattern associated with group 2 plutons has a northwest trend. This indicates that although the two groups of rocks appear almost identical in the field, group 2 plutons are older and have been deformed mildly by a northwest-trending fold system which has not affected the group 1 plutons.

2) The foliated margins of plutons b and d may be up to 1 mile wide, while the foliated margins of group 1 plutons are much narrower, on the order of several hundred feet.

A point of note is the geographic distribution of these plutons. The younger group 1 plutons (a, c, e, and f) define a crude line which trends northeast and bisects the area. The group 2 plutons tend to lie northwest of this line. This belt continues to the northeast into the area mapped by Stevenson (1967), as indicated by the aeromagnetic map pattern.

Unit 9 — Hornblende adamellite

One large pluton of hornblende-bearing adamellite was found during the mapping. It is strikingly similar to the plutons of unit 8 and was probably emplaced at about the same time as these rocks. The major differences from unit 8 plutons are:

1) There is no aeromagnetic anomaly associated with the pluton.

2) The mafic mineral is hornblende.

3) The phenocrysts of potash feldspar are not as common as they are in unit 8 rocks, nor are they are large (10 x 15 mm).

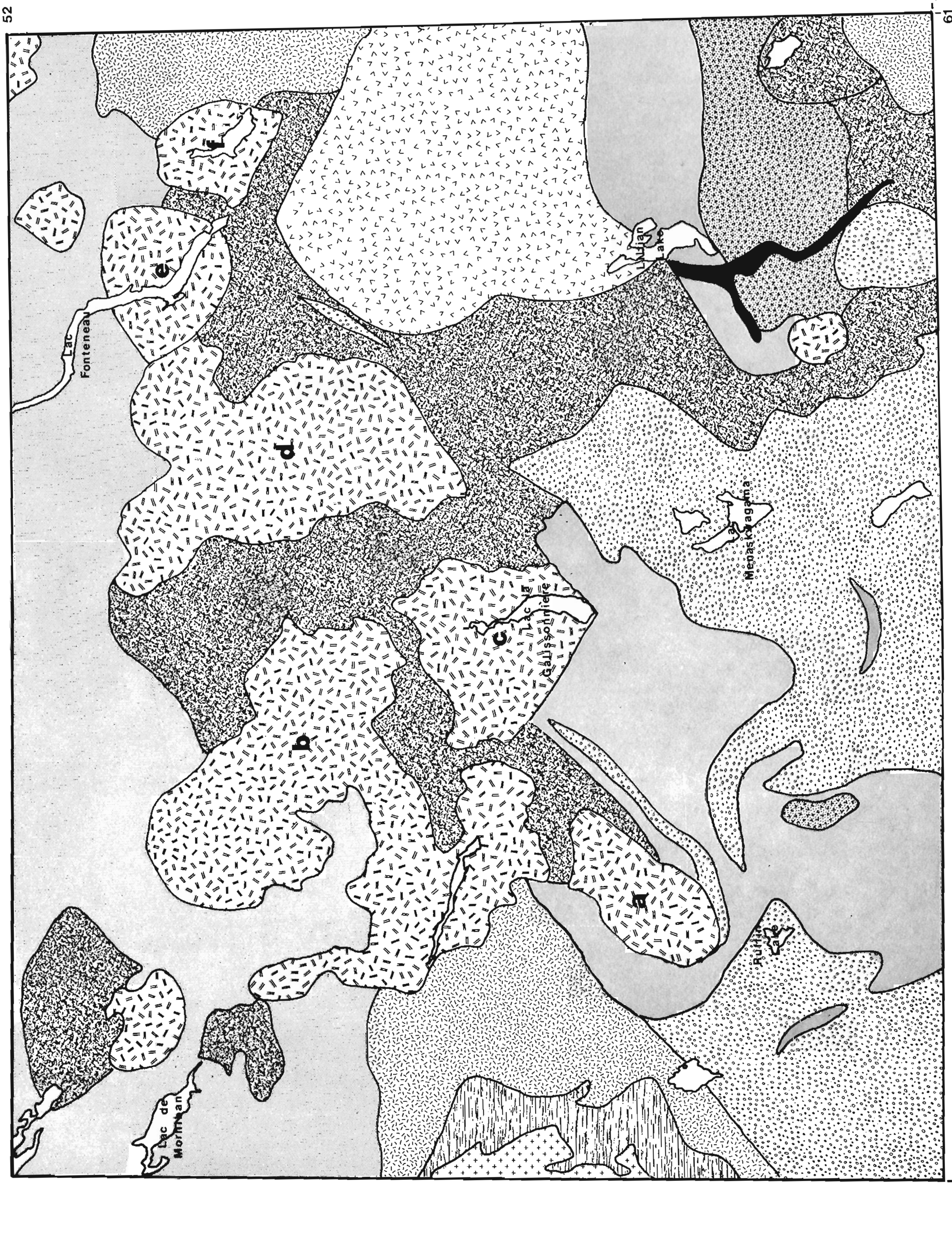
The area underlain by this mass is heavily drift-covered so the boundary indicated on the map is only approximate.

Unit 10 — Diabase dykes

A swarm of diabase dykes trending approximately 85° is clearly visible on aerial photographs. They cross-cut the hornblende adamellite (unit 9) and are undeformed, making them the youngest rocks in the area. Volumetrically they are insignificant.

Structural Geology

Evidence exists for at least three periods of deformation, although it is doubtful that any one unit of rocks was affected by all three phases. In the northeast, the gneiss unit (unit 1) contains refolded folds in several places, which constitutes good evidence for two periods of deformation in these rocks.



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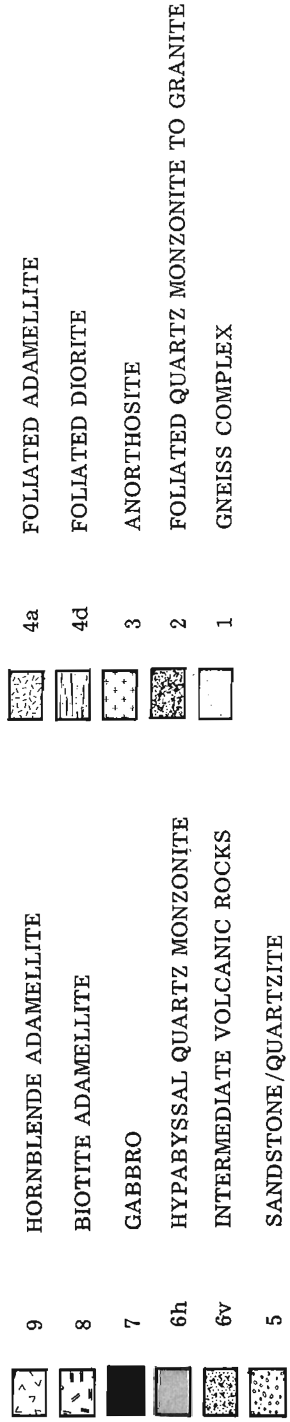

Figure 40.1. Sketch map showing the distribution of lithologic units within the area mapped.

The second period of folding (F_2) resulted in the formation of a series of tight folds with a southeasterly trend. All units from unit 1 up to and including the group 2 plutons of unit 8, were affected by this second period of folding. This southeast trend is the strongest trend found in the region.

The F_2 folds have been refolded about northeast-trending axes (F_3) with a southwest plunge, evidence for which is clearly displayed in the sandstone unit in the south-central part of the area.

The time gap between the second and third periods of deformation was probably not great. It has been mentioned that the group 2 plutons of unit 8 have been affected by the southeast deformational event, and that these plutons appear similar to the group 1 plutons as well as to the hornblende adamellite, both of which clearly postdate the last period of folding since folds with the northeast trend have been deformed by the emplacement of the plutons.

Metamorphism

The metamorphic grade of the gneiss unit is difficult to determine because of its granitic to tonalitic composition. In one outcrop a quartz-biotite-garnet-sillimanite assemblage was observed, suggesting that middle amphibolite facies conditions or higher were attained. The presence of a phase ascribed here to partial melting supports this conclusion. The metamorphic grade of units 2 to 4 appears to be as high as that of unit 1; however, no evidence of partial melting was observed. The partial melting observed in unit 1 was associated with an earlier, higher grade metamorphic episode which was retrograded by a later metamorphic event which affected unit 1 to 7 inclusive.

The metamorphic grade of the metasedimentary and metavolcanic rocks is much lower. In these rocks biotite-epidote-ankerite-quartz assemblages are found, and in the metamorphosed gabbros intruding the sandstone, epidote-actinolite-quartz assemblages are common. Phenocrysts in the metavolcanics have compostions less than An_{15}. These data suggest a middle to upper greenschist facies classification for units 5 to 7. It has already been mentioned that indirect evidence exists of a decrease in metamorphic grade from west to east.

Rocks of units 8 to 10 are unmetamorphosed.

Economic Geology

No showings of any importance were discovered during the mapping. The only showing which has been reported (Sharma and Jacoby, 1973, p. 14) mentions chalcopyrite mineralization near the contact of gabbro and sandstone. Pyrite is common in the gabbro sills throughout the area.

In recent years, considerable interest has been expressed in the uranium potential of granitic bodies located to the south (Cooper, 1957, Ministère Richesses Naturelles du Québec, 1974). Descriptions in the literature suggest that some may well be related to the rocks of unit 8 described here, particularly the Ferland Lake granite (Cooper, 1957) and the southern granite stock of Claveau (1949).

Summary and Interpretation

The gneiss unit is clearly the oldest unit within the area mapped. The rocks comprising this unit are so varied and complex that no satisfactory origin for all of these rocks can be offered. It is possible that rocks of more than one origin have been included in unit 1. The primary age(s) of the unit is unknown. The gneiss has been subjected to two periods of folding. The first (F_1) postdated the climax of the first discernible period of metamorphism (M_1) since irregular veinlets of assumed partial melt origin have been deformed by F_1. The temperatures associated with M_1 must have been high to have resulted in partial melting of a granitoid of intermediate composition. Evidence of any earlier deformational episode(s) was not found.

The foliated quartz monzonite to granite (unit 2) cuts across the rocks of unit 1 and contains evidence of only one period of deformation. This unit appears to be older than the three associated rock types of the Romaine River anorthosite complex, although no contact relationship between the foliated adamellite and unit 2 rocks was observed in the field. If the primary age of the adamellites in Labrador of 1350 ± 100 m.y. (Krogh and Davis, 1973) and the primary age of the foliated adamellites mapped here are similar, as is reasonable, then the age of the unit 2 rocks would be of the order of 1400-1500 m.y., while unit 1 rocks would be Aphebian or older.

Contact relations between the foliated adamellite of the anorthosite complex and the metasandstone and metavolcanic rocks are obscured by drift; however, it is likely a fault contact since the boundary between the two rock types coincides with a major lineament seen on the LANDSAT imagery. The strike direction of the fault is parallel to the general northeast trend defined by the group 1 plutons of unit 8. This, and the large amount of intermediate to acid volcanics of possible alkaline affinities, suggests that the metasandstones and metavolcanic rocks may have formed in association with the development of a rift structure.

Emslie (pers. comm., 1976) has suggested a relationship between the time and mechanism of formation of anorthosite rocks and the deposition of sedimentary rocks (which have been intruded by gabbro sills) adjacent or near to them. He considers them all to be related to the development of a central North American rift system. He cites the Harp Lake anorthosite/Seal Lake sedimentary and volcanic rocks as an analogous example, among others. If this is the case, then the time gap separating the formation of the anorthosite and the deposition of the sandstone and volcanic rocks cannot have been great.

The grade of metamorphism of the sandstone and volcanic rocks is considered middle to upper greenshist, while the grade of the foliated adamellite and other rocks of the anorthosite complex, as well as of the foliated quartz monzonite to granite (unit 2) is amphibolite facies. It is assumed that this difference is related to different intensities of metamorphism in the areas underlain by the two different rock suites during a second period of metamorphism (M_2). The M_2 metamorphism resulted in the downgrading of the middle to upper amphibolite facies rocks of unit 1 to the same metamorphic grade as the rocks of units 2 to 4 inclusive.

Associated with this second period of metamorphism is the development of two sets of folds. The earlier of these (F_2) is a series of tight northwest-trending fold structures. This northwest trend is the dominant structural trend within the map-area. The F_2 folds are folded by a series of northeast-trending F_3 folds which are of local occurrence. The time gap between the development of the F_2 and F_3 folds was small.

The final event recorded in the area is the emplacement of a series of diabase dykes trending approximately 85°. Volumetrically they are insignificant.

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GEOLOGICAL STUDIES: THIRTY MILE LAKE AND MACQUOID LAKE MAP-AREAS, DISTRICT OF KEEWATIN

Project 750005

A.N. LeCheminant, D.H. Blake[1], R.W. Leatherbarrow[2], and L. deBie[3]
Regional and Economic Geology Division

The 1: 250 000 scale mapping of the west half of MacQuoid Lake (55 M) and the east half of Thirty Mile Lake (65 P) map-areas was completed during the 1976 field season. D.H. Blake spent one and one-half months in the map-area as part of a regional study of the nature and environment of Dubawnt volcanism. Detailed mapping of all known copper and uranium occurrences was undertaken by A. Miller of the Uranium Section.

South of the unconformity underlying the Dubawnt Group the region consists of east to northeast trending belts of Archean(?) metasediments and metavolcanics separated by broad bands of felsic gneisses (Fig. 41.1). The gneisses are intruded by a suite of alkalic rocks ranging in composition from biotite pyroxenite to biotite pyroxene syenite. Epizonal granite plutons intrude both the alkalic rocks and the gneisses.

Kaminak Group(?)

Migmatized Kaminak Group(?) metasediments and metavolcanics underlie a 2 km wide east-northeast trending belt in map-sheet 65 P/1 (Fig. 41.1). This extension of a major belt of Kaminak Group(?) rocks previously mapped to the east (LeCheminant *et al.*, 1975; Reinhardt and Chandler, 1973) is characterized by thin-layered amphibolite and hornblende-plagioclase gneiss associated with garnet and biotite-rich paragneiss and schist. Mineral assemblages in metapelites commonly contain sillimanite and cordierite. Rare oxide facies iron-formation locally contains fayalite. The gneissosity trends northeasterly and has moderate to steep northwest dips. Some deformation postdates metamorphism and migmatization as evidenced by numerous deformed and boudinaged white pegmatitic granitic layers within the unit.

East to east-southeast trending belts of dominantly metavolcanic rocks form prominent ridges between Bissett and Parker lakes and south of Kazan Falls. Fine grained amphibolite is interlayered with thin laminated mafic schist and hornblende-plagioclase gneiss. These rocks are probably metamorphosed mafic to intermediate lava flows, tuffs and volcanigenic sediments. The typical mineral assemblage is hornblende-plagioclase-epidote-chlorite-sphene; garnet is present in some layers. Small meta-gabbro bodies within the belt may be dykes or sills intruding the mafic volcanic pile. Iron-formation and quartzofeldspathic paragneiss are present in minor amounts. Gneissosity generally trends east-west and dips steeply north.

Kaminak Group(?) amphibolite, garnet-biotite paragneiss, and garnet-muscovite schist underlie Dubawnt Group rocks in a small inlier 3 km west of Bissett Lake and near the eastern boundary of the area studied. In both areas the metapelite assemblage (quartz-muscovite-biotite-garnet-plagioclase) reflects lower amphibolite grade metamorphism.

Aphebian or Archean Gneisses

The area south of the Dubawnt unconformity can be broadly divided into three domains of Aphebian or Archean gneisses. Southern and northern domains of well foliated felsic gneisses are separated by a northeast trending zone of irregularly layered migmatitic gneiss intruded by alkalic and granitic plutons.

The southern domain consists of well layered grey to white granodioritic gneisses intimately infolded with amphibolite and paragneiss of the Kaminak Group(?). The gneisses have been affected by at least three phases of penetrative deformation. The dominant structural trend is northeasterly, and reflects F_2 tight to isoclinal folds which have shallow plunging hinge lines. The association of thin amphibolite and paragneiss bands with well layered granodiorite gneiss suggests part of the layered gneiss unit may be derived from intermediate volcanics or greywackes. Thick units of homogeneous composition are thought to be orthogneisses.

Migmatitic gneisses in the central domain are dominantly granitic in composition and have an irregular foliation which is broadly conformable with the contacts of younger alkalic and granitic plutons. The gneisses contain mafic-rich streaks, metamorphosed diabase and lamprophyre segments, and have been intruded by numerous sheets of granite, aplite and pegmatite.

The northern domain is characterized by a variety of felsic ortho- and paragneisses in which cataclastic textures are predominant. East of the Kazan River gneissic quartz monzonite and granodiorite, locally containing large potash feldspar augen, are infolded with belts of Kaminak Group(?) metavolcanics. To the west felsic augen gneisses contain numerous lenses, layers and boudins of hornblende-rich gneiss and amphibolite. Near the west end of Thirty Mile Lake two areas of migmatitic granitoid gneiss are separated by a northwest trending belt of thinly interbanded leucocratic and mafic gneisses. This 'black and white' gneiss unit contains isolated garnet and hypersthene porphyroblasts, partly altered to chlorite or biotite, and may represent retrograded granulite facies gneiss.

[1]Bureau of Mineral Resources, Geology and Geophysics, Canberra, Australia.

[2]Carleton University, Ottawa, Ontario.

[3]University of New Brunswick, Fredericton, New Brunswick.

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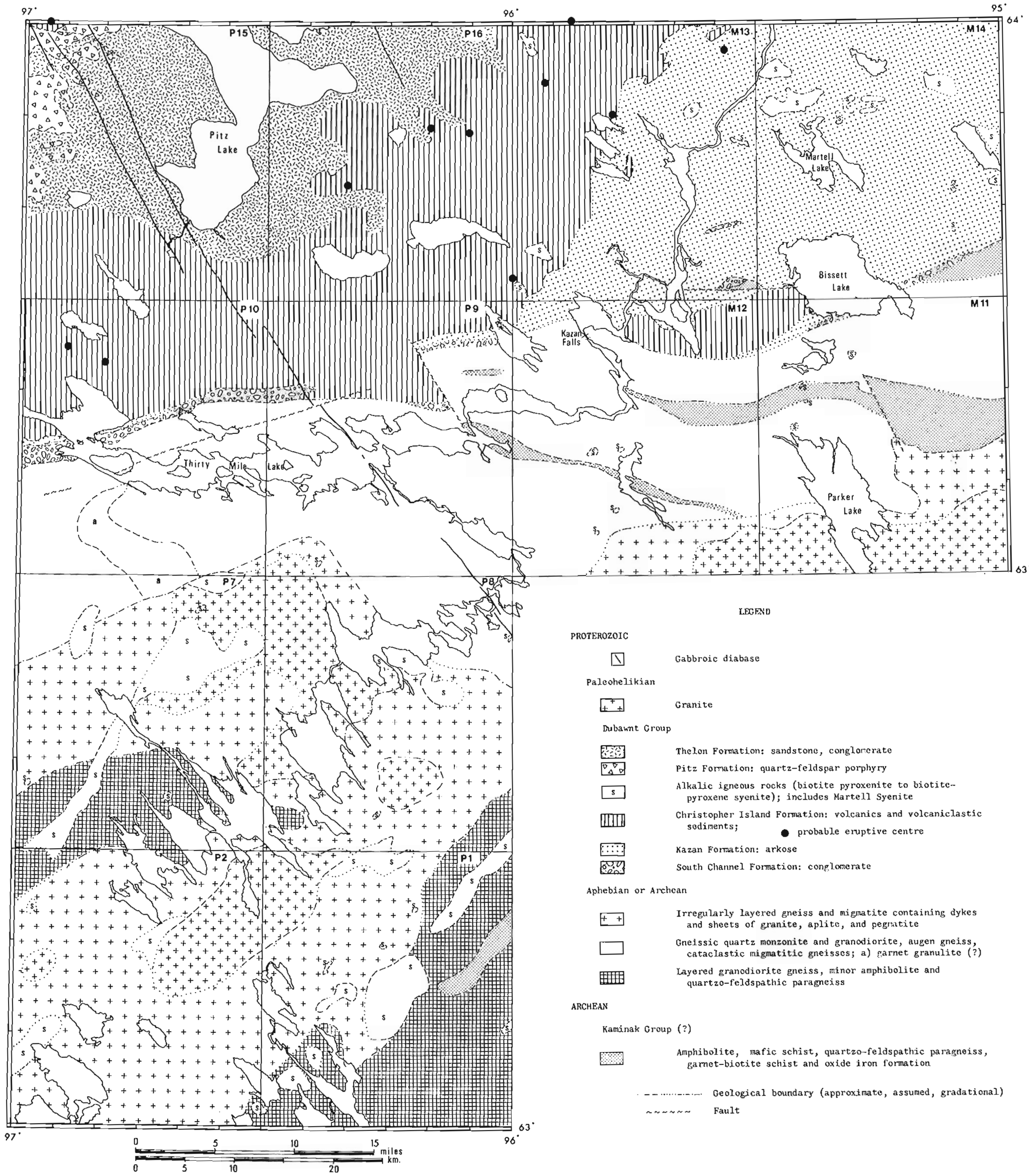

Figure 41.1. Geological sketch map of Thirty Mile Lake (65 P east half) and the northwest corner of MacQuoid Lake (55 M) map-areas.

An east-northeast trending fault crosses the west end of Thirty Mile Lake. The fault is marked by a prominent linear, local mylonite zones, and intense chlorite, epidote, quartz, and carbonate veining and alteration. Mylonite, protomylonite, augen gneiss, and retrograded garnet granulite outcrop in the narrow zone between the base of the Dubawnt Group and the fault. These gneisses have a characteristic aeromagnetic expression that suggests the fault continues to the west of the study area.

South of the fault numerous east to east-northeast trending diabase dykes and south to southeast trending lamprophyre and feldspar porphyry dykes cut the northern domain gneisses. North of the fault the predominant diabase dyke trend abruptly changes to northeast.

Dubawnt Group

The northern part of the area consists of poorly exposed unmetamorphosed continental sediments and volcanics of the Dubawnt Group (Fig. 41.1). The group comprises the South Channel, Kazan, Christopher Island, Pitz and Thelon formations and the Martell Syenite (Donaldson, 1965), and is isotopically dated by Rb/Sr methods at about 1725 m.y. (Wanless and Loveridge, 1972).

The basal unit of the group, the South Channel Formation, unconformably overlies felsic gneisses in the west and Kaminak Group(?) metasediments and metavolcanics in the east. The coarse polymictic conglomerate is formed almost entirely of locally derived basement fragments and ranges in thickness from 0 to about 1800 m. Lenses of similar conglomerate occur locally within the conformably overlying Kazan Formation, which is predominantly a red, cross-bedded, fine to medium grained arkose containing a high proportion of sodic feldspar. Calcite, dolomite and hematite are common cements. The thickness of the Kazan Formation cannot be determined accurately, because of inadequate exposures, but is probably well over 1000 m in the east.

Volcanics of the Christopher Island Formation succeed the Kazan Formation. The contact between these two formations is conformable in the east, where an interfingering relationship is evident, but may be disconformable or unconformable in the west. The Christopher Island Formation, at least 300 m thick in places, is made up of predominantly subaerial lava flows, agglomerate and tuff, volcanic and tuffaceous sediments, and various associated dykes. These rocks apparently retain their original attitudes over much of the area. The lavas are mainly dark purple, maroon, and brown mafic trachytes containing biotite and augite phenocrysts, but some flows of extremely fine grained feldspar porphyry, locally with sanidine phenocrysts, and aphanitic dacite and flow-banded rhyolite are also present. Propylitic alteration, characterized by calcite, chlorite, quartz, and minor epidote, is widespread throughout the formation. The sites of local thick accumulations of agglomerate and/or dacitic to rhyolitic extrusives are shown as probable eruptive centres in Figure 41.1.

Lamprophyre dykes similar in mineralogy and composition to lavas of the Christopher Island Formation intrude the Kazan and South Channel formations and northern and southern domain gneisses. Dykes range in width from several centimetres to 20 m and trend south to southeast. Rounded gneiss xenoliths are locally abundant within the dykes, both north and south of the Dubawnt unconformity.

The Pitz Formation is exposed only in the northwest corner of map-sheet 65 P/15 and has a maximum thickness of about 100 m. The formation comprises two acid lava flows, containing abundant large feldspar phenocrysts and subordinate rounded quartz 'eyes'. The topography west of Pitz Lake reflects in part the exhumed surface on the top of the Pitz Formation flows. In the upper flow, but not in the more restricted lower flow, the feldspar phenocrysts are pseudomorphed by kaolinite. The margins of the upper flow, including its top, are autobrecciated and intimately mixed with water-laid, mainly tuffaceous sediments interpreted as hyaloclastites. In the north the upper flow rests conformably on similar hyaloclastites at the top of the Christopher Island Formation.

The Thelon Formation, confined to the northwest corner of the area, has a minimum thickness of 100 m. It consists of flat-lying, cross-bedded orthoquartzite and conglomerate, locally with volcanic clasts, which overlie and abut against volcanics of the Christopher Island and Pitz formations, filling depressions between tongues of lava.

The Martell Syenite (Donaldson, 1965) refers to a group of irregular laccoliths, sills and dykes which intrude the Kazan and Christopher Island formations in the east half of the area studied. The intrusive rocks are pinkish to maroon, fine to medium grained, and range in composition from monzonite to syenite. They commonly contain augite, biotite, and feldspar phenocrysts. Thermal metamorphic effects adjacent to sharp intrusive contacts extend up to 30 m into Kazan Formation arkoses.

South of the Dubawnt unconformity numerous small stocks and dykes of alkalic igneous rocks intrude felsic gneisses and Kaminak Group(?) rocks. In map-sheets 55 M/11 and 55 M/12 these rocks are texturally, mineralogically, and chemically similar to the Martell Syenite but to the southwest they are generally coarser grained and range in composition from biotite pyroxenite to syenite. Many of the bodies appear to be multiple intrusions and several are marginally deformed. The gneissosity in adjacent gneisses is generally parallel to intrusive contacts but sharp truncations occur. The intrusives are cross-cut by pink granitic veins and offset by northwest trending minor faults.

Granite

Plutons of Paleohelikian(?) biotite granite underlie an area of sparse outcrop, felsenmeer, and numerous lakes in the southwest corner of the study area. The granite is mainly fine to medium grained, light pink, massive and leucocratic. Purple fluorite occurs locally. Sheets and dykes of similar granite intrude gneisses adjacent to the plutons. The pluton in the south half of map-sheet 65 P/8 contains numerous gneissic xenoliths and rafts and is locally weakly foliated. Shallow dipping foliations in the flanking gneisses and rare subhorizontal fabric within the granite suggests the present level of exposure may be close to the roof of the pluton. The granites postdate the east trending diabases and the southeast trending lamprophyre and feldspar porphyry dykes.

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GEOLOGY OF THE TULEMALU LAKE MAP-AREA, DISTRICT OF KEEWATIN

Project 740004

K.E. Eade and D.H. Blake[1]
Regional and Economic Geology Division

During the 1976 field season, mapping was completed in the western half of the Tulemalu (65 J) map-area. D.H. Blake spent one month in this half of the area, concentrating on the Dubawnt Group rocks. Preliminary results of the mapping in the east half were reported previously (Eade, 1976). Regional reconnaissance mapping (Wright, 1967) included the map-area and broadly outlined the distribution of some of the rocks. The simplified geological sketch-map (Fig. 42.1) delineates the major rock units in the west half of the map-area.

The Archean metavolcanics in the vicinity of Angikuni Lake consist of flows and pyroclastic rocks ranging from mafic to felsic composition. Compared to metavolcanics of similar age occurring elsewhere in southern District of Keewatin, the unit here contains a higher proportion of pyroclastic rocks (lapilli tuffs, crystal tuffs and agglomerate) and these are mainly of intermediate to felsic composition. The lavas, which unlike the pyroclastics are chiefly of mafic composition, are for the most part massive, but pillows are present locally. Sills of medium-grained metagabbro occur with the basic lavas. Dykes and plutons of a different metagabbro cut the metavolcanics in some places. Amphibolite present in a small area northeast of Angikuni Lake, is probably derived from basic extrusive rocks.

Metagreywacke is present in limited amount with the metavolcanic rocks east of Angikuni Lake, and a band of mixed metagreywacke and tuff occurs within migmatized paragneiss at the south boundary of the map-area. Some black slate and quartzite to arkosic quartzite, including bands of quartz pebble conglomerate, are present on the west side of Angikuni Lake. Disseminated pyrite characterizes some of the quartzite. The black slate appears to underlie the quartzite for the most part, but some intercalation of the rocks exists. The slate and quartzite overlie metavolcanic rocks, possibly disconformably, although in some places quartzite is interfingered with felsic tuff. A narrow band of chert-siderite iron-formation occurs with the slate and quartzite.

Migmatized paragneiss in the southern part of the area contains prominent white pegmatitic leucosomes. The presence of garnet, cordierite, and sillimanite, indicates a middle amphibolite grade of metamorphism. The paragneiss is probably derived from greywacke and tuff similar to that associated with the metavolcanics. In the east half of the map-area, pink quartz monzonite veins and masses intrude the migmatized paragneiss in many places. In the west half of the map-area, veins and small plutons of grey to white granodiorite to quartz monzonite cut the migmatized paragneiss.

The granodiorite gneiss, mixed gneiss and granodiorite map-unit (Fig. 42.1) groups together a variety of igneous and metamorphic rocks, mainly of granodiorite composition. The gneissosity ranges from pronounced layering to slight mineral foliation. Although the granodiorite is in part massive, a faint foliation is commonly evident. Dark bands and inclusions are abundant in the mixed gneiss but are rare in the granodiorite gneiss.

The Archean quartz monzonite is a medium grained, pink, homogeneous rock with inclusions present only close to contacts. It is cleaved and a vague secondary foliation is locally present.

Porphyritic quartz monzonite in the north part of the area is considered to be of late Aphebian age. It seems to be undeformed, unlike the Archean quartz monzonite.

Hornblende syenite, a pink, medium grained, massive rock, occurs along the east side of Tulemalu Lake. It intrudes Archean quartz monzonite and also forms small dykes cutting granodiorite gneiss southwest of Tulemalu Lake. The contact of the syenite with porphyritic quartz monzonite in the northeast is not exposed, and the assigned Helikian age for the syenite is uncertain.

The porphyritic granite in the northwest corner of the map-area, west of Tulemalu Lake, is a medium to coarse grained massive rock petrographically similar to the Nueltin Granite in the region to the south (Eade, 1973). It contains both hornblende and biotite and also scattered grains of purple fluorite. Its contact with the Dubawnt Group volcanic rocks is not exposed, but the volcanics are believed to overlie the granite unconformably.

The main outcrop areas of the Dubawnt Group volcanic and sedimentary rocks are shown in Figure 42.1; in addition there are several small outliers around Angikuni Lake. The group is dated isotopically at about 1725 m.y. (Wanless and Loveridge, 1972) and has not been affected by regional metamorphism. The Dubawnt rocks were laid down on an irregular land surface and in places have a regolith preserved beneath them. Initial dips probably ranged up to 35 degrees and over much of the area these are retained. Steeper dips can be attributed to the effects of penecontemporaneous faulting.

Sedimentary rocks are exposed locally at the base of the Dubawnt Group (Fig. 42.1). Some outcrops consist of polymictic conglomerate formed of locally derived basement fragments, in places accompanied by clasts of Dubawnt volcanics, but elsewhere, the sediments are represented by coarse grained to pebbly sandstone overlain by thinly interbedded fine sandstone, siltstone and mudstone. However, most of the Dubawnt Group in the area consists of volcanic rocks; these

[1]Bureau of Mineral Resources, Geology and Geophysics, Canberra, Australia.

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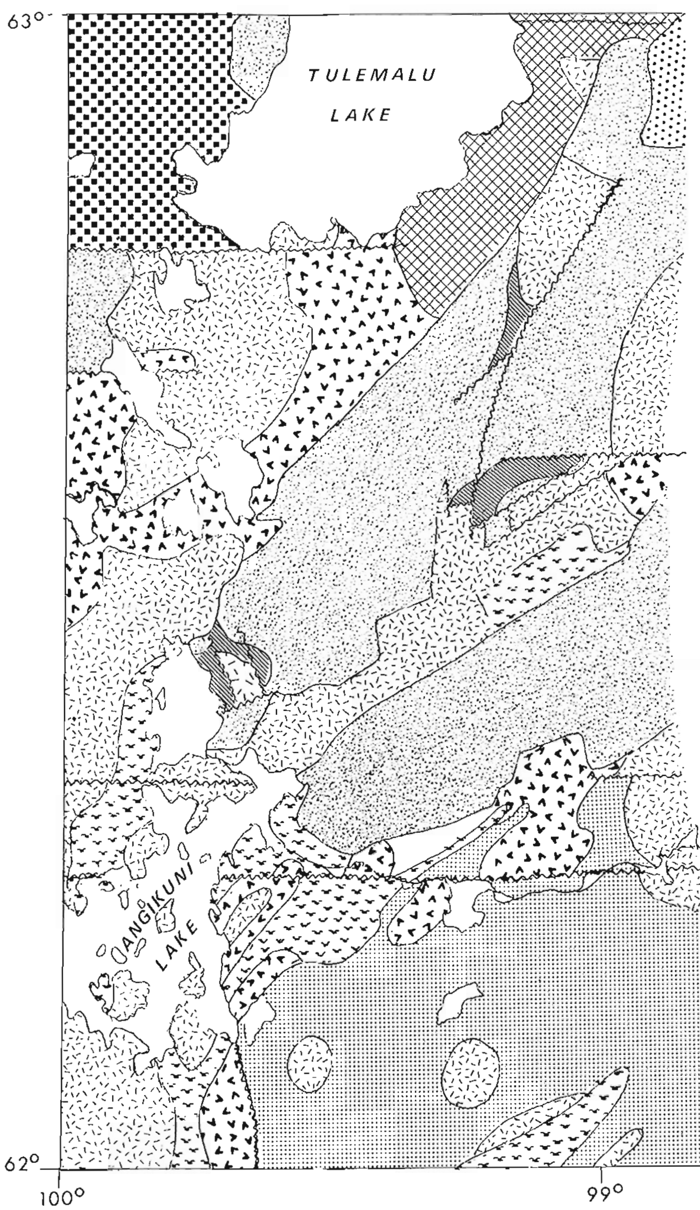
overlie crystalline basement unconformably and the polymictic conglomerate conformably. Following Donaldson (1965), the conglomerate is equivalent to the South Channel Formation, the feldspathic sandstone and overlying sediments may be correlated with the Kazan Formation, and the volcanics belong to the Christopher Island Formation.

The Dubawnt volcanics comprise subaerial lavas and pyroclastics, bedded hematitic volcaniclastic and tuffaceous sediments, vent breccia, and various associated high level minor intrusions. Southeast of Tulemalu Lake the volcanics attain a thickness of at least 1000 m. The sites of eruptive centres are indicated by exposures of vent breccia, agglutinate lava, and thick accumulations of massive pyroclastics. The lavas are mainly potassium-rich trachytes containing biotite and augite as phenocrysts and alkali feldspar as the dominant groundmass phase; some also contain phenocrysts of alkali feldspar, sodic plagioclase and pseudomorphs after olivine. Although not regionally metamorphosed, the volcanics show the effects of propylitic alteration, and secondary carbonate, chlorite and quartz are widespread.

Gabbro dykes, not shown on the accompanying sketch-map, abound in the area. In the southern part of the area slightly metamorphosed, east trending dykes, informally named the Tulemalu dykes, are prominent in the migmatized paragneiss unit. The dykes, commonly from 25 to 35 m wide, cross-cut the regional structure. Older northeast trending metagabbro dykes are common in the vicinity of Angikuni Lake. They are markedly more metamorphosed than the east trending dykes and show a foliation parallel to that of the adjacent country rocks. Fresh gabbro dykes of the northwest trending Mackenzie swarm occur in a few places.

Lamprophyre dykes associated with the Dubawnt Group volcanic rocks cut both the Dubawnt Group and the basement rocks. The dykes, discontinuous and narrow (average 2 m wide), are characterized by abundant biotite phenocrysts up to 5 cm in diameter. Some of these dykes contain rounded xenoliths of country rock.

In the Archean rocks south and east of Angikuni Lake two periods of deformation, similar to those to the east (Eade, 1976), are recognized. The earlier deformation resulted in a northeast trending foliation and in folds typically plunging gently to the northeast. The later deformation resulted in broad open flexures with poorly developed axial plane cleavage striking approximately 150 degrees. A different structural domain apparently exists in the Archean rocks northwest and



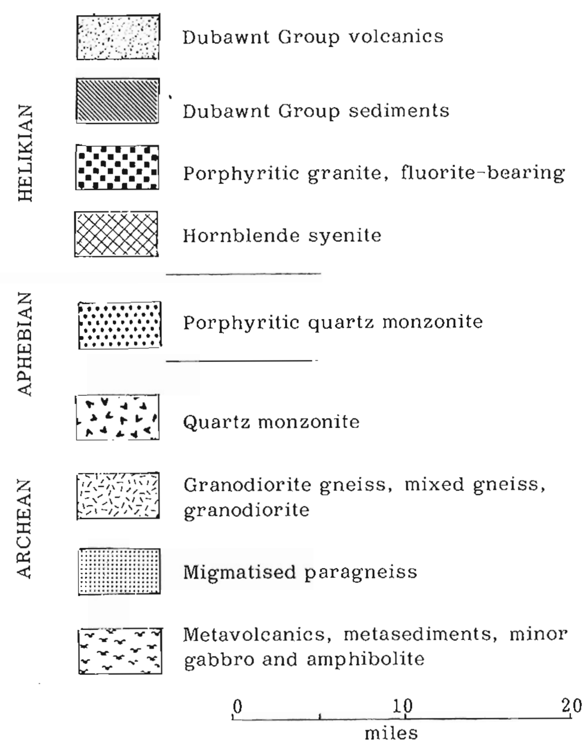


Figure 42.1. Simplified geological sketch map of the west half of the Tulemalu Lake map-area.

north of Angikuni Lake. On the northwest side of the lake, the foliation and folding trend north-northeast, but north from the lake foliation trends gradually change to north and north-northwest.

Deformation of the Dubawnt Group rocks is limited to minor warps and tilting associated with faults.

Major east trending faults mapped to the east (Eade, 1976) extend westward into the area and there are also some northeast and northwest trending faults that cut Dubawnt Group and basement rocks. The northeast faults are probably old faults which have been reactivated from time to time, whereas the northwest faults are considered to be younger.

Several companies carried out exploration for uranium in the map-area during the past field season. Five prospecting permit areas were held, as well as a number of claim groups. Interest was centred on the Dubawnt Group sedimentary rocks adjacent to their contact with the basement rocks, particularly where they are fractured along shear zones. During the present survey, scintillometer readings four to five times background were noted in these rocks. As well, readings three to four times background were found in some of the Dubawnt Group volcanic rocks and in fluorite-bearing porphyritic granite in the northwest corner of the map-area.

Mineralization was noted at several places in the Archean rocks. On a small island or reef in the south central part of the lake immediately north of Angikuni Lake, white quartzite is cut by quartz and quartz-carbonate veins that contain chalcopyrite and galena and there is a similar occurrence on another small island in the southwest corner of this lake. On the southern part of the west shore of this same lake, Archean lapilli tuff is cut by a complex network of quartz and quartz-carbonate veins, some of which contain galena. West of Angikuni Lake in white medium to coarse grained quartzite, a quartz vein stockwork is partially stained by malachite and contains some chalcopyrite and galena.

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GEOLOGY OF THE BARROW RIVER MAP-AREA, MELVILLE PENINSULA, DISTRICT OF FRANKLIN

Project 760026

Andrew V. Okulitch, Terry Gordon, J.R. Henderson[1], J.E. Reesor, and I.E. Hutcheon[2]
Regional and Economic Geology Division

This project extends previous work in the Foxe Fold Belt in southwestern parts of Melville Peninsula (Reesor *et al.*, 1975), eastward to Foxe Basin. Two and one half months were spent mapping eleven 1: 50 000 scale map-areas east of 83°30'W and between 67°00 and 67°30'N (south half of 46 O and P). The information and maps in this report are primarily the product of compilation and preliminary interpretation in the field and may be subject to considerable revision when analysis of data is completed. Much additional data is on 1: 50 000 scale maps, with marginal notes and structure sections, of areas 46 O/1, 2, 3, 6, 7, 8 and 46 P/4, 5, 6, 11 which have been placed on open file.

Figure 43.1 illustrates the lithologic distribution of major units in the area. No detailed differentiation of rock types within the Archean basement complex was attempted in the field although numerous varieties were recognized. Lenses of paragneiss, amphibolite and orthoquartzite observed near the coast may be related to the Archean Prince Albert Group but the affinity of

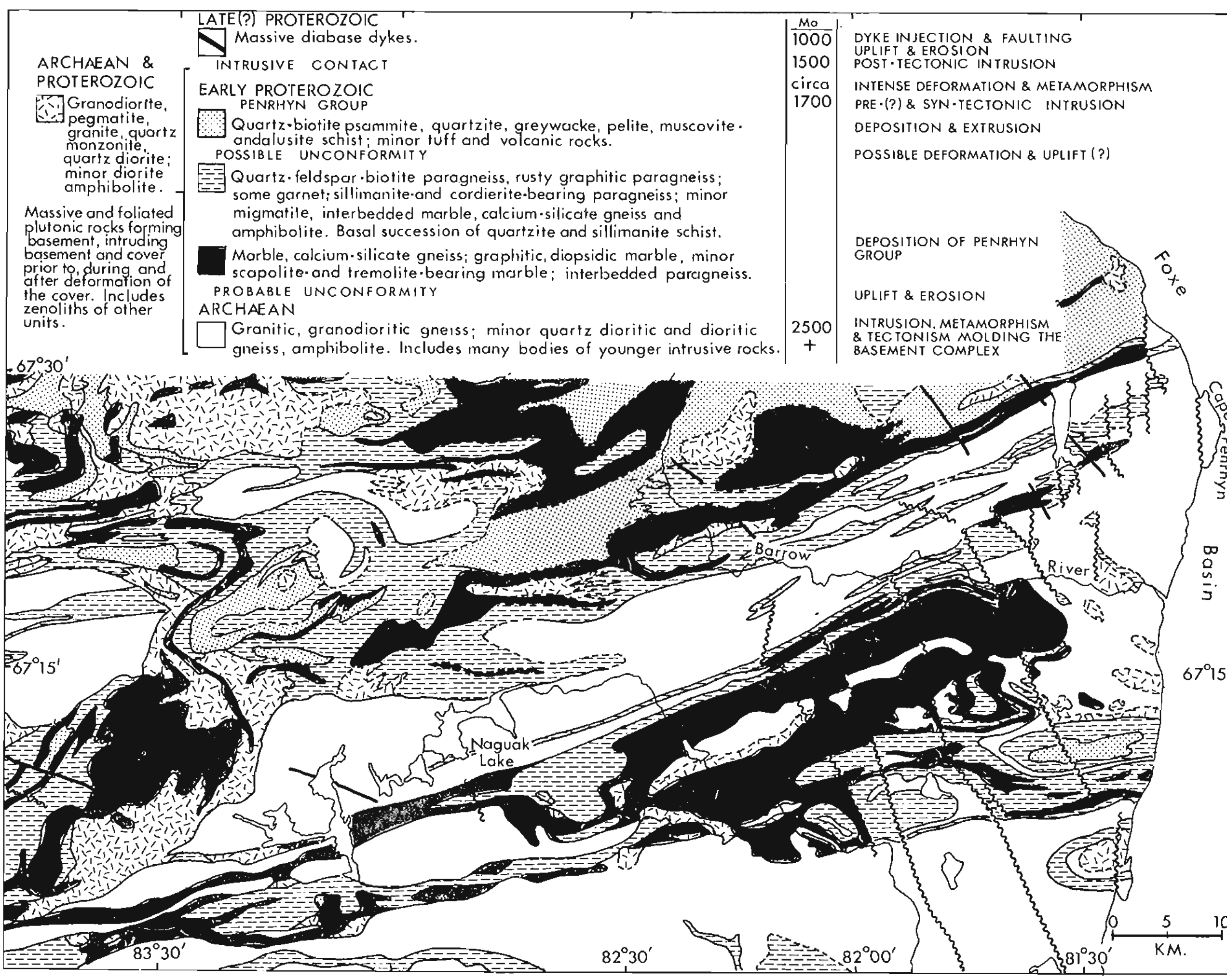

Figure 43.1. Distribution of major lithologic units, southern half, Barrow River map-area.

[1]Department of Geology, Grand Valley State College, Allendale, Michigan, 49401.

[2]Department of Geology, Carleton University, Ottawa, Ontario, K1S 5V6.

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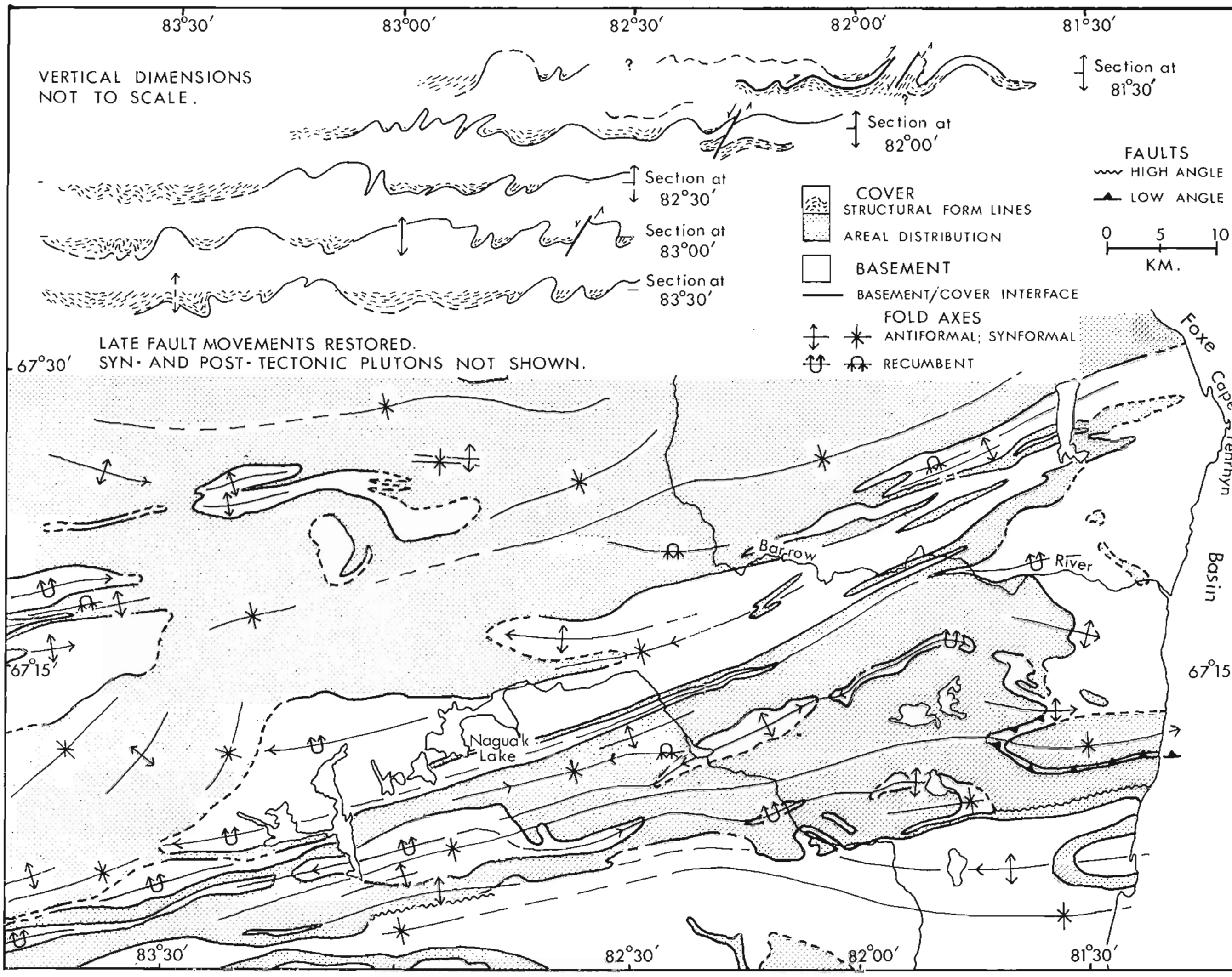

Figure 43. 2. Major structural elements, southern half, Barrow River map-area.

these units is uncertain. Unconformable relationship between the basement and the Penrhyn Group is based on the obvious lithologic contrast and the common presence of a thin unit of orthoquartzite with rare feldspathic grit beds lying upon a variety of gneissic rock types in the complex. The intensity of later tectonic events has obliterated any pre-existing angular discordance.

Units of the thin (50-100 m) basal sequence of the Penrhyn Group are not everywhere present together but form a distinctive assemblage of orthoquartzite, rusty sillimanite schist, grey-green micaceous rock suspected to be a metaregolith and minor amphibolite, marble and feldspathic grit. The basal sequence is overlain in succession by a predominantly calcareous unit (marble, calcium-silicate gneiss and interbedded quartz-biotite-feldspar paragneiss) ; a thick unit of a variety of paragneissic rocks with a thin bed of schistose paragneiss at its base; a unit of marble, calcium silicate gneiss and biotite quartzite; and at the highest observed structural and stratigraphic levels, by a unit of quartz-biotite-muscovite psammite and greywacke. A possible unconformity separating the last unit from the rest of the Penrhyn Group is suggested on the basis of contrasts in metamorphic grade and intensity of deformation observed in the two parts of the Group. Such contrasts can also be explained by rapid transitions or faulting rather than any discontinuity.

Only this generalized concept of the stratigraphy of the Group has emerged at this time. In detail, most units appear to be discontinuous and lensoid. Such relationships can be ascribed to facies changes of unusual frequency, numerous unconformities, or early, largely unrecognizable structures such as attenuated folds or thurst faults with trends transverse to later structures, or to all three features.

Metamorphism of the Penrhyn Group produced two distinct lithologic suites. Greenschist facies rocks have been observed in the upper unit of the Penrhyn Group in northern and easternmost parts of the map-area. Elsewhere the rocks are in uppermost amphibolite facies.

Pelitic assemblages in the lower grade rocks include chlorite-muscovite-quartz. Porphyroblasts of a mineral tentatively identified as andalusite are common in meta-psammitic units and tremolite occurs sporadically in calcareous rocks.

Assemblages in high grade rocks include garnet-biotite-sillimanite in paragneiss. Cordierite-sillimanite-garnet rocks occur mainly in the southernmost part of the map-area. Marble contains diopside-forsterite-calcite as well as a humite group mineral and scapolite.

The nature of the metamorphic boundary and its bearing on the possible unconformity within the Penrhyn Group will be the subjects of future petrographic work. This work will also attempt to place constraints upon tectonic models of the fold belt.

Preliminary modelling of major structural elements within the map-area is illustrated in Figure 43.2. Six phases of deformation can be observed in the various units. D_1, the nature of which is not known, embraces those events which molded the basement complex prior to deposition of the Penrhyn Group. D_2 formed attenuated isoclinal folds and the ubiquitous foliation in the Penrhyn Group, deformed the basement complex and obliterated most effects of D_1. Meagre evidence suggests that the trend of D_2 structures may have been northerly. In all but a few outcrops the associated foliation (S_2) is parallel to bedding (S_0). As mentioned previously, structures formed in D_2 may be responsible for some of the observed discontinuity of units.

D_3 and D_4 produced prominent mesoscopic and megascopic folds and imposed an east-northeast structural grain on the Foxe Fold Belt. D_3 folds are nearly isoclinal and recumbent and cannot with confidence be distinguished from D_2 structures in all instances. D_4 folds are more open and upright or overturned. Reversals of plunge of major D_3 and D_4 folds by later transverse warps (D_5) limit the exposed depth of the structural cross-section. Only in easternmost parts of the map-area is there any evidence of nappe-like gneissic sheets overlying the Penrhyn Group. The predominant direction of tectonic transport during D_3 and D_4 appears to be southward. The two fold phases are nearly coaxial with mainly horizontal axes and both deform S_0 and S_2. Little development of new mesoscopic structures took place and analysis of these phases by conventional geometric and statistical methods is thereby hindered. The lack of mineral recrystallization that could serve to define new foliation or lineation is all the more puzzling as metamorphism appears to have acted during and after deformation. Solutions to this problem are not yet evident and await petrofabric examination.

D_5 resulted in formation of broad transverse flexures trending north to northeast which produce variation in plunge of earlier structures. This effect can be seen in the area northeast of the river draining Naguak Lake. Earlier structures are deformed in western parts of the study area, thereby preserving higher structural and stratigraphic levels and burying rocks of the basement complex.

Steeply dipping fractures and faults formed in D_6 and are spatially related to emplacement of diabase dykes in eastern parts of the area. Most observed fault displacements appear to be left-lateral and east-side-up.

The timing of events in the map-area is summarized in Figure 43.1. No precise limits can be set on the age of the Penrhyn Group and considerable latitude exists for postulating the age of its units. Numerous undetectable unconformities or nappe structures may be present and would compound stratigraphic problems described above. Future analysis of zircons from a unit believed to be an acid volcanic rock found in the upper part of the Penrhyn Group may help to define its age.

No mineral occurrences of economic importance were discovered. Scattered rusty zones and gossans were found to contain mostly pyrite, graphite and minor pyrrhotite. A locality with high background radioactivity (Geological Survey of Canada, 1976) proved to consist of a lense of rusty pegmatite within the basal sequence of the Penrhyn Group, presumably containing small amounts of a disseminated uranium mineral (uraninite?).

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44. ARCHEAN SUPRACRUSTAL-BASEMENT ROCK RELATIONSHIPS IN THE KESKARRAH BAY MAP-AREA, SLAVE STRUCTURAL PROVINCE, DISTRICT OF MACKENZIE

Project 760024

John B. Henderson and R. Michael Easton[1]
Regional and Economic Geology Division

Introduction

Granitic rocks unconformably below the Archean volcanic and sedimentary rocks of the Yellowknife Supergroup are preserved and particularly well exposed at Point Lake, District of Mackenzie. A 1: 50 000 scale mapping project in the Keskarrah Bay map-area (parts of 86 H/2, 3, 6, 7) was undertaken in order to better appreciate the extent and significance of this granitic basement, the interrelationships of the various supracrustal facies at an apparent Archean basin margin, and ultimately to achieve a better understanding of the evolution of an Archean basin in the Slave Province. The region is currently an area of considerable interest for base metal exploration and recently for gold as well.

The existence of granitic basement at Point Lake was first recognized by Stockwell (1933) in an early reconnaissance expedition into the central part of the Slave Structural Province. The region of which this area is a part has been mapped at 1: 250 000 scale by Bostock (1976).

The area is situated on the western margin of an extensive basin of supracrustal rocks that is bounded on the west by a heterogeneous assemblage of granitic gneisses and schists (Fig. 44. 1). Basin is used in the sense of an outcrop area of supracrustal rocks. However, as has previously been suggested (McGlynn and Henderson, 1970) and as will become apparent later, the present outcrop basin margin may in part approximate the original depositional basin margin. The basin is open to the north, east and southeast beyond the borders of the map-area. The supracrustal rocks are part of the Yellowknife Supergroup and the formational nomenclature was established by Bostock (1976). All rocks in the area are Archean except for one small outlier of Goulburn (Aphebian) dolomite that occurs at the contact between the Itchen and Contwoyto formations about 2. 5 km north of Point Lake.

General Geology

Point Lake Formation

Mafic volcanic rocks of the Point Lake Formation occur along the western margin of the basin and consist of a relatively thin member of thin bedded mafic tuffs and a thick sequence of dominantly pillowed and massive mafic flows with minor medium to fine grained fragmental units of similar composition. Very minor and small dacite bodies also occur locally with the mafic flows. Thick coarse grained gabbro sills that are probably contemporaneous with the mafic volcanism are abundant within the volcanics in the southwestern and north-central part of the map-area. The mafic volcanic flows are similar to the mafic volcanic sequence in the Yellowknife area. The volcanic strata east of Keskarrah Bay locally contain mudstone units that are commonly carbonaceous, and in some cases sulphide-bearing. Some of these mudstone units contain thick, massive, very coarse grained (centimetre scale) turbidites of exclusively granitic provenance. These coarse, still porous, beds commonly contain sulphides (mainly pyrite) derived from the enclosing sulphide-rich mudstones.

The major felsic volcanic unit of the Point Lake Formation occurs in the central part of the area. Thin bedded tuffs and locally volcanic conglomerate of intermediate composition occur on the west side and generally massive but locally coarsely fragmental rhyolite on the east. Volcanic lenses of varied composition also occur within the sandstones and conglomerates of the Keskarrah Formation (shown as unpatterned lenses in Fig. 44. 1).

Contwoyto and Itchen Formations

The areally most extensive sedimentary units are the greywacke-mudstone turbidites that are characteristic of Yellowknife sediments throughout the Slave Province.

The Contwoyto Formation occurs in the north-central part of the map-area. Sedimentary structures characteristic of turbidity current deposition are well preserved. Composition of the greywackes is locally variable with more volcanogenic quartz-poor varieties commonly occurring near the volcanic rocks. In general, the sediments are more thinly bedded, more fine grained and have a greater mudstone component than similar sediments east of Yellowknife. The formation commonly contains units of thin bedded iron-formation of all facies. Oxide and carbonate facies occur in the west half of the outcrop area of the formation, and silicate iron-formation with locally associated sulphide iron-formation dominates in the east. Carbonate beds (both dolomite and limestone) also occur within the greywacke sequences and are commonly associated with iron-formation.

The Itchen Formation, another greywacke mudstone turbidite, differs from the Contwoyto in that iron-formation does not occur. Within the map-area the rocks are everywhere in medium metamorphic grade with assemblages of quartz-biotite-muscovite-cordierite-andalusite-sillimanite and garnet. In the southern half of its outcrop area the formation contains abundant pegmatites, locally tourmaline-bearing, and several very small, medium grained adamellite bodies.

[1]Department of Geology and Geophysics, University of Hawaii, Honolulu, Hawaii.

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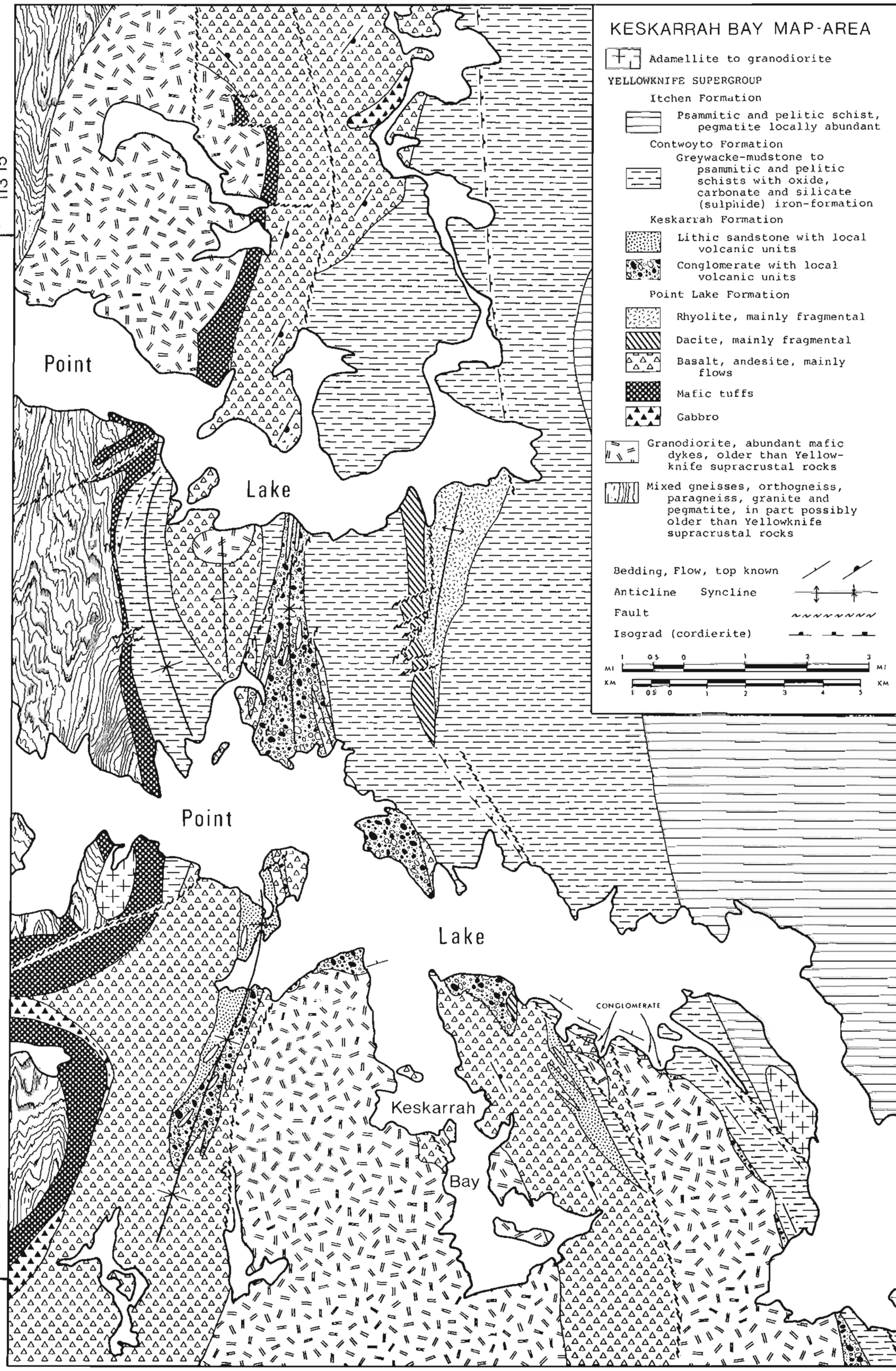

Figure 44. 1. Geological map of the Keskarrah Bay map-area.

Keskarrah Formation

The Keskarrah Formation consists of a conglomerate and a sandstone member. The conglomerate is composed of mafic volcanic and granitic clasts in relative proportions that reflect the local source. Granitic clasts predominate where the conglomerate occurs above the granitic basement, particularly in the eastern exposures. Within or adjacent to volcanic tuffs, the conglomerate cobbles are dominantly volcanic lithologies largely derived from the adjacent volcanic strata and reflect the local volcanic variation. The sandstone member locally interbedded with the conglomerate consists of fine to medium grained commonly cross-bedded lithic sandstone of predominantly felsic volcanic provenance. The medium to locally large scale cross-bedding in the sandstone and its close association with the thick massive conglomerates, suggests a subaerial environment of deposition for the formation. Intermediate and mafic volcanic rocks occur as lenses in both members of the Keskarrah Formation (unpatterned lenses in Fig. 44.1).

Granitic Basement

A massive, medium to locally coarse grained granodiorite characterized by abundant, commonly irridescent, quartz and by the presence of chlorite as its dominant mafic mineral, occurs in the southern and northwestern part of the map-area and as a small body between the two arms of Point Lake. The granodiorite is basement to the supracrustal succession. At many localities the unconformity is clearly exposed and is usually marked with a basal conglomerate derived in part from the underlying granodiorite (Henderson, 1975a, Fig. 4). Contact metamorphic effects suggesting an intrusive relationship with adjacent supracrustal rocks were not observed. In general, the granodiorite is massive, although in the northwestern body it becomes increasingly foliated towards the gneisses that lie to the west. In the southern body the granodiorite is locally gneissic in places west of Keskarrah Bay and it becomes increasingly cataclastically deformed to the south. Almost everywhere the granodiorite is cut by altered but undeformed diabase dykes. Similar dykes are present in the overlying supracrustal rocks, but are less numerous. In addition to these undeformed dykes, older, severely sheared mafic dykes occur locally that may have been feeders to the overlying Point Lake volcanic rocks.

At several localities where the unconformity is exposed, a weathered horizon can be seen in the underlying granodiorite. The weathered material is characterized by the extremely altered feldspar, although the original texture of the rock is preserved. At one locality a 15 cm layer of coarse quartz and sand presumably derived from this regolith occurs between the weathered granite and the succeeding conglomerate. Cobbles of granodiorite regolith are commonly found in the conglomerates, along with fresh granodiorite cobbles.

Gneisses and Schists

Along the west margin of the area is a heterogeneous assemblage of gneisses and schists. The rocks vary from orthogneisses, most common in the southwest corner of the map-area, to paragneisses, biotite schists and minor amphibolite, that commonly contain varied proportions of granitic to pegmatitic dykes and sills to the north. Adjacent to the contact with the Yellowknife supracrustal rocks the rocks of this unit are highly cataclastic to mylonitic. The degree of cataclasis of the gneisses declines to the west over a distance of 1 km from the supracrustal rocks. The extreme structural and metamorphic contrast between these gneisses and the adjacent supracrustal rocks suggest that this gneissic unit may be in part older than the Yellowknife rocks. However, no direct evidence of an unconformable relationship between these units was observed. Wherever it was exposed the contact was faulted.

Intrusive Granitic Rocks

Three small plutons that postdate the deposition of the Yellowknife rocks occur within the area. South of the west part of Point Lake a medium grained biotite adamellite, massive in the centre and strongly foliated at its margins, occurs between the Point Lake mafic tuffs and cataclastic gneisses to the west. Near the southeast corner of the map-area is a weakly foliated, medium grained biotite-muscovite adamellite with minor pegmatite. Abundant pegmatitic and granitic sills related to this pluton are present in the enclosing sediments. The edge of a third pluton composed of a massive, medium grained, coarsely porphyritic, biotite adamellite occurs at the southeast corner.

Structure

Throughout the area mapped the structural trend is generally northerly (Fig. 44.1). The Point Lake mafic volcanics in the northern part of the area occur in an east-facing, steeply dipping, homocline, but become increasingly more folded toward the south. The belt of mafic volcanic rocks east of Keskarrah Bay is also a steeply dipping, east-facing homoclinal succession. The Contwoyto and Itchen formations in contrast are commonly tightly folded into steep, often closely spaced, isoclines. In its main outcrop area the Keskarrah Formation occurs in a synclinal structure complicated by smaller-scale folds. Throughout the area all units have a northerly-trending cleavage of varied intensity. The major faults are also northerly trending. An older fault set with an easterly trend and relatively small displacement locally offsets the north-striking units.

Metamorphism

Isograds also trend northerly. The lowest metamorphic grade rocks are near the west margin of the basin. The grade rises both to the east and to the west. The cordierite isograd to the east occurs in the central part of the Contwoyto outcrop area and passes through the Point Lake felsic volcanics south of the northerly part of Point Lake. The assemblage biotite-muscovite-cordierite-andalusite, with or without garnet, is common.

Staurolite is rare. Garnet and amphibolite are spectacularly developed in the silicate iron-formation. Coarse sillimanite is common in parts of the Itchen Formation to the east. To the west there is a narrow zone of medium grade metamorphic rocks east of the gneisses, south of the north arm of Point Lake. The metamorphic as well as structural contrast between the relatively low grade supracrustal rocks and the cataclastic granitic gneisses adjacent to them, is striking.

Economic Geology

The area and surrounding region has been of considerable economic interest in recent years. The Izok Lake deposit of Texas Gulf, among the largest deposits in the Slave Province, occurs about 12 miles north of the northeast corner of the map-area. Within the map-area Zn-Cu mineralization occurs on the peninsula on the south shore of Point Lake west of Keskarrah Bay (Henderson, 1975a) in mafic volcanics within the Keskarrah Formation. Minor copper mineralization is locally present in black mudstone units of the Contwoyto Formation and also in the Point Lake mafic volcanics south of Point Lake. The silicate-sulphide facies iron-formations of the Contwoyto Formation have been actively prospected for gold. A more detailed discussion of the economic geology of the area is presented by Bostock (1976).

Intraformational Relationships

A major problem in dealing with the stratigraphy of Archean rocks is the difficulty of determining the intraformational relationships, even within a small area. The rocks are commonly structurally complex and in most parts of the Canadian Shield the typically steeply-dipping strata outcrop on an essentially horizontal plane. Without a reference horizon of some sort, it is difficult to relate one part of a basin with another, particularly across the structural trend of the various rock units.

In the Keskarrah Bay map-area, the unconformity surface between the granitic basement rocks and the overlying Yellowknife supracrustal rocks is such a reference horizon. South of Point Lake there is what amounts to a series of four northerly-trending "cross-sections" through the upper part of the Archean crust (Fig. 44.1). The bottom of each "section" is fault bounded with the sections facing east. The first section starts in the gneiss terrain (basement?) west of the map-area, passes eastward up through the mafic tuffs and flows of the Point Lake Formation and into the Keskarrah Formation, where it ends at the synclinal axial plane. The second section starts at the fault between the Point Lake volcanics and the basement granite west of Keskarrah Bay, and again extends eastward up through the Point Lake Formation, the Keskarrah Formation and into the Contwoyto Formation, where it ends against the fault. The third section is in the small triangular fault block and consists of basement granite, Keskarrah conglomerate, Point Lake volcanic rocks, more conglomerate, and finally, the Contwoyto greywacke-mudstone turbidites. The last section starts with the basement granite, passes into conglomerate which is followed by the Contwoyto Formation, with no volcanic rocks present in the section.

These four sections show the transition from the volcanic rocks that outcrop along the basin margin to the sediments that occur in the central part of the basin. The basement granodiorite lies unconformably below both the volcanics and sediments of the Yellowknife Supergroup. The granodiorite contributed detritus while both volcanism and sedimentation were taking place, as conglomerate and detritus of granitic provenance are found in both the volcanic and sedimentary units in the vicinity of the granitoid basement. All the formations of the Yellowknife Supergroup are essentially contemporaneous; the volcanic rocks are interbedded with the Keskarrah sediments and the contact between the Keskarrah and the Contwoyto formations, where exposed, is abruptly gradational. Thus, there are no significant internal unconformities within the Yellowknife Supergroup. In the westernmost section the mafic volcanics are thickest and no Contwoyto sediments are present. To the east into the basin the volcanic section becomes progressively thinner, until in the easternmost section no volcanics are present. Apparently, the mafic volcanism is restricted to what is the present-day margin of the basin and does not underlie the basinal turbidites of the Contwoyto and Itchen formations. This further supports the suggestion that the Archean supracrustal basins in the Slave Province were ensialic (McGlynn and Henderson, 1970).

Comparison between Keskarrah Bay map-area and the Yellowknife Region

The similarity between the Archean geology of the Keskarrah Bay area and the Yellowknife region (Henderson, 1975b, 1976) 200 miles to the south, is striking. The gneissic terrain south of Yellowknife and east of Ross Lake in the Yellowknife region corresponds to the gneisses in the Keskarrah Bay area, both of which are considered to be in part older than the supracrustal rocks (Henderson, 1976). No plutonic basement analogous to the granodiorite basement at Keskarrah Bay has been recognized near Yellowknife, but the plutonic body mantled by the mafic volcanics near Sunset Lake (Henderson, 1976) bears a close similarity in composition and texture to the basement granite in the Keskarrah Bay area. A re-examination of that granodiorite is perhaps in order. The supracrustal units of the Yellowknife Supergroup in both areas are similar, with the Kam and Duck formations at Yellowknife corresponding to the mafic parts of the Point Lake Formation, the Banting Formation with the intermediate and felsic members of the Point Lake, and the Burwash and Walsh greywacke turbidites with the Itchen and Contwoyto formations. Iron formations are absent in the sedimentary sequence at Yellowknife. The conglomerate and subaerial sandstones of the Keskarrah Formation are analogous to the essentially identical lithologies in the Jackson Lake Formation at Yellowknife. One difference between the two areas is that the Yellowknife area is perhaps less "complete" than Keskarrah Bay area and the stratigraphic relationships that have to be inferred at Yellowknife (Henderson, 1975b) are more clearly apparent around Keskarrah Bay. The similarity between the two areas, however, does not suggest that the various formations are physically correlative, but does indicate that similar sedimentological and volcanological events were taking place in similar environments in the two areas.

Acknowledgments

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STRATIGRAPHIC AND STRUCTURAL RELATIONS BETWEEN THE SELWYN BASIN, PELLY-CASSIAR PLATFORM, AND YUKON CRYSTALLINE TERRANE IN THE PELLY MOUNTAINS, YUKON

Project 730037

D.J. Tempelman-Kluit
Regional and Economic Geology, Vancouver

Introduction

This study of the stratigraphic and structural relations within and between the Omineca Crystalline Belt, Pelly-Cassiar Platform, Selwyn Basin and Yukon Crystalline Terrane has focussed on Quiet Lake (105 F) and Finlayson Lake (105 G) map-areas where exposure is generally excellent. During the 1976 field season mapping was completed in Quiet Lake and Finlayson Lake map-areas for publication at 1: 250 000 scale with detailed study of selected areas for publication at 1: 50 000. S.P. Gordey spent three weeks completing field work for his doctoral dissertation at Queen's university. G. Abott, S. Gordey, K. Pivnick, D. Moncrieff and R. Wahlgren provided able assistance in the field. This summary, together with earlier reports (Tempelman-Kluit *et al.*, 1975, 1976), is a preliminary account of the geology of the Pelly Mountains.

General Setting

The shallow marine miogeoclinal sequence found in the Pelly Mountains (Tempelman-Kluit *et al.*, 1975, 1976) occupies an area up to 70 km wide that extends southeast to the Cassiar Mountains in northern British Columbia, a distance of 600 km. This northwest-trending belt of platform carbonates and related rocks ranges in age from Cambrian through Mississippian and has been referred to as the Pelly-Cassiar Platform (Gabrielse, 1967). Northeast of the platform carbonates are time equivalent shales that constitute Selwyn Basin. Southwest of the Platform are metamorphosed shale, quartzite and volcanic rocks, also time equivalents of the carbonates of the Pelly-Cassiar Platform. These metamorphic rocks are covered locally by late Paleozoic serpentinized peridotite, basalt and chert thought to have been thrust over them. The metamorphic rocks and the overthrust peridotite

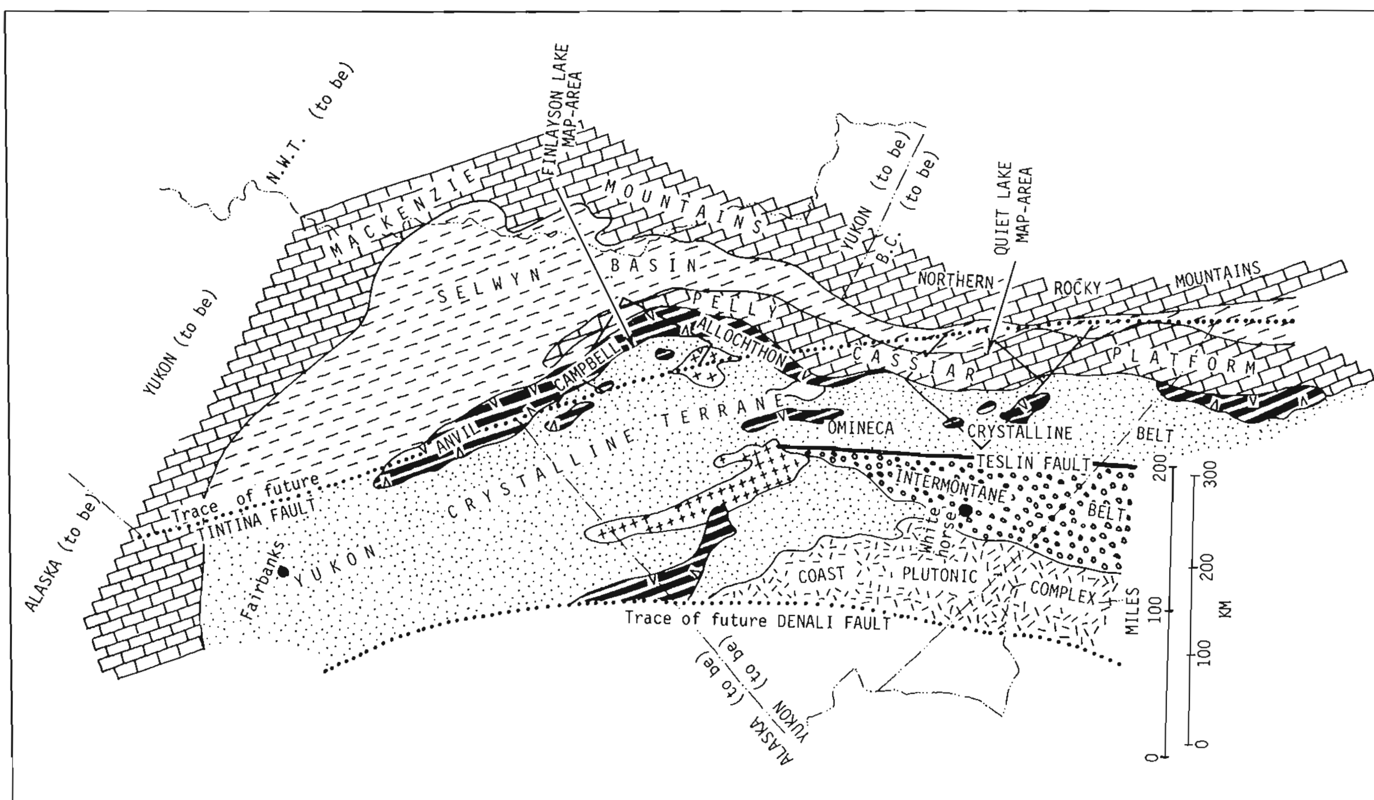

Figure 45. 1. Distribution of major tectonic elements in a part of the northern Cordillera about 100 m. y. ago prior to transcurrent movement on the Tintina Fault. The reconstruction is made by restoring the right lateral slip of 450 km along the trace of Tintina Fault. Note that Quiet Lake and Finlayson Lake map-areas, now adjacent to each other, give cross-sections of the same tectonic elements which were originally about 350 km apart along the depositional strike.

From: Report of Activities, Part A;
Geol. Surv. Can., Paper 77-1A (1977).

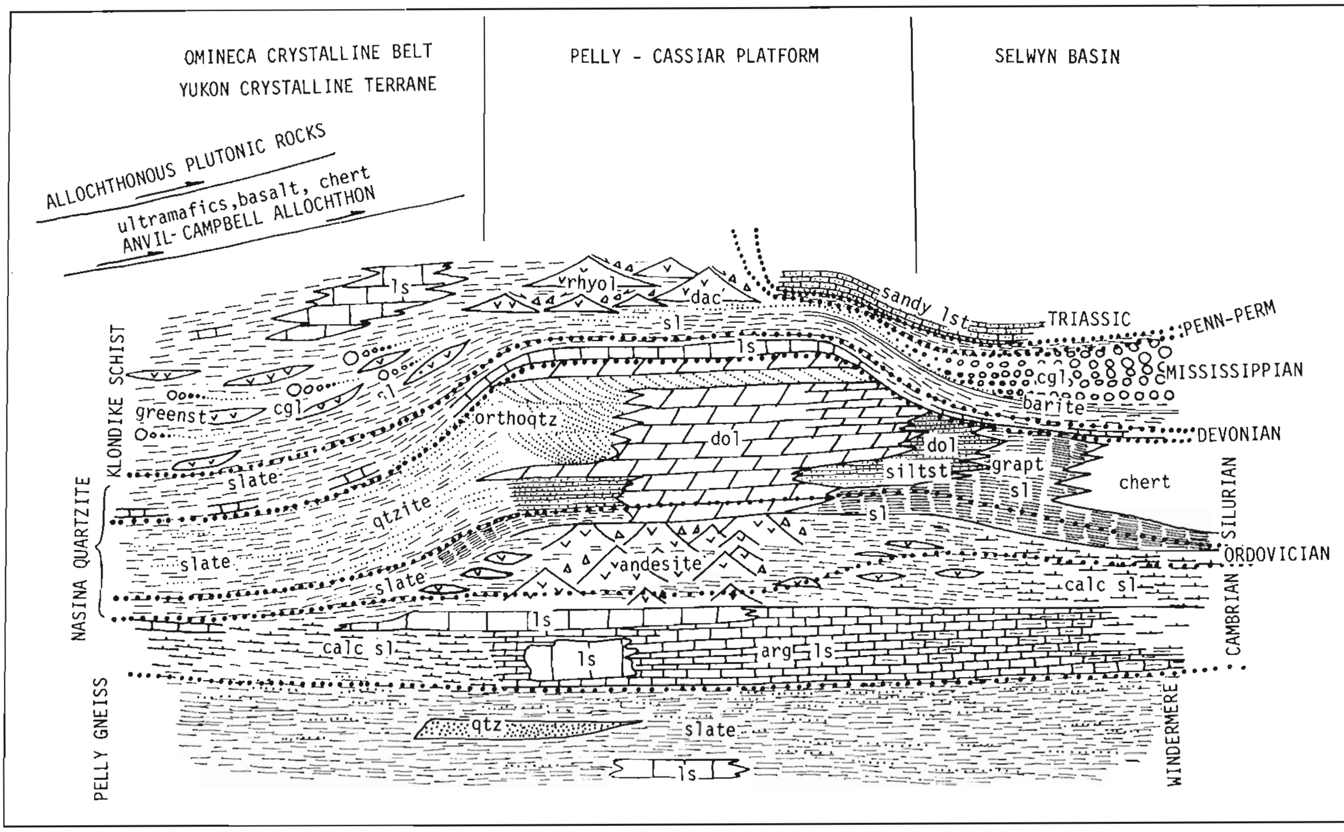

Figure 45. 2. Restored section across the Pelly-Cassiar Platform through Quiet Lake map-area to illustrate the facies relations of the main stratigraphic units with those found in the flanking tectonic elements, Selwyn Basin and the Omineca Crystalline Belt-Yukon Crystalline Terrane. Time lines shown by heavy dots across the diagram are only approximately located in the Omineca Crystalline Belt part of the diagram because no diagnostic fossils have been found there. (Compare with Fig. 21. 1 of Tempelman-Kluit et al., 1976).

and basalt constitute the Omineca Crystalline belt and its northwestward continuation, the Yukon Crystalline Terrane. In southern Quiet Lake and Finlayson Lake map-areas the metamorphic rocks together with the overthrust ultramafic and mafic rocks are thrust northeastward over upper Triassic rocks at the southwest edge of the Pelly-Cassiar Platform. The Platform itself is internally repeated by folds and northeast directed thrust faults which involve Upper Triassic strata. The entire foreshortened assemblage is intruded extensively by mid-Cretaceous granitic rocks. Late Cretaceous right lateral movement of 450 km along the Tintina Fault has displaced the tectonic elements relative to each other. Figure 45. 1 is a reconstruction of their distribution prior to the transcurrent movement.

Pelly-Cassiar Platform

The reconstruction in Figure 45. 1 emphasizes that, although the same tectonic elements are adjacent in the northern part of Finlayson Lake and the southern part of Quiet Lake map-area on either side of the Tintina Trench, these two areas were originally a considerable distance apart. In Finlayson Lake map-area northeast of Tintina Trench a cross-section extends from the edge of the Selwyn Basin across the northern extremity of the Pelly-Cassiar Platform into the Yukon Crystalline Terrane whereas in southern Quiet Lake map-area a cross-section intersects the same tectonic elements that were originally some 350 km south along the depositional strike. The succession and facies relations of the Pelly-Cassiar Platform in Quiet Lake map-area shown schematically in Figure 21-1 of Tempelman-Kluit *et al.* (1976) are generally valid for the extension of the Cassiar Platform in northern Finlayson Lake map-area. However Lower Cambrian strata are not so widely developed and tend to be dominantly shaly limestone with few thick limestones like those seen locally in the Pelly Mountains. Similarly the Siluro-Devonian carbonate sequence is generally thinner in northern Finlayson Lake map-area than in southern Quiet Lake map-area. The Mississippian acid volcanic rocks so prominent in parts of Quiet Lake map-area are commonly absent in northern Finlayson Lake map-area although the orange weathering chert, which is its lateral equivalent, is widespread. The condensed sequence of rocks that ranges from Pennsylvanian to Triassic seen in Quiet Lake map-area is also widespread north and west of the Pelly-Cassiar Platform in Finlayson Lake map-area.

Omineca Crystalline Belt

The metamorphic rocks southwest of Pelly-Cassiar Platform constitute the Omineca Crystalline Belt and also make up its northern extension, the Yukon Crystalline Terrane (Figure 45.1). Original layering in these rocks is so strongly transposed on near-horizontal slip surfaces that primary layering is largely obliterated. However, because the deformation has been so penetrative, strain has occurred internally and the original sequence of gross lithological units has been preserved although depositional thicknesses are probably considerably modified. In southern Quiet Lake map-area the penetratively deformed metamorphic rocks are thrust onto the Pelly-Cassiar Platform so that the transition from deformed and metamorphosed rocks to unmetamorphosed rocks is abrupt and foreshortened. In northern Finlayson Lake map-area this transition is gradational and apparently not foreshortened. It coincides roughly with the sedimentary facies boundary that separates the Pelly-Cassiar Platform from the Yukon Crystalline Terrane. Although details of the interrelations between the various stratigraphic units in Pelly-Cassiar Platform and the Yukon Crystalline Terrane remain to be worked out, it is possible in a general way to correlate rock units across this facies boundary. Figure 45.2 illustrates these correlations. No fossils have been discovered in the transposed metamorphic rocks to substantiate these correlations and time lines drawn across from the Pelly-Cassiar Platform into the Omineca Crystalline Belt are therefore approximate.

The metamorphic sequence in southern Quiet Lake map-area includes more extensively developed Silurian

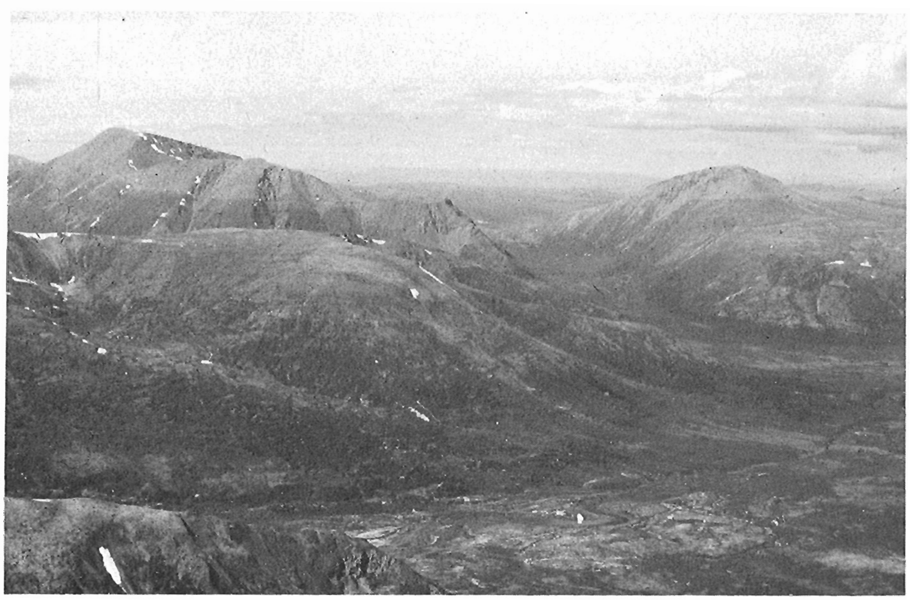

Figure 45.3.

Serpentinized peridotite (Anvil-Campbell allochthon) caps the mountains in this view of part of east central Finlayson Lake map-area and lies here on schist and quartzite of the Klondike Schist. Klippen like these are seen at many places in the Pelly Mountains on both sides of Tintina Fault (compare this with Fig. 21.11 of Tempelman-Kluit *et al.*, 1976).

Figure 45.4.

Pink quartz monzonite overlies sheared basalt and serpentinized peridotite of the Anvil-Campbell Allochthon in this view in southwest Finlayson Lake map-area. The contact is sharp and consists of a polished slickensided surface; above it the quartz monzonite is shattered and recemented in a zone about 100 m thick and below it the basalt is sheared. In contrast to some other allochthonous masses of plutonic rocks, which are accompanied by much mylonitzation, displacement here evidently occurred at high structural levels judging by the brittle style of structural response.

to Devonian black graphitic quartzite, slate and minor limestone than is seen in northern Finlayson Lake map-area in the equivalent interval. Similarly the schist and gneiss considered the equivalent of Windermere and Lower Cambrian greywacke and calcareous shale, which is seen extensively in Finlayson Lake map-area, is not exposed in Quiet Lake map-area.

The dark grey slate, volcanic rocks (now chlorite schist), blue quartz and feldspar conglomerate and the Klondike schist, their more metamorphosed and sheared equivalents, are extensive in the southern parts of both Finlayson Lake and Quiet Lake map-areas. These rocks are the Englishman's Group of Mulligan (1963). Although they resemble gritty rocks of Windermere age found extensively in the Cordillera they are a separate unit as their stratigraphic position shows.

The slates, volcanic rocks and conglomerates are broadly correlative with the Upper Devonian and Mississippian black slate and chert pebble conglomerate known informally as the "Black Clastic". The provenance of these two roughly time equivalent units is distinct. The "Black Clastic" contains much detrital chert with less quartz and very minor feldspar and is derived locally from the Road River chert basin (Blusson, 1976) because its coarsest members are arranged concentrically around this basin and generally fine away from it. The blue quartz and feldspar grits are arkoses in many places and contain no chert. They were probably derived from the southwest from an igneous terrane no longer present or much modified.

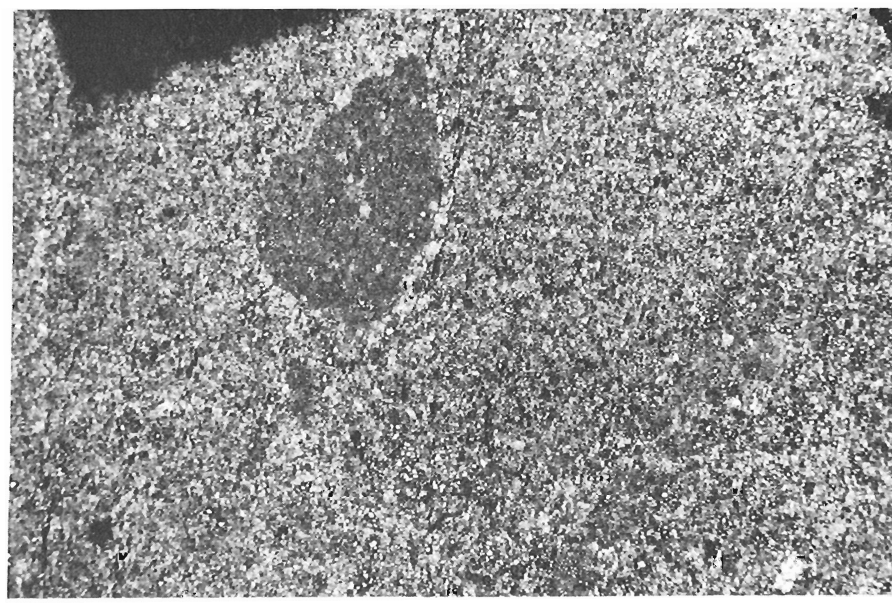

Figure 45.5.

Hornblende granodiorite like that shown here makes up the bulk of a klippe at least 18 km across that is thrust above serpentinized peridotite in southeast Finlayson Lake map-area. Similar rocks occur in adjacent Watson Lake and Frances Lake map-areas where the structural relations may be identical. The granodiorite resembles parts of the Klotassin batholith in the Yukon Crystalline Terrane. The dark inclusion is about 3 inches long.

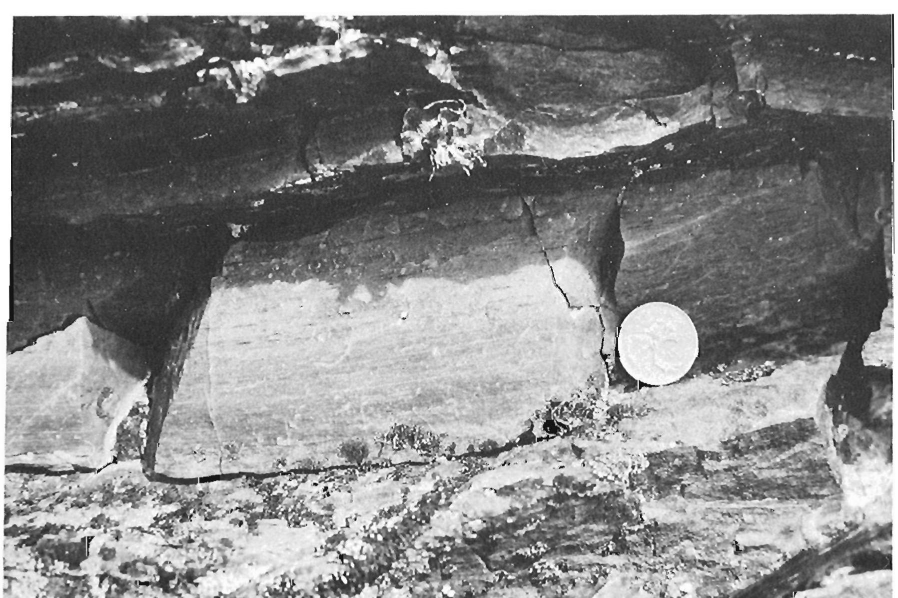

Figure 45.6.

Ultramylonite as seen in places within and near the margins of the allochthonous hornblende granodiorite. This mylonite grades over a distance of 3 m into the fresh hornblende granodiorite illustrated in Figure 45.5. Elsewhere the mylonite zones are hundreds of metres thick. The degree of cataclasis within the mylonite zones varies from place to place.

Allochthonous Rocks of the Omineca Crystalline Belt

Ultramafic rocks, basalt and chert of the Anvil-Campbell allochthon are presumed to be thrust onto the metamorphic rocks in Finlayson Lake map-area, a relationship similar to that seen in southern Quiet Lake map-area at Dunite Mountain (compare Fig. 45.3 with Fig. 21-11 of Tempelman-Kluit *et al.*, 1976). A number of bodies of mafic and ultramafic rocks that are scattered through Yukon Crystalline Terrane are probably klippen of the same allochthon. In turn, above the mafic and ultramafic allochthonous rocks are thrust batholith sized bodies of plutonic rocks now preserved as klippen (Fig. 45.4). These include medium grained equigranular hornblende granodiorite to quartz diorite and porphyritic (pink K-feldspar) medium grained biotite quartz monzonite. In places, both at the margins of klippen and within them, the granitic rocks are mylonites and blastomylonites. Elsewhere the rocks are less sheared and show only a weak cataclastic texture so that every gradation from mylonite to a granitic-textured rock is seen (*see* Fig. 45.5 and 45.6). The orientation of the mylonite zones is generally near horizontal and their thickness ranges from centimetres to hundreds of metres. Klippen of pink quartz monzonite are not generally sheared and instead lie on polished planar surfaces above which the quartz monzonite is only shattered and recemented in a zone about a hundred metres thick that grades upward into fresh granitic-textured unbroken rock. The ages of the hornblende granodiorite and pink quartz monzonite are unknown, but they resemble rocks of the Klotassin and pink quartz monzonite suites, respectively, in Yukon Crystalline Terrane (Tempelman-Kluit and Wanless, 1975). A thrust relationship has not been demonstrated beneath these plutonic rocks in Yukon Crystalline Terrane. However such a relationship there is not incompatible with the field evidence.

Mineral Deposits

No new mineral occurrences were discovered, but several showings in southern Finlayson Lake map-area were briefly examined to determine their stratigraphic association. The Hoo, Fyre, North Lake and Wolverine Lake base metal showings all occur in the sequence of dark slates with greenstones and gritty rocks here considered to be Devono-Mississippian. Several small base metal occurrences are also known in equivalent rocks in other parts of the Yukon Crystalline Terrane (Lucky Joe Creek, Loon Lake). No mineralization has yet been discovered in the schist and gneiss thought to be equivalent to the Windermere-Lower Cambrian or in the graphitic quartzite sequence known informally as the Nasina quartzite and thought to be Ordovician to Silurian. The ultramafic and mafic rocks of the Anvil-Campbell allochthon are apparently barren of sulphide mineralization although all known asbestos occurrences are confined to this assemblage. Similarly the plutonic rocks thrust above the Anvil-Campbell allochthon contain no known sulphide occurrences.

Mineralization at the Hoo and Fyre is prominently layered and this layering is parallel with, and similar to, the strong foliation of the rocks implying that sulphides were emplaced before the deformation and metamorphism of the rocks. Local controls to the mineralization at these occurrences are unknown, but at the Hoo a spatial association of mineralization with a thin limestone has been noted.

Small bodies of quartz feldspar porphyry and other intermediate and acidic volcanic rocks, which are thought to be equivalent of the South Fork volcanics, occur in places in Finlayson Lake map-area. They are particularly extensive 16 km northwest of Fire Lake. These rocks locally contain much disseminated pyrite, but no other sulphide minerals were noted.

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Project 750014

G. H. Eisbacher
Regional and Economic Geology Division, Vancouver

Introduction

Copper mineralization in Proterozoic basins of the Mackenzie Mountains has been known since 1962 when the first showings were discovered in the Redstone area. Since then geological studies and prospecting have extended the knowledge of potentially copper bearing strata and related units towards the northwest into the Keele River area (1971) and the Hayhook Lake region (1975). During 1976 field work rocks of the Redstone Formation, the Coppercap Formation, and the Rapitan Group were studied as far as Mountain River (Fig. 46. 1). Part of this work was carried out in conjunction with J. D. Aitken to whom the author of this report is indebted for stimulating discussions of Proterozoic geology.

The outcrop pattern of the copper-bearing Redstone and Coppercap formations is controlled by (a) original eastward embayments of the sedimentary basins, and (b) westward erosion of the overlying Rapitan Group. The lateral extent of Redstone and Coppercap rocks, now known to be more than 300 km along strike, suggests stongly that the Redstone and Coppercap basins were only slightly more restricted than the subsequent Rapitan basin. Faults active during deposition of the Redstone Formation also seem to have influenced facies pattern in the Rapitan Basin. To demonstrate the mappable relationships between tectonics and sedimentation two areas will be discussed in some detail: the Coates Lake area and the Mountain River area. As in other parts of the outcrop area the formations have been elevated and tilted above major Laramide(?) thrust faults, and form prominent erosional scarps. The stratigraphic units involved are briefly summarized in Table 46. 1.

Table 46. 1

Sheepbed Formation – shale

Keele Formation – limestone, dolomite, quartzite

"Shale" Formation (Rapitan 3) – siltstone, shale, limestone Rapitan Group

"Diamictite" Formation (Rapitan 2) – glaciomarine conglomerate

"Maroon" Formation (Rapitan 1) – siltstone, breccia, iron-formation

Coppercap Formation – limestone, dolomite

Redstone Formation – siltstone, breccia, conglomerate, evaporite, dolomite

Little Dal Formation – dolomite, stromatolites

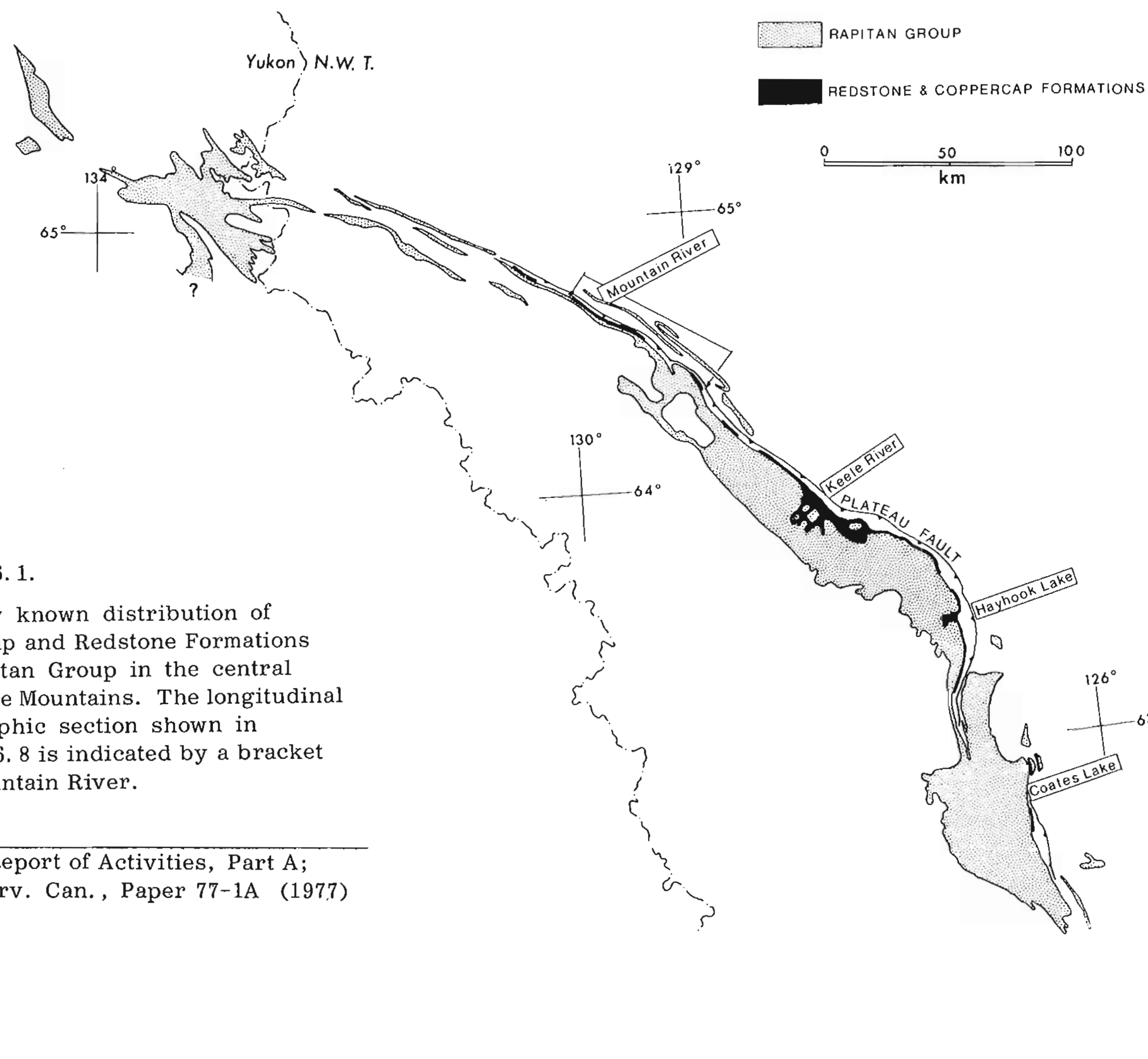


Figure 46. 1.

Presently known distribution of Coppercap and Redstone Formations and Rapitan Group in the central Mackenzie Mountains. The longitudinal stratigraphic section shown in Figure 46. 8 is indicated by a bracket near Mountain River.

From: Report of Activities, Part A; Geol. Surv. Can., Paper 77-1A (1977)

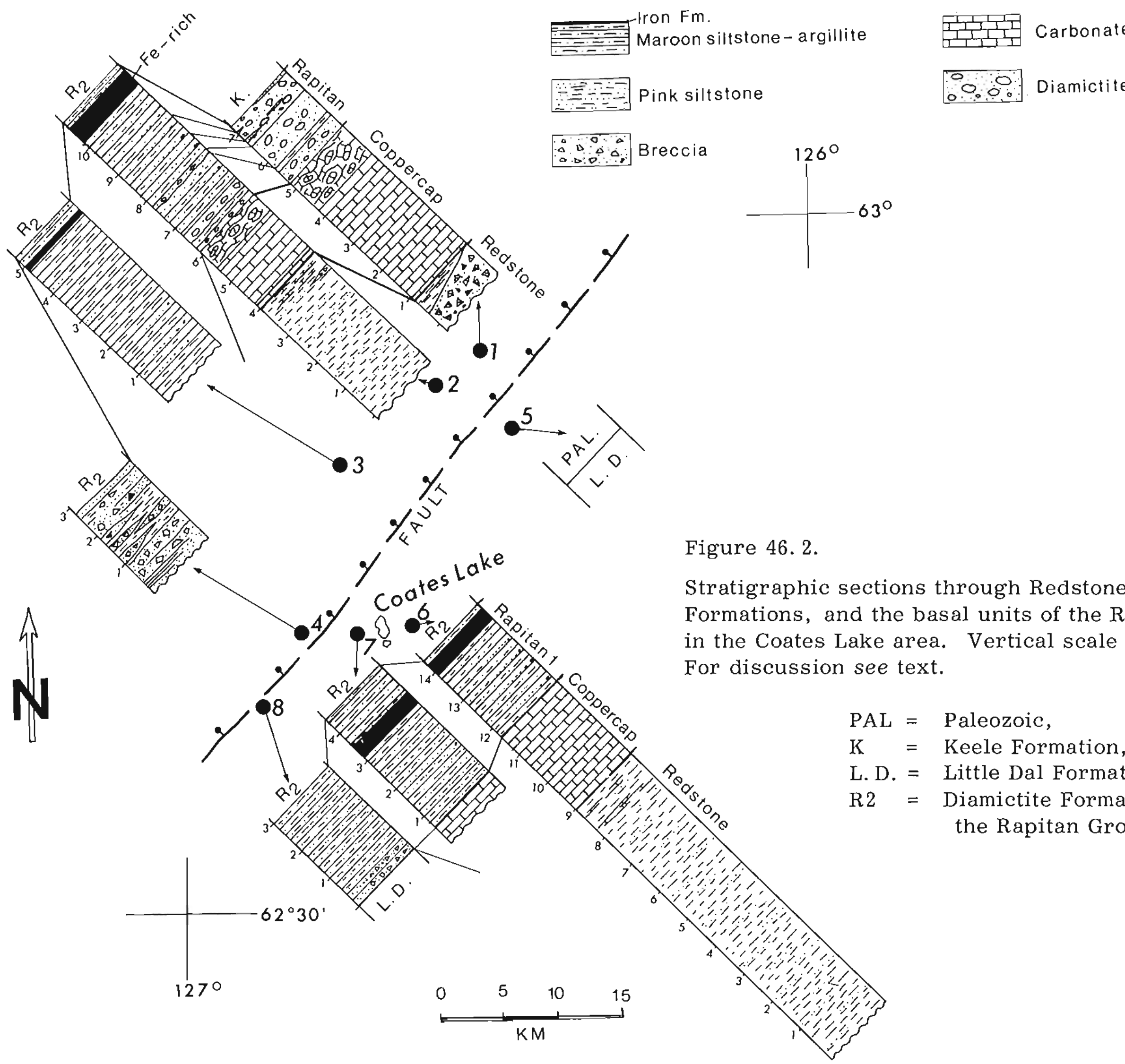


Figure 46. 2.

Stratigraphic sections through Redstone and Coppercap Formations, and the basal units of the Rapitan Group in the Coates Lake area. Vertical scale is 100 m. For discussion *see* text.

PAL = Paleozoic,
K = Keele Formation,
L. D. = Little Dal Formation,
R2 = Diamictite Formation of the Rapitan Group.

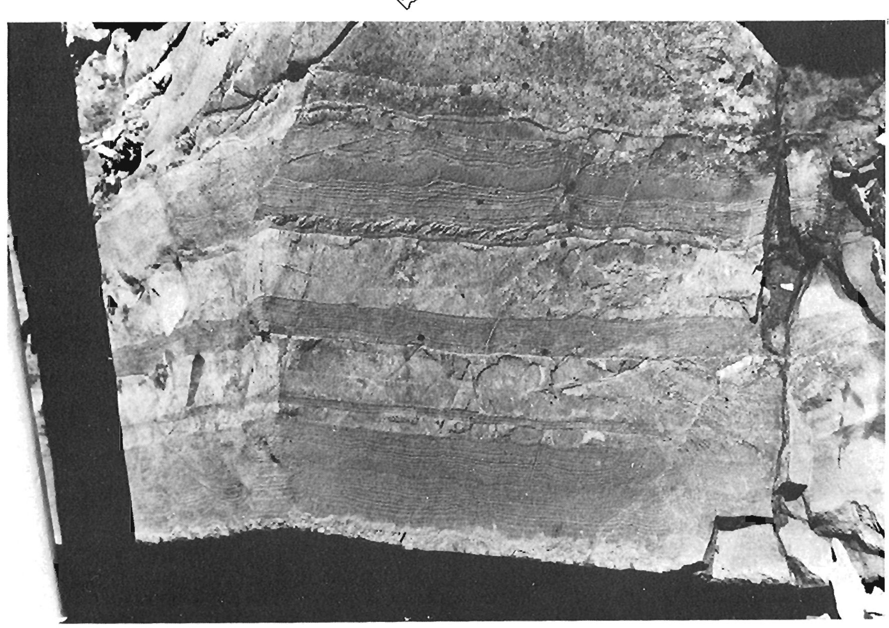


Figure 46. 3.

Dolomitic cryptalgal laminites and two bands of dolomitic siltstone in the Redstone Formation of Section 1 (Fig. 46. 2). Vertical distance covered in photograph is 15 cm.

Figure 46. 4.

Distinctly graded allodapic limestones of the Coppercap Formation (Section 1). Flat-edge pebble conglomerates consist generally of reworked dolomitic micrite clasts in a matrix of calcarenite. Part of camera cap at the top of picture for scale.

Figure 46. 5.

Slump fold in graded allodapic limestone of the Coppercap Formation (Section 1). White scale bar on pole is 10 cm.

Figure 46.6.

Boulder of Redstone breccia in slightly cleaved red matrix of Rapitan diamictite of Section 1. White scale bar on pole is 10 cm.

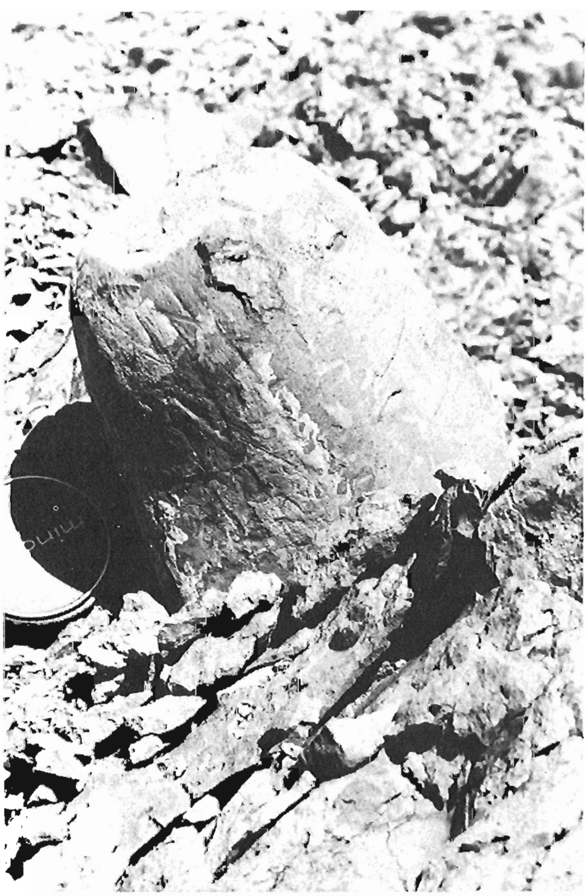
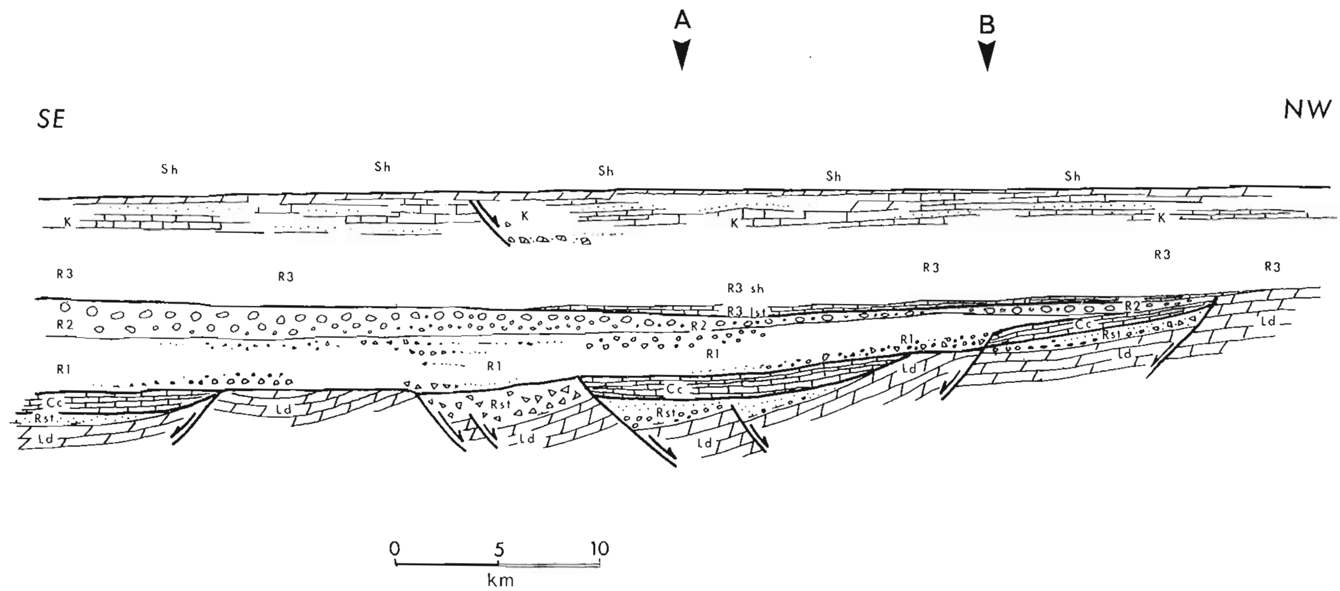


Figure 46.7. Glacially striated cobble of dolomite in Rapitan diamictite of Section 1.

Coates Lake

Mapping and stratigraphic-sedimentologic study of the original discovery area demonstrates a complex synsedimentary tectonic control of the basin which affected sedimentary facies and later structural evolution. The principal element is a steep northeasterly trending fault (Fig. 46.2). The principal displacement during Proterozoic sedimentation was normal, with the northwest side down dropped. There might have been a small right-lateral shift as well.

Section 1 is northeast of Coates Lake where the formations were previously mapped with older Proterozoic units (Gabrielse *et al.* 1973). Exposures display an apparent northwestward intertonguing of a fanglomerate breccia with typical pink siltstones and brown dolomitic siltstones of the Redstone Formation. The fanglomerate thickens southeastward where it directly overlies Little Dal Formation. It must have been derived from a high-standing fault scarp to the south where the entire Redstone – Coppercap – Rapitan section is missing (Section 5). The dolomitic siltstones above the breccia are interbedded locally with cryptalgal laminites (Fig. 46.3) and probably represent deposits of a supratidal mudflat. Some of the calcareous cryptalgal laminites are stained with malachite. Overlying the Redstone Formation are thin-bedded detrital dolomites and limestones (Fig. 46.4) with shale partings between the carbonate beds. As in other areas of the Redstone Copper Belt both the dolomitic layers near the base and the graded limestones above are allodapic and were probably deposited by turbidity or density currents. The allodapic dolomite must have been derived from submerged dolomitic silts and sands of the Redstone Formation. The base of the Coppercap Formation thus is definitely a transgressive surface and the gradual change to detrital limestones probably indicates deepening



Ld	= Little Dal Formation,	Rst	= Redstone Formation,
Cc	= Coppercap Formation	R1	= Maroon Formation (Rapitan 1),
R2	= Diamictite Formation (Rapitan 2),	R3 lst	= laminated, dark grey limestone of the Rapitan Group,
R3sh	= Shale Formation of the Rapitan Group,		
K	= Keele Formation,	Sh	= Sheepbed Formation

Figure 46. 8. Longitudinal cross-section in the Mountain River area illustrating the possible influence of comtemporaneous faults on facies in Proterozoic rocks. Vertical scale is schematic.

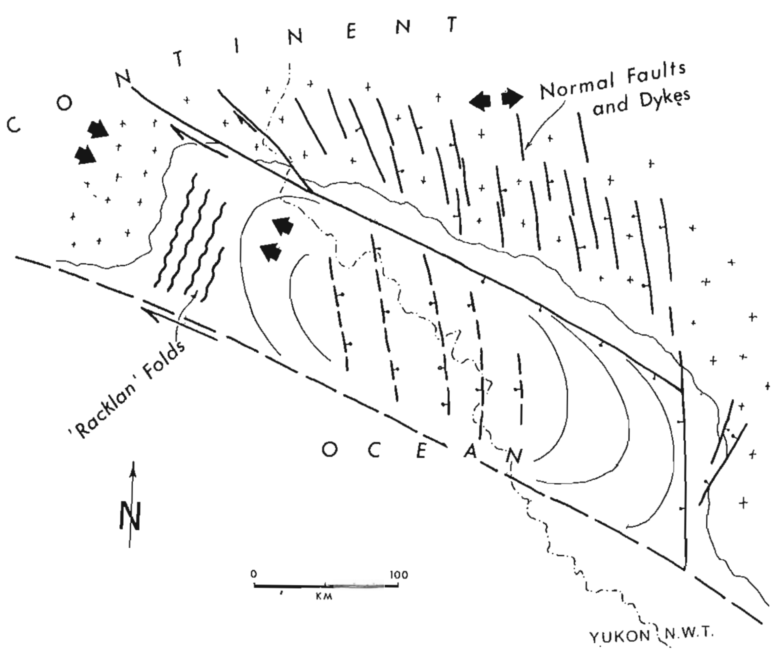


Figure 46. 9. Preliminary tectonic model for the late Proterozoic basin of the Mackenzie Mountains. A dilation of the continenal crust is indirectly related to a slowly stretching ocean basin south of a trans-current fault. The Racklan Orogeny is a direct outcome of this process.

of the basin. Slump folds (Fig. 46. 5) suggest a west-northwest dipping paleoslope. The Coppercap Formation is topped by a grey, marly diamictite consisting predominantly of subrounded limestone cobbles distributed chaotically in a calcareous matrix. Upsection this unit grades into brick-red to maroon diamictite of the Rapitan Group. The latter contains cobbles and boulders of dolomite, limestone, quartzite, and greenstone. Cobbles with copper mineralization and a boulder of the distinct Redstone breccia (Fig. 46. 6) suggest that diagenesis and copper mineralization in the Redstone Formation were well advanced by the time the Rapitan Group was laid down. Sporadic cobbles display remarkable glacial striae (Fig. 46. 7) known from other sections of the Rapitan Basin (Ziegler, 1959; Eisbacher, 1976; Young, 1976).

In Sections 2 and 6, to the west, the Redstone Formation thickens and consists of anhydrite-cemented laminated siltstones with mudcracked bedding surfaces. Cross-lamination at section 6 indicates northerly flowing paleocurrents. The Maroon Formation of the Rapitan Group (Rapitan 1) also increases dramatically in thickness and only occasional brick-red diamictite layers in the siltstone-argillite succession suggest a correlation between Sections 2 and 1. Iron-formation occurs near the top of the Maroon Formation.

At Section 4 thick beds of angular cobbles and boulders of dolomite, limestone, and reworked Rapitan Group, including fragments of iron-formation, make up the bulk of the Maroon Formation. These peculiar "sharpclast – siltstones" were probably derived from south of the fault where stripping of Coppercap, Redstone and Little Dal formations is suggested in Section 8. Paleocurrent data from the fine grained layers indicate northward flow. Syndepositional displacement along the northeasterly trending fault was active during Redstone and Rapitan time in the Coates Lake area.

Mountain River

Mapping reveals abrupt facies changes along strike of the upper Proterozoic Formations (Fig. 46.8). Synsedimentary tectonics must have been particularly active during Redstone time. The stratigraphically deepest part of the Redstone Formation (at A Fig. 46.8) consists of evaporites. These are overlain by conglomerates or dolomitic siltstones and minor dolomitic cryptalgal(?) laminites. The Coppercap Formation begins with detrital dolomite beds, and grades upward into graded limestone beds and cherty dolomite. Copper mineralization near the contact between the Redstone and the Coppercap Formation was found at 129°33'00"W, 64°35'30"N.

Faulting continued in Rapitan time although in some areas the field relationships suggest that the throw was reversed (e.g. B, Fig. 46.8). The edges of anomalous facies in the Rapitan Group, such as the detrital limestone laminites (R3 lst) overlying the glaciomarine diamictites (R2) also seem to be localized by hinges above deep faults. Spectacular olistoliths of Keele limestones within shales of Rapitan character (R3) were also shed from a northerly trending synsedimentary scarp. Synsedimentary movements therefore extend throughout the time interval of Redstone and Coppercap formations, and Rapitan Group, along faults that ran oblique to the later compressional structures of the Laramide Orogeny.

During the deposition of the Keel Formation the early structural trends in general became subsidiary to the regional southwestward dipping paleoslope in the area. Towards the southwest the carbonates form aprons of subaqueous debris flow deposits and the boundary between the Rapitan shales and the Keele Formation is rarely sharply developed. For purposes of regional correlation the Keele Formation will be included within the Rapitan Group (Eisbacher, in prep).

Tectonic control of the Redstone, Coppercap and Rapitan basins

The hypothetical scheme for the origin of the Redstone Basin and the strata above is given in Figure 46.9. Aitken and Cook (1974, and pers. comm.) have amply documented north-northwest trending normal faults which are commonly intruded by diabase dykes and which cut rocks as young as Little Dal Formation in the northern Mackenzie Mountains. Northeast trending in folds have been mapped by Wheeler (1954) in the Yukon Territory to the west. Gabrielse (1967) suggested that the folds were created during a 'Racklan Orogeny' and that they are overlain by Rapitan-like rocks. Observations made during a visit to the type area by the author and J.D. Aitken strengthened the possibility that Rapitan, Coppercap and Redstone equivalents may indeed overlie older folded dolomitic rocks, deformed during a 'Racklan Orogeny'. D.K. Norris (pers. comm. 1975) points out that northwesterly trending right-lateral faults in northern Yukon Territory may have great antiquity. Taken these seemingly diverse pieces of evidence and considering the extensional tectonics of the fanglomerate – evaporite-carbonate setting of the Redstone Formation, it might be suggested that a broadly dilating crust bounded by a transcurrent fault was the basement of the Redstone, Coppercap, and Rapitan basins. The process of crustal thinning and rifting may, at the same time, have caused shortening and folding along a continental buttress to the northwest ('Racklan Orogeny'). The Redstone Formation thus signifies the initiation of a fault-controlled basin that maintained a similar shape throughout the rest of Proterozoic and Paleozoic time (Cordilleran Miogeosyncline).

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47. ROCKSLIDES IN THE MACKENZIE MOUNTAINS, DISTRICT OF MACKENZIE

Project 760059

G.H. Eisbacher
Regional and Economic Geology Division, Vancouver

Introduction

During the field season of 1975 and 1976 the author began a study of large postglacial rock slides in the central part of the Mackenzie Mountains. Preliminary airphoto scanning suggested high incidence and a remarkable degree of preservation of large slides and debris streams along the valley bottoms of the Backbone Ranges, Central Mackenzie Mountains. The Mackenzie Mountains form part of the Eastern Cordilleran Fold Belt: well exposed folded and faulted Proterozoic and Paleozoic successions protrude as long ridges, dip-slopes, and scarps from drift-covered glacial valleys. Rock slides are particularly common in massive Paleozoic carbonate rocks.

At present, increasing mining exploration and the possible construction of ground transportation routes justify a careful consideration of the rock slide hazard in this, as yet, remote region of the Canadian Cordillera. Only three localities have been chosen for discussion in this preliminary report. Other instructive slide areas were studied or briefly visited and will be described in forthcoming papers. However, the three examples chosen do illustrate most of the features encountered, particularly the aspects of "streaming behavior" of the debris after the initial collapse of rock faces. The three localities are shown on the index map (Fig. 47.1), and precise co-ordinates are listed with the subheadings below.

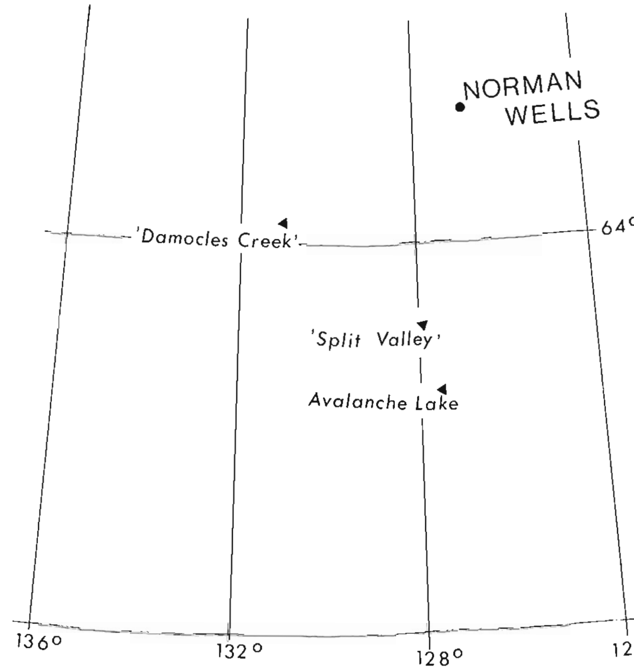

Figure 47.1. Index map showing the location of three rock slide areas discussed in the report.

A) "Damocles Slide" (131°08', 64°17')

The Damocles Slide is situated near the Yukon – District of Mackenzie border in an area lacking named topographic features. The rocks involved are massive carbonate banks of Early Cambrian age which grade latreally into argillaceous limestones and shales.

The Damocles Slide originated on the north side of a glacial valley by the collapse of a massive carbonate cliff (Fig. 47.2 and 47.4). The scar left behind is defined by a bedding plane dipping 17°/SSE – the slide surface – and a steep joint-controlled side wall trending 10°NNE. The intriguing aspect of the collapse is that the rock mass adjacent to the scar almost came down as well, but only somewhat broken up and internally rotated, maintained its position along the valley wall ("Jumbled Castle"). Many of the house-size blocks in the Jumbled Castle were rotated backward into the opening gap behind the rock mass. Undisturbed bedding planes adjacent to the Jumbled Castle dip on the average 12°ESSE. During the failure of the cliff face the difference between 12° and 17° in the dip towards the valley might have been the parameter separating a stable (or 'just' stable) from the unstable rock mass.

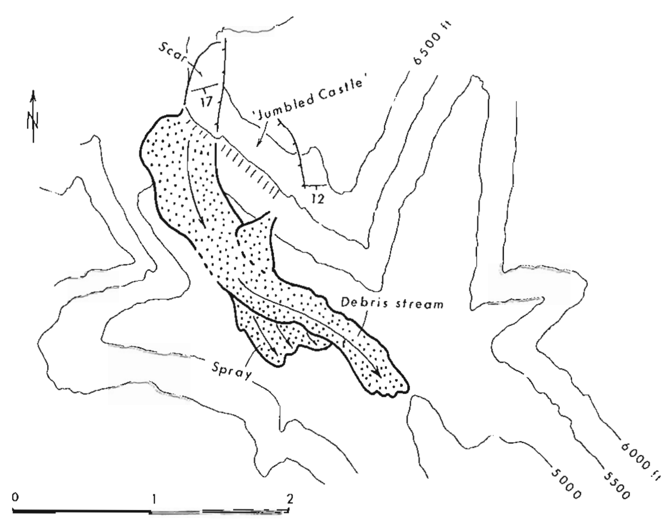

Figure 47.2. Sketch map illustrating the principal features of the Damocles Slide. The arrows indicate direction of movement in the debris stream. Virtually all of the material in the debris stream was derived from the scar.

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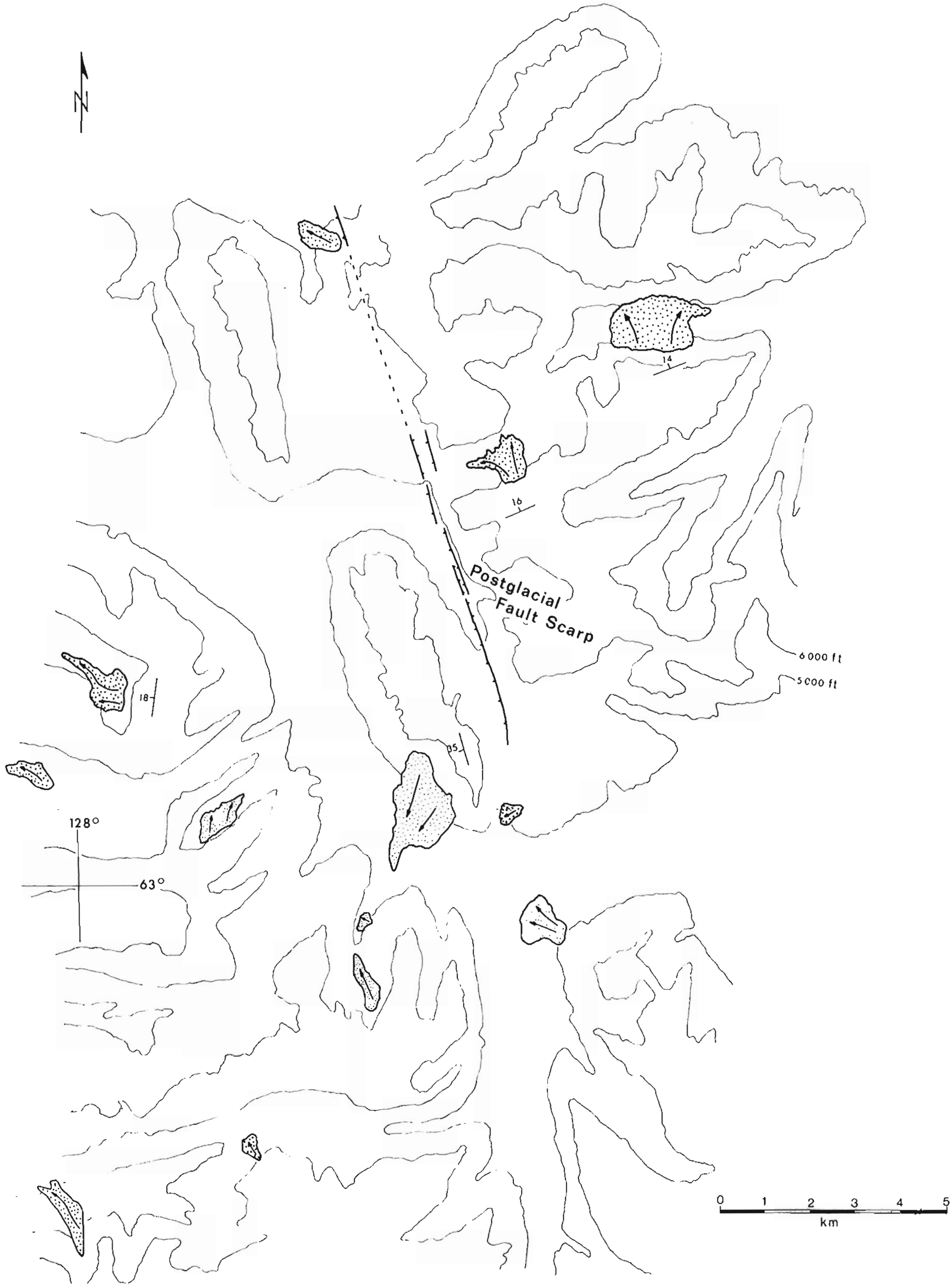

Figure 47. 3. Sketch map of Split Valley showing the location of postglacial fault scarps and distribution of rock slides. The lightly stippled slides are older than the heavily stippled ones. All the slides took off on bedding planes. Barbs on the fault scarps indicate sense of surface displacement.

Only a small subsidiary debris lobe issued from the Jumbled Castle during the failure of the cliff to the west. However, at present the 200-m-high cliff seems to hang on the proverbial hair and is bound to collapse in the future. About 20 x 10^6 m^3 of material collapsed at the scar and fell another 50 – 100 m before it began to stream down the valley. This phase of the slide, the "Sturzstrom" of Heim (1932) and "debris stream" of Hsu (1975) must have occurred in two pulses or waves: first a thin debris spray lapped up and over a low morainal ridge (Fig. 47. 2 and 47. 5), followed by the enormous debris stream which splashed down the valley changing its direction twice (Fig. 47. 2). The border of the debris stream is generally sharp (Fig. 47. 4) although a few blocks up to 2 m across were hurled as much as 15 m outside the main stream. Near its front the stream contains many blocks of 200 – 300 m^3, and one block of 800 m^3. The angle measured from the front of the Damocles Slide to the head of the scarp is 11°. The slide clearly belongs to the category of debris streams with excessive travel distance (Hsu, 1975).

The spray or first pulse of the slide might be interpreted as a frontal portion of the collapsing cliff which was accelerated by momentum transfer from the main disintegrating mass behind. From a close inspection of the 'almost failed' Jumbled Castle of the Damocles Slide it is evident that during collapse large rock units can maintain their integrity until they hit the bottom of the cliff. It is postulated that kinetic energy might have been transferred from large blocks in the back to smaller masses in the front thus propelling the fontal portion forward in an almost independent wave seconds ahead of the main debris stream. Some of the excessive travel distance may also be due to momentum transfer in the early stages of collapse.

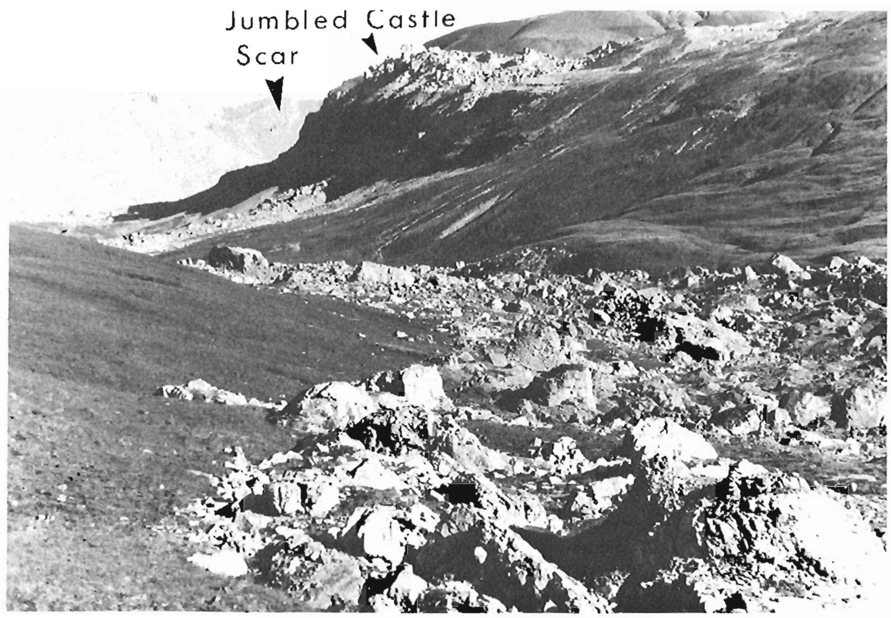

Figure 47. 4.

View looking southward across the Damocles debris stream. Note the sharp outer rim of the thin spray of debris which preceded the main debris stream.

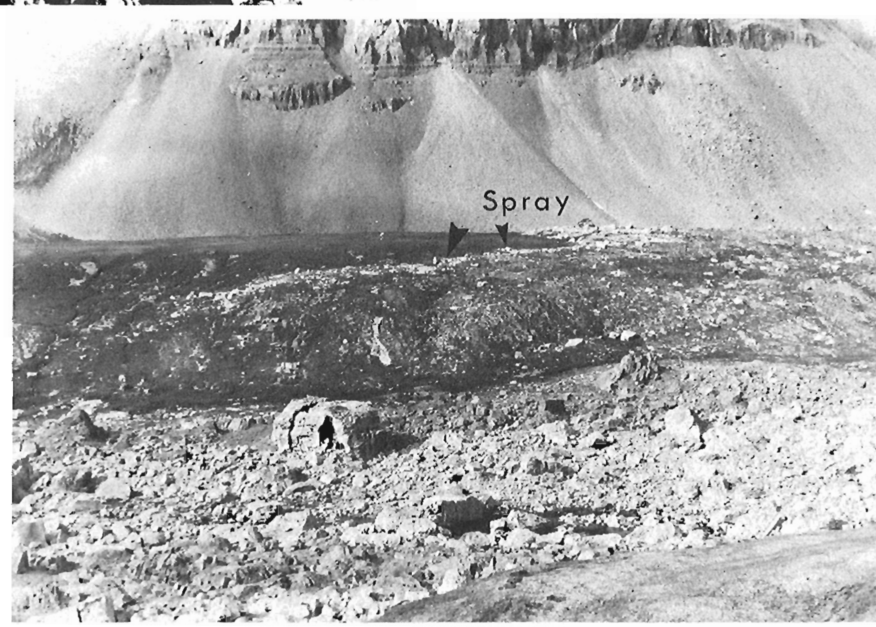

Figure 47. 5.

View of the Damocles Slide. The Jumble Castle in the background consists of white dolomite and limestone broken up into large blocks which give an impression of the situation that must have prevailed at the cliff beyond just before it collapsed.

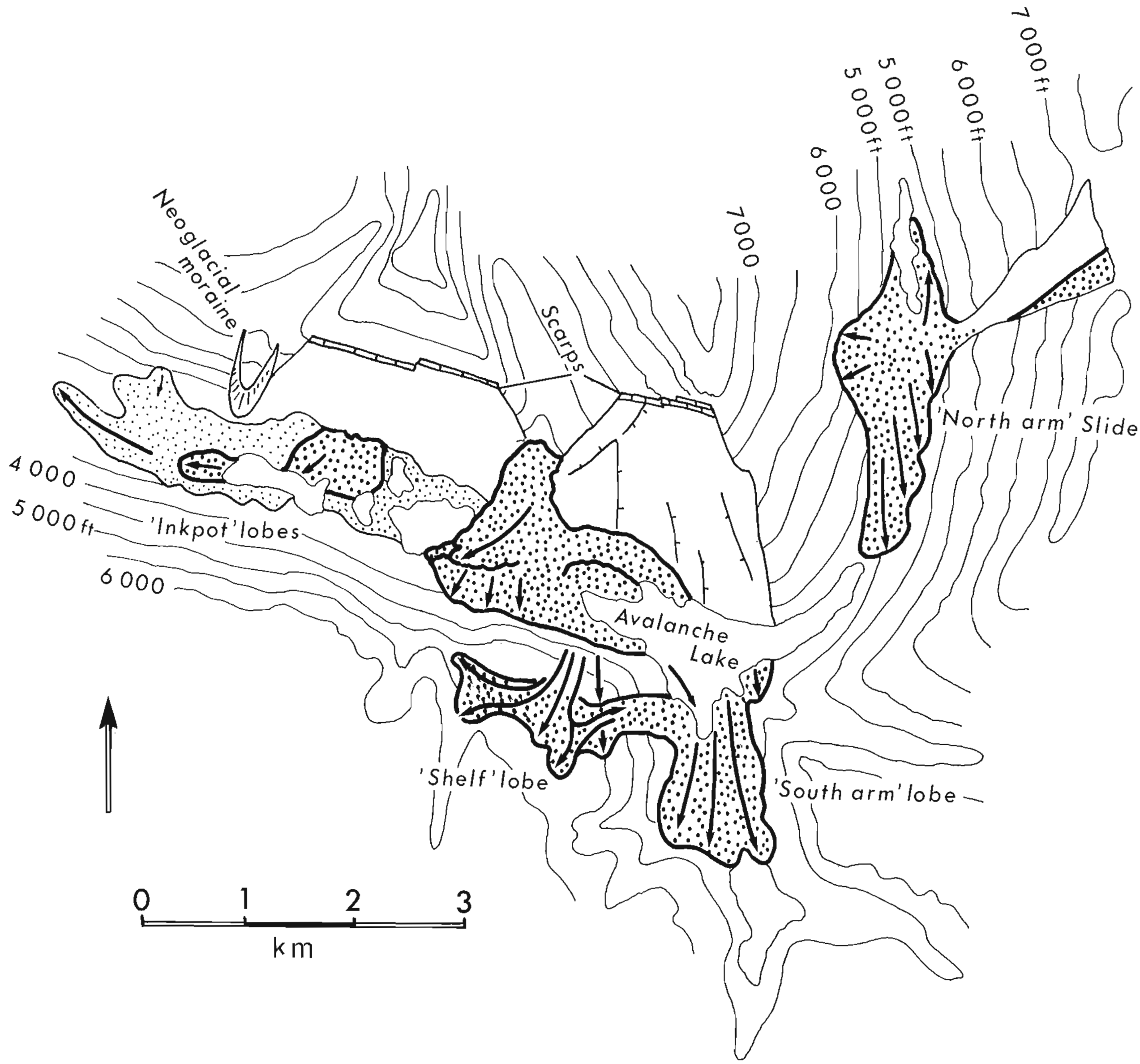

Figure 47.6. Sketch map of Avalanche Lake area. Older slide material ('Inkpot' lobes) is stippled lightly, the main young slide material is stippled heavily. Arrows indicate direction of motion. Note the change in the direction of streaming after the debris stream reached the Shelf south of Avalanche Lake.

B) Split Valley (127°50', 63°05')

The Split Valley area is situated in the Backbone Ranges of the Mackenzie Mountains and is chosen as an example to illustrate the possible relationship between active faulting, earthquakes, and rock slide hazard. The area is underlain by lower Paleozoic carbonates and quartzites which are cut by the Natla Fault, a high-angle east-directed reverse fault (Gabrielse *et al.*, 1973). The Split Valley area contains the trace of a recently active fault (Fig. 47.3 and 47.7). The trace cuts the Holocene surficial material and follows the Natla Fault. In contrast to older eastward directed reverse displacement of the Laramide (?) Natla Fault, the Recent fault displays predominant normal displacement west-down with offsets ranging between 0.5 m to 5 m. Grabens and sag-ponds define the remainder of the fault. Regionally, closely related, are rock slides of two ages: an older group of post-glacial rock slides with a well established cover of vegetation, and a younger group with angular, unvegetated blocks. All of the slides took off on bedding surfaces inclined towards glacially oversteepened valleys. Judging from the concentration of the six younger slides around the fault scarp it is possible that faulting triggered the younger set of slides. The existence of older slides indicates the repetition of similar types of slope failure and the continued slide hazard in this area. This region is clearly located within a belt of intraplate earthquakes (e.g. Horner *et al.* 1974).

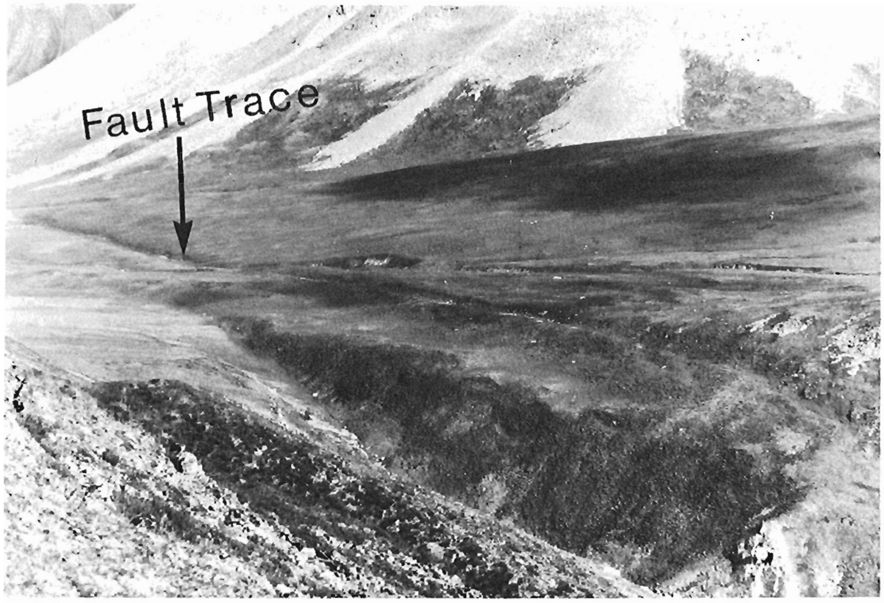

Figure 47.7.

Northeastward view of the postglacial fault trace near its southern end. Note the very fresh rock fall talus in the background.

Figure 47.8.

View looking up the surface of detachment for the principal debris stream at Avalanche Lake.

C) Avalanche Lake (127°15', 62°15')

The area around Avalanche Lake hosts the most spectacular rock slide deposits in the Mackenzie Mountains. Vegetation, and Neoglacial morains suggest several episodes of sliding. In the present context only the youngest phase will be discussed. It is probably of post-Neoglacial age. During this phase a bedding-parallel slope in thick bedded Devonian carbonates collapsed along the north side of the valley and slid along a bedding plane dipping 31°/S (Fig. 47.6 and 47.8). As the material crashed onto the valley bottom it broke up into two debris streams: one moved south-southeastward forming the 'South arm' lobe, the other charged southward up a rock slope of about 45° to an elevation 600 – 700 m above the valley floor and continued to break up into debris streams which followed the pre-existing topographic lows on the shelf. (Fig. 47.9 and 47.10). From energy considerations along it can be calculated that the velocity of the material as it rushed across the valley floor must have been at least 110 m/s before surging up the shelf. From the fact that at least 10 per cent of the rock fragments on the shelf were derived from the cliff below, considerable frictional energy must have been expanded in the gouging of loose talus and solid rock during the ascent of the shelf. The velocity quoted above is therefore an absolute minimum. The deposits on the shelf yield fascinating insights into the mechanism of debris streaming. Part of the debris on the shelf clearly arrived in the form of individual blocks hurled through the air after surging over the rim. The bulk of the material, however, continued to stream at angles up to 90° divergent from the initial direction of movement up

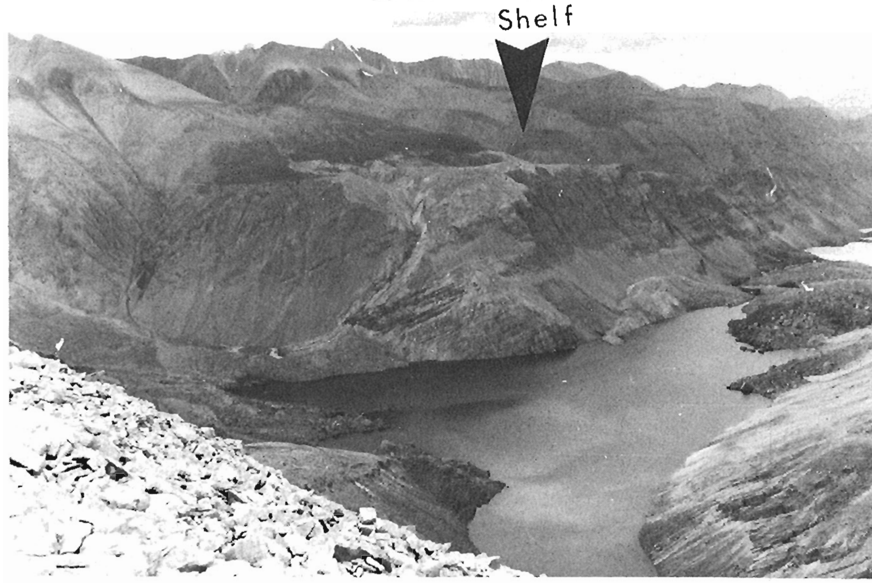

Figure 47. 9.

Southwestward view of Avalanche Lake and the debris – covered Shelf which is about 600 m above the present lake level.

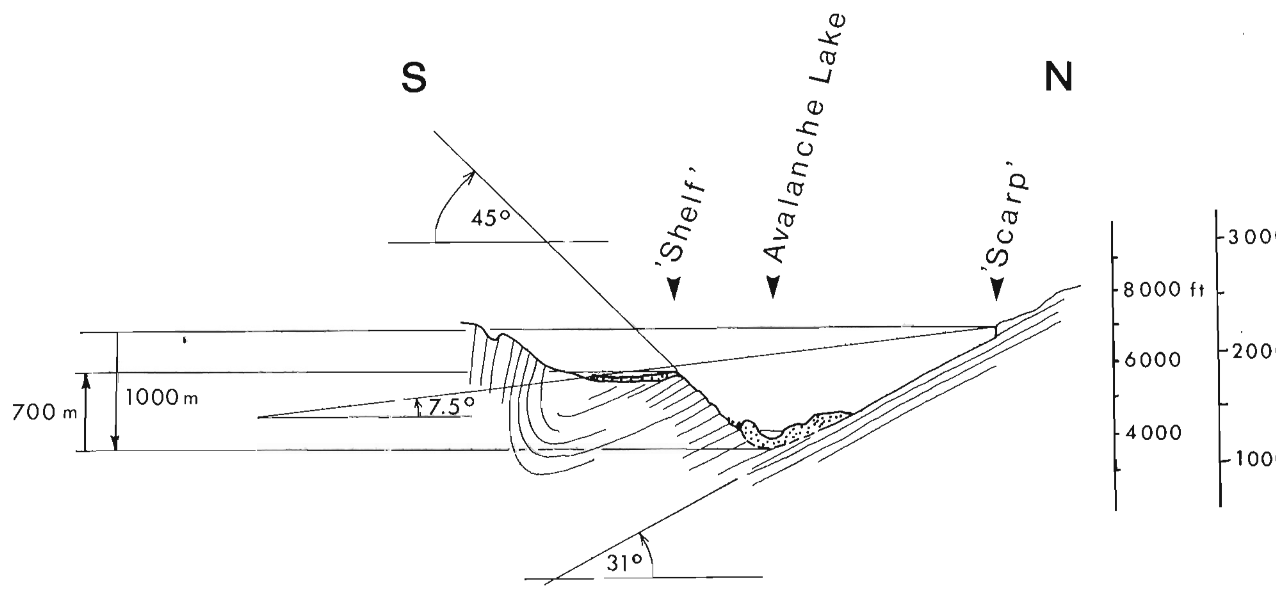

Figure 47. 10. Cross-section through the Avalanche Lake slide near the western end of Avalanche Lake.

the cliff (Fig. 47.6). Blocks up to 400 m^3 participated in the streaming! Along the east rim of the shelf evidence exists that the debris stream was made up of two waves: one slightly preceding the other (Fig. 47.11). While the first wave surged southward across a natural gully in the slope the second wave broke within the gully and continued to flow both up and down the gully. All these features indicate that this debris stream did not move on an 'air cushion' in the sense postulated by Shreve (1968).

It cannot be ruled out that Avalanche Lake existed prior to the slide and water participated as a significant lubricant in the debris stream. But if Avalanche Lake did exist, it was smaller than today. The total volume of the collapsed face is difficult to appraise because of the uncertainty of the pre-slide topography below Avalanche Lake, but a conservative guess would be about 520 x 10^6 m^3.

Conclusions

1) The major rock slides in the Mackenzie Mountains occur in massive carbonate rocks.

2) The principal detachment surfaces are bedding planes which dip 10° to 35° towards the adjacent valleys. The significance of regional bedding plane trends for the direction of sliding has been recognized in other parts of the Cordillera (Eisbacher, 1971; Cruden, 1976).

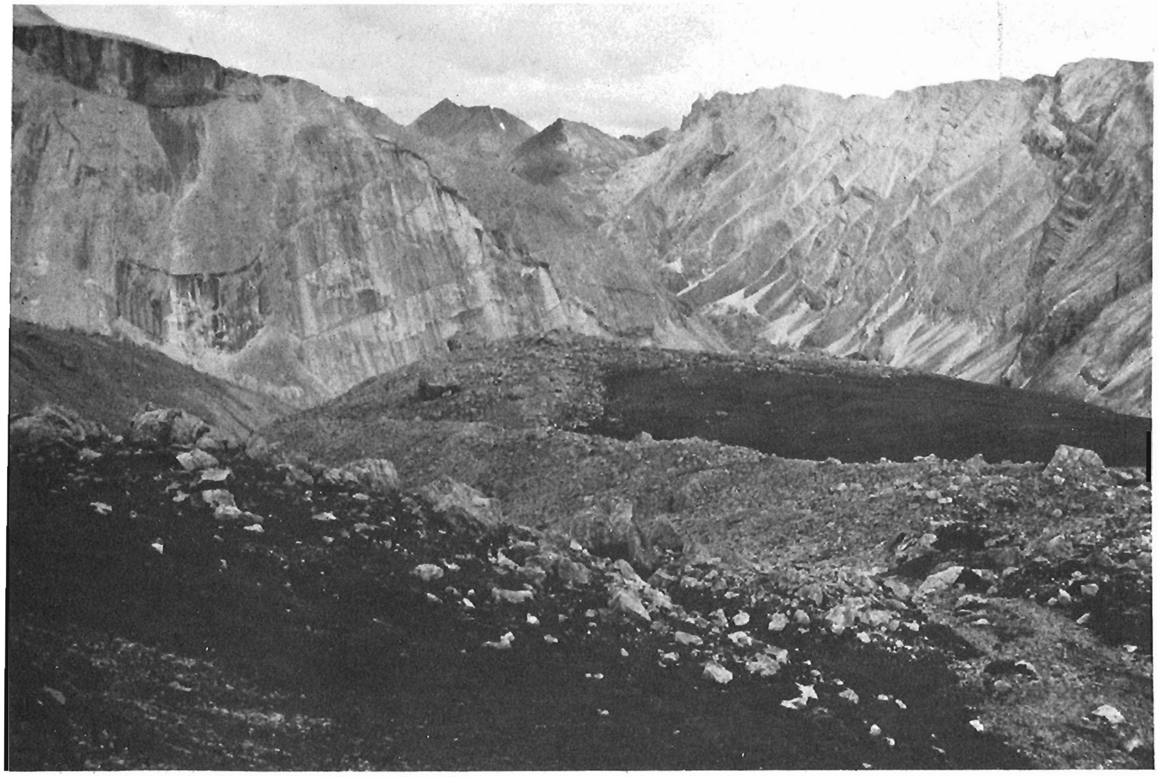

Figure 47.11. The overlapping of two debris waves along the east edge of the Shelf (*see* Fig. 47.6). The earlier wave crossed the gully, the later one followed it.

3) Measured angles between the front of the debris streams and the head scarp can be as low as 7.5° and are commonly less than 20°.

4) Earthquakes were significant trigger mechanisms for some of the postglacial slides in the Mackenzie Mountains.

5) In several localities more than one generation of slide material can be recognized.

6) Valleys adjacent to glacially undercut bedding plane slopes should be investigated carefully prior to location of permanent structures.

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Project 700047

H. Gabrielse, C. J. Dodds and J. L. Mansy[1]
Regional and Economic Geology Division, Vancouver

During the 1976 field season, geological reconnaissance was completed in Toodoggone River (94E) and Ware (west half, 94F W½) map-areas. In addition, some work was carried out in the southern part of Kechika (94L) map-area in order to obtain stratigraphic and structural information pertinent to the solution of problems in the areas to the south and southeast. A few of the more significant observations are outlined below.

Stratigraphy

Marked changes are evident in thickness and facies of Upper Proterozoic and Lower Paleozoic strata across regional faults in northeastern Toodoggone map-area. In the narrow panel between Pelly and Ridgeway faults (Fig. 48.2) fossiliferous Lower Cambrian carbonate strata, about 65 m thick, include conspicuous members of oolitic limestone and dark grey oncolitic dolomite. Presumed correlative strata northeast of Ridgeway Fault consist entirely of limestone no more than 20 m thick. Also, northeast of the fault the rocks are more complexly deformed and more metamorphosed than those to the southwest.

The age of metamorphic strata (to kyanite grade) in the block between Pelly and Ridgeway faults on the west and Kechika fault on the east, including Sifton Range, has not been fully resolved. An assemblage of amphibolite with distinctive members of carbonate and quartzite in the hanging wall of the Sifton thrust fault closely resembles the Upper Proterozoic assemblage east of Rocky Mountain Trench in Fort Grahame E½ map-area to the southeast (Gabrielse, 1976). On the basis of stratigraphic position and general lithology the structurally underlying rocks are believed to be late Proterozoic to early Ordovician in age.

Sequences of Lower Tertiary(?) conglomerate (Sifton Formation) in the Spinel fault zone are as much as 250 m thick and include boulders of metamorphic and granitic rocks derived from a westerly source in

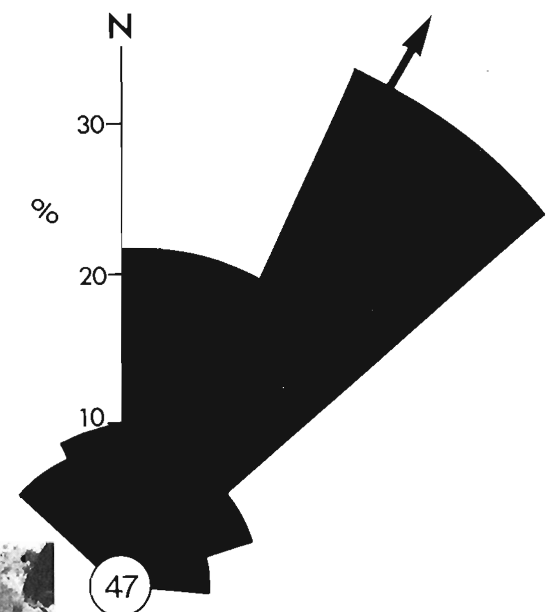

Figure 48. 1a.

Rose diagram showing dominant trends of linear sole markings (current directions are mainly from prod marks as illustrated in Fig. 48. 1b).

Figure 48. 1b.

Linear sole markings in Upper Devonian(?) clastic rocks east of Gataga Lakes.

[1]University of Lille, France

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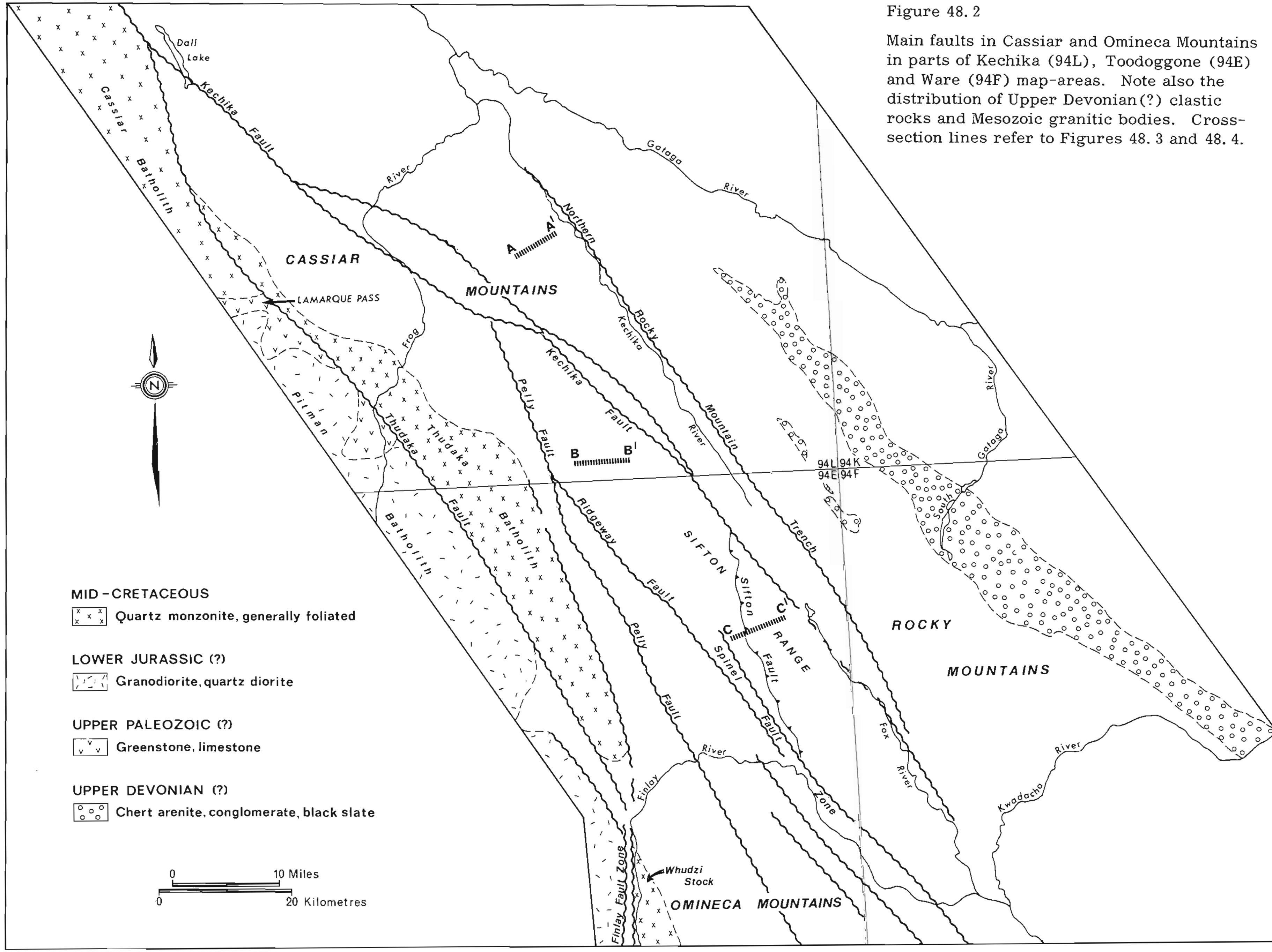

Figure 48.2

Main faults in Cassiar and Omineca Mountains in parts of Kechika (94L), Toodoggone (94E) and Ware (94F) map-areas. Note also the distribution of Upper Devonian(?) clastic rocks and Mesozoic granitic bodies. Cross-section lines refer to Figures 48.3 and 48.4.

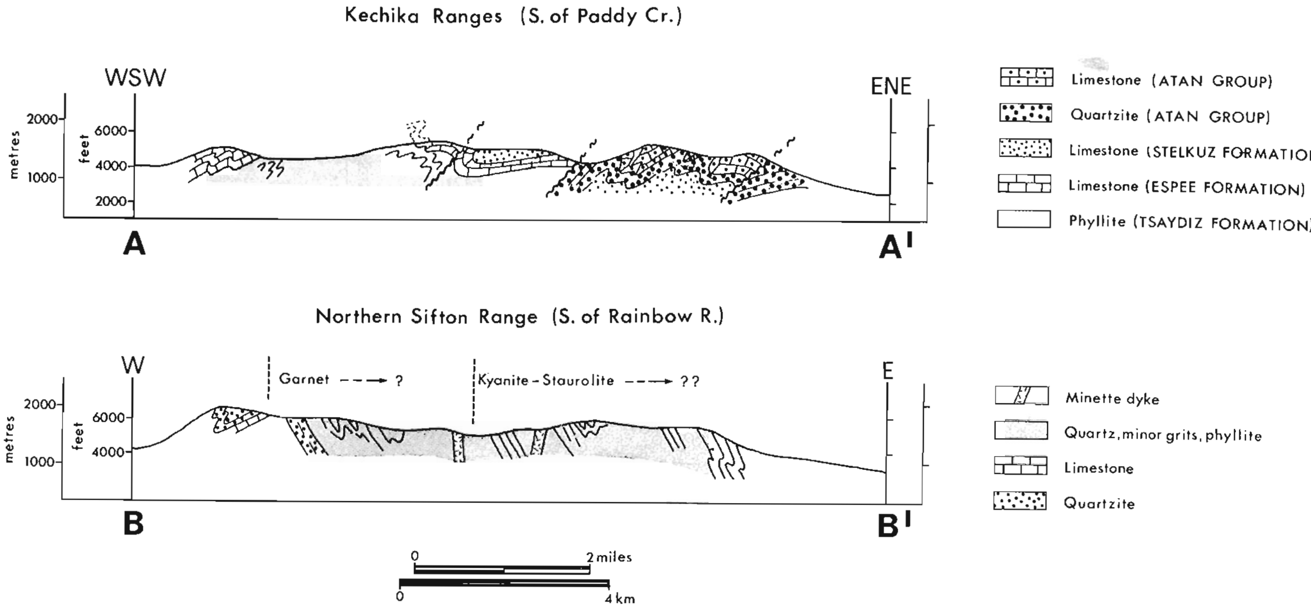

Figure 48. 3. Cross-sections, Kechika and Northern Sifton Ranges.

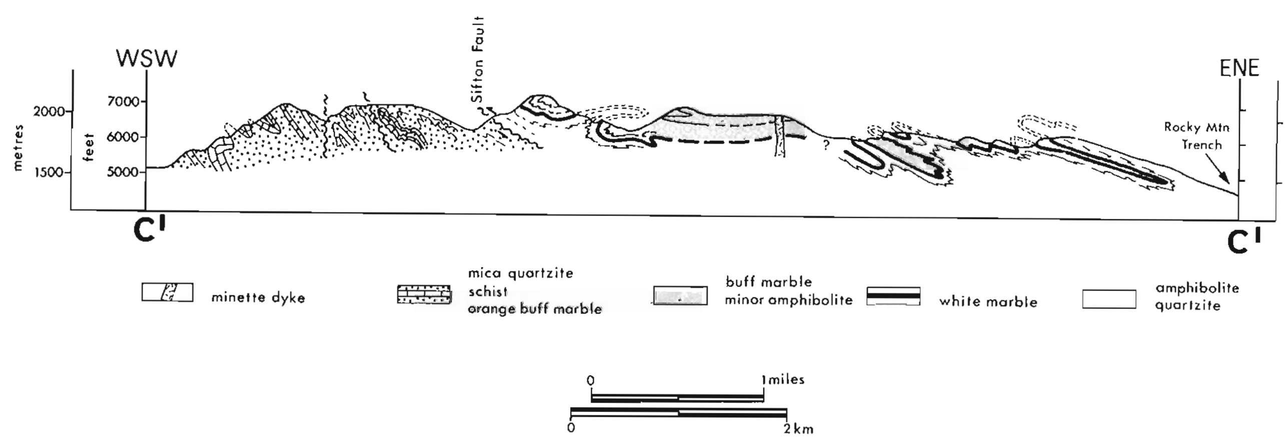

Figure 48. 4. Cross-section, Central Sifton Range (Fox Peak Area).

Swannell and Thudaka ranges. Lamprophyre dykes locally cut the Sifton Formation and are abundant in the region bounded by Kechika, Pelly and Ridgeway faults, Spinel fault zone, and the Rocky Mountain Mountain Trench.

In Ware (west-half) map-area a belt of Upper Devonian(?) noncalcareous shale, sandstone, gritty greywacke, chert-pebble conglomerate and polymictic conglomerate trends northwesterly from the upper reaches of the Kwadacha River to beyond South Gataga Lakes (*see* Fig. 48. 2). These strata extend into the southwesternmost part of Tuchodi Lakes map-area and southeastern Kechika map-area. East of Gataga Lakes the sandy and conglomeratic beds contain spectacular sole markings that consistently indicate northeasterly and northerly sedimentary transport (Figs. 48. 1a, 1b).

No fossils have been found in the clastic rocks discussed above but they overlie Middle Devonian carbonate strata south of Kwadacha River, and a coarse conglomerate believed to be correlative with part of the succession is overlain by Lower Mississippian carbonate west of South Gataga Lakes.

Structure

A number of remarkably continuous faults or fault zones in Cassiar and Omineca mountains trend generally parallel with or more northwesterly than the Rocky Mountain Trench and are marked by zones of cataclasis and by linear valleys. They separate relatively narrow panels that show significant differences in stratigraphy, structural style and degree of metamorphism. Thudaka fault, characterized by a wide, intensely sheared zone in quartz monzonite and intercalated metasediments was traced northerly in Kechika map-area to beyond Lamarque Pass. Southeast of Lamarque Pass the fault separates mid-Cretaceous foliated quartz monzonite of the Thudaka batholith on the northeast from quartz diorite and granodiorite of the Pitman batholith and metavolcanic and sedimentary rocks of late Paleozoic(?) age to the southwest. For about 15 km north of Lamarque Pass foliated quartz monzonite occurs on both sides of the fault zone and still farther north quartz monzonite of the Cassiar batholith lies west of the fault and abuts little metamorphosed Upper Proterozoic strata to the east. Presumably Thudaka fault continues northerly along the east side of Cassiar batholith and intersects Kechika fault west of Dall Lake. If the Thudaka batholith is an offset portion of the Cassiar batholith then Thudaka fault has an apparent dextral displacement in excess of 20 km.

At its southern end Thudaka fault swings into or intersects the Finlay fault zone which separates Upper Proterozoic rocks and quartz monzonite to the east from Mesozoic volcanic and granitic rocks to the west. Dextral offset in the order of 15 km is suggested by the distribution of Thudaka batholith and the lithologically and structurally similar Whudizi stock to the south.

In southern Kechika map-area Pelly fault bounds easterly to southeasterly trending low-grade Upper Proterozoic rocks to the west and southerly to southeasterly trending, relatively high grade metamorphic rocks, locally pegmatitic, to the east.

Kechika fault zone contains highly cataclastized rocks from Frog River to the Rocky Mountain Trench. It marks the boundary between northeasterly directed structures to the northeast and southwesterly directed structures to the southwest (*see* Fig. 48.3).

Sifton thrust fault is a northerly trending, east dipping structure truncated just north of Finlay River by the Spinel fault zone (*see* Fig. 48.4). Intensely folded amphibolite, carbonate and quartzite in the hanging wall underlie a long, narrow strip south of Sifton Range between Spinel fault zone and the Rocky Mountain Trench.

Economic Geology

A minor occurrence of chalcopyrite-pyrite (with associated azurite-malachite staining), was noted in complexly deformed quartz-calcite veins in chlorite grade calcareous argillite and siltstone of the Misinchinka Group on the east side of the Rocky Mountain Trench about 5 km, west of Chisholm Peak.

Many large limonitic gossan deposits occur in areas underlain by Upper Devonian (?) clastic rocks in northern Rocky Mountains. One such deposit in southwesternmost Tuchodi Lakes map-area has been explored by trenches that reveal bedded barite, pyrite, sphalerite and galena in the underlying bedrock. The minerals occur in black, or, locally, green noncalcareous pyritic slate.

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Project 720038

T. A. Richards
Regional and Economic Geology Division, Vancouver

The summer of 1976 was spent mapping the Hazelton map-area, west half (93M-W½) and comments offered here are restricted to data and ideas arising from the season's work. For reports of earlier work *see* Richards and Jeletzky, 1974 and Tipper and Richards, in press.

Most of the area is underlain by fluvio-deltaic and shallow marine clastics of the Middle to Upper Jurassic Bowser Lake Group (Richards and Jeletzky, 1974; Jeletzky, 1976). Sedimentary and volcanic rocks of the Skeena and Sustut groups (or correlative units), volcanic rocks of the Brian Boru Formation, and the 50-80 m.y. old Bulkley Intrusives underlie smaller areas.

Much time was spent differentiating between the Skeena Group and the Bowser Lake Group. It was found that the transition between the two units involved a change in sedimentation and a probable pre-Hauterivian hiatus. Occupying the interval between the youngest known Bowser Lake sediments and the oldest known Skeena sediments, is a suite of coal-bearing fluviatile clastics with genetic affinities to both groups. A similar assemblage, the Jenkins Creek assemblage, has been suggested for the McConnell Creek map-area (Eisbacher, 1974). The Bowser Lake Group has been interpreted as a prograding fluvio-deltaic suite (Richards and Jeletzky, 1974; Jeletzky, 1976). The "intermediate" facies represents further progradation northward, appears to be continuous with the Bowser Lake Group, and therefore has been tentatively included with it. The Skeena Group is lithologically similar to the underlying, "intermediate" facies but is distinguishable by the presence of locally abundant muscovite. The base of the Skeena may be defined as a black marine shale or a coarse cobble conglomerate containing detrital muscovite flakes and pebbles of milky vein quartz, chert, and volcanics. This marks the first appearance of muscovite and quartz detritus in any abundance in sedimentary rocks younger than Early Oxfordian. Its age is inferred to be Hauterivian? based on its lithologic and stratigraphic similarities to black clastics in the lower Skeena Valley near Skeena Crossing.

The major difficulty in distinguishing the Bowser Lake and Skeena groups comes from the possibility that deposition continued without major interruption in Hazelton map-area but as most rocks are in a continental facies, no reliable guide fossils are available. Floras of this time interval have not proven useful.

The Brian Boru Formation (Sutherland Brown, 1960) in the southern and extreme western part of the Rocher Deboule Range, consists of over 4000 feet of acidic pyroclastics centred around Brian Boru Peak. They appear similar to rocks mapped as Ootsa Lake Group to the south, and are tentatively correlated with them. They are of probable rhyolite to rhyodacite composition but may include minor quantities of andesite and basalt. They comprise an assortment of breccia, lapilli tuff, air-fall tuff, ash-flow tuff and sedimentary rocks derived from the pyroclastics, some of which are spectacular conglomerates and fanglomerates that are particularly well exposed on Highway 16, immediately north of Beament Station. In places these rocks are intensely pyritized producing an impressive gossan. Along the spur leading south from Brian Boru Peak, the volcanics are nearly flat lying and rest with angular discordance on augite porphyry volcanics of the Lower Cretaceous Rocky Ridge volcanics. This discordance is probably an unconformity of local extent. Brian Boru volcanics around Brian Boru Peak could possibly be considered the volcanic ejecta of the Roche Deboule stock (Sutherland Brown, 1960), dated at 70 m.y.

The structure of the area is dominated by block faults. Most major river valleys represent grabens, in particular the southern part of the Skeena, Bulkey, Kispiox, Nichyeskwa, Shelagyote and upper Babine rivers. The pattern of grabens is not simple but may be best described as large polygonal blocks with a general northerly elongation. In general, relatively younger rocks outcrop in the valleys and older rocks on the adjacent mountain masses. Youngest rocks known to be involved in the block faulting in the map-area are of Paleocene-Eocene age. Folds trend northwest and are usually asymmetrically overturned to the northeast. Along the southern margin of the map-area an east-west trend persists. Structures around the Bulkley Intrusions are controlled by these bodies and do not conform to the regional patterns.

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ZEOLITIZATION OF THE TELKWA VOLCANICS, SMITHERS, BRITISH COLUMBIA

Project 720038

J.S. Dudley[1]
Regional and Economic Geology Division, Vancouver

Introduction

An investigation of the secondary mineralization in the Jurassic Telkwa Formation underlying the west half of the Smithers map-area (93L) is being undertaken as a thesis study to investigate the problem of burial metamorphism vs hydrothermal metamorphism. In 1976 only preliminary mapping and identification of the low grade secondary mineral assemblages was carried out.

General Geology

The study area is underlain by the Howson subaerial facies of the Telkwa Formation (Fig. 50.2) a Lower Jurassic formation of the Hazelton Group as defined by Tipper and Richards (in press). It is predominantly basaltic to rhyolitic pyroclastics with subordinate flows and sediments.

Lapilli tuff, ash tuff, accretionary lapilli, welded tuff, lahar, and fine to coarse breccia comprise the pyroclastics. The fine ash tuffs and accretionary lapilli are distinctive and serve as useful markers. In one instance, correlation of sections over six miles apart was possible.

A feature of the flows and pyroclastics is the preservation of primary textures. Basaltic flow top breccias and amygdaloidal zones are easily distinguished, as well as zones of the flow bottom. Many andesite-dacite flows exhibit scoriaceous beds. The volcaniclastics show cross-bedding and primary porosity preserved by mineral filling.

The structural style is regional block faulting with good sections exposed in cliffs. The dip is low to moderate.

As stated by Tipper and Richards (in press), the rocks are extensively altered with a regional development of zeolites, epidote, prehnite, and calcite. This alteration is the core of this study.

Occurrence of Alteration

Within the Howson subaerial facies the low grade secondary minerals are in three major forms:

1. as veins, from approximately 30 cm wide to fine veinlets (<1 cm) that cut the strata.
2. as primary porosity fillings that form amygdules and the cement of breccias.
3. as a matrix component of secondary minerals in pyroclastics and flows.

The veins and porosity fillings are medium to coarse minerals observable with the unaided eye. The occurrences as matrices of rocks are not evident without thin section and X-ray diffraction study.

Minerals

The secondary minerals that have been identified with the help of X-ray diffraction are epidote, albite, adularia, prehnite, barite, chlorite, calcite, celadonite, and numerous zeolites such as mesolite, heulandite, stilbite, thomsonite, laumontite, and analcime. The occurrence of wairakite has also been determined by T.A. Richards (pers. comm.) from previous X-ray work. Some malachite, chalcocite, and native copper was found in some of the coarser veins of zeolite.

Burial or Geothermal Mineral Genesis

Assemblages of zeolites and other low grade secondary minerals have been attributed in the literature to pressure-temperature conditions of burial metamorphism (Fig. 50.2). Many of the established phase relationships are summarized on Seki (1969). The trends proposed from the past studies do not appear to be applicable to the rocks of the Howson facies. The most common vein mineral in the area is laumontite and the veins cut across hundreds of feet of strata commonly containing zeolites that supposedly demark lower temperatures of formation, e.g. stilbite, heulandite. Although not verified by X-ray diffraction, mapping suggests that epidote and prehnite are found in vein assemblages with laumontite. X-ray diffraction has verified the occurrence of pumpellyite and laumontite together. These assemblages juxtapose minerals indicative of prehnite-pumpellyite facies with those of moderate zeolite facies at the same stratigraphic level. This contradiction coupled with Richards' discovery of wairakite, suggests a geothermal origin of the zeolites as opposed to burial metamorphism. Wairakite is characteristic of zeolite facies alteration in a geothermal area (Seki, 1969).

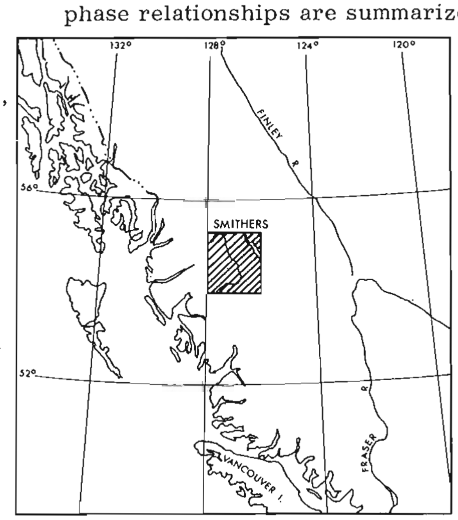

Figure 50.1. Index map.

[1] Department of Geology, University of Calgary, Calgary, Alberta.

From: Report of Activities, Part A;
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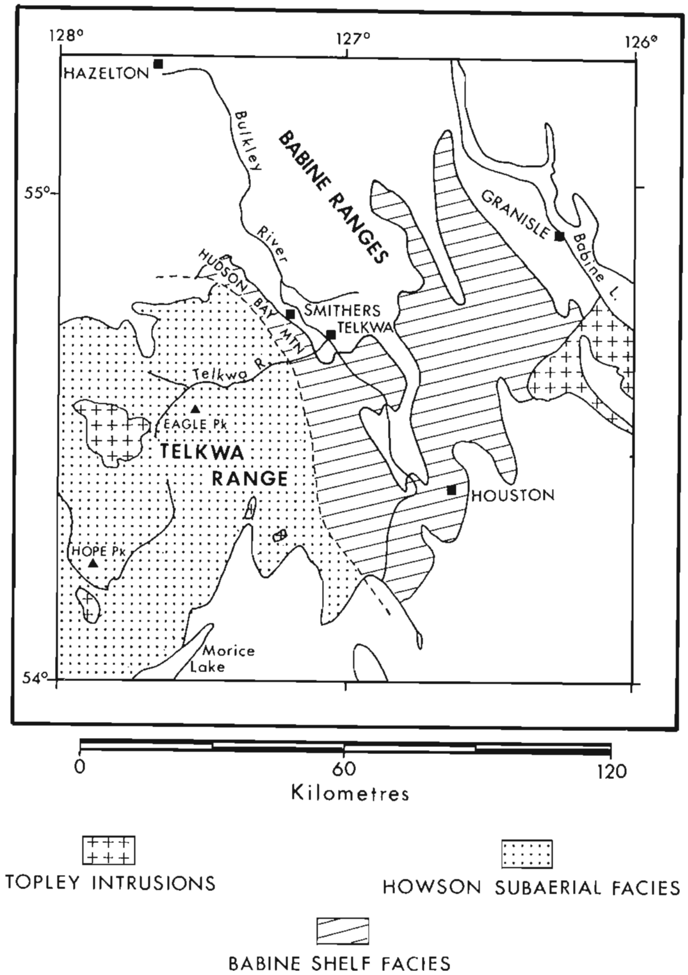

Figure 50. 2. Distribution of the Howson Subaerial Facies Adapted from: Tipper and Richards (in press)

Any substantial burial before development of zeolite cements in the volcaniclastics seems doubtful, as the resultant decrease in porosity would prevent the zeolite from attaining the coarseness observed in these cements. This implies early porosity filling suggesting that the secondary minerals may represent more of a diagenetic process than one of metamorphism.

The zeolitization tends to favour basaltic and andesitic rather than rhyolitic rocks. Barite, calcite, and epidote have been found in the rhyolitic material but no zeolites. A high concentration of CO_2 and other chemical characteristics of the hydrothermal solutions have been used to explain this lack of zeolitization in active geothermal areas (Miyashiro, 1973).

That basalts of the Howson facies flowed into topographically low areas as suggested by Tipper and Richards (in press) is confirmed by the writer's mapping. These basins would be suitable areas for the

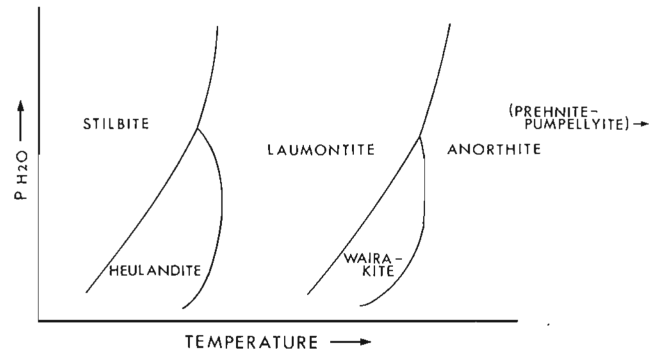

Figure 50. 3. Possible phase boundaries of some zeolite species identified in field area and direction in which higher grade prehnite-pumpellyite facies is found. From: Miyashiro, 1973, p. 162.

circulation cells of hydrothermal solutions that may have precipitated the zeolites. This could also explain the preferential zeolitization of more basic rocks.

Significance of Study

Zeolitization studies in active geothermal fields have had to rely upon the restricted sampling of drill core as in Wairakei, New Zealand (Coombs *et al.*, 1959) and Onikobe, Japan (Seki, 1969). The Telkwa study area may offer an opportunity to examine cross sections of an extinct geothermal field. The strata are well exposed both horizontally and vertically in simple, gently dipping fault block mountains giving exposure unavailable in active fields. Lack of alteration of the primary minerals suggests that the interpretation suggested here may be valid as later alteration does not appear to have been superimposed on the zeolitization.

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JURASSIC STUDIES IN QUEEN CHARLOTTE ISLANDS, HARBLEDOWN ISLAND, AND TASEKO LAKES AREA, BRITISH COLUMBIA

Projects 610011 and 750035

H.W. Tipper
Regional and Economic Geology Division, Vancouver

The Vancouver Basin (Tipper and Richards, in press) was studied during the 1976 field season in Queen Charlotte Islands, on Harbledown Island, near Cowichan Lake on Vancouver Island and along Tyaughton Creek in Taseko Lakes area (Fig. 51.1) to establish the biostratigraphic succession and to define the Lower and Middle Jurassic lithologic subdivisions. In addition further investigations were carried out along Tyaughton Trough (Jeletzky and Tipper, 1968) and the Yalakom Fault system to complete the mapping of Taseko Lakes map-area (NTS 92 O).

Lower and Middle Jurassic

Queen Charlotte Islands

Extensive collections of fossils from beds lithologically similar to Maude and Kunga formations confirm the extension of these formations from the exposures on Maude Island and other islands of Skidegate Inlet (Tipper, 1976) to central Graham Island along Ghost, Phantom, and King creeks (Sutherland Brown, 1968, p. 50-66). There Early (?) and Late Sinemurian and Early and Late Pliensbachian ammonites are abundant and well preserved. Some of these beds were of interest in recent years as possible oil shales, particularly the black fetid shales of the Maude Formation. However in many places the sections comprise interbedded shales and siliceous tuff, the latter making up as much as 50 per cent of the rock by volume.

The Kunga Formation displays prominent bands of white, grey, and reddish weathering siliceous, fine grained tuff interbedded with black to dark siltstone and shale (Fig. 51.2). The tuff is in beds up to 15 cm thick, but commonly is about 2 cm. In places it is mixed with silt to form a tuffaceous siltstone. A few, thin, white or reddish weathering tuffaceous beds are interbedded with the Pliensbachian sediments of the Maude Formation but none were noted in the Toarcian part. These tuffs are thought to be mainly air-fall tuff but no source is suggested Sutherland Brown (1968) did not record volcanics of Sinemurian or Pliensbachian age.

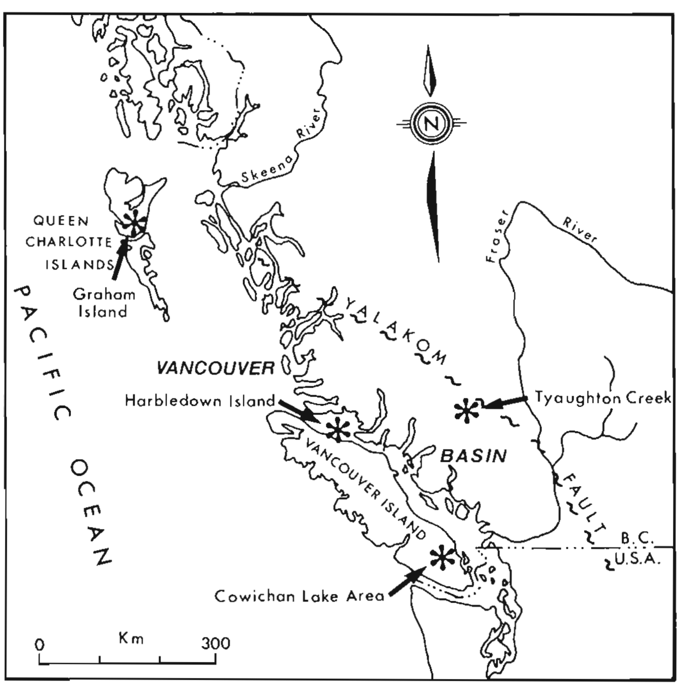

Figure 51.1 Location of areas studied.

Figure 51.2. Interbedded tuff (white to light grey) and siltstone (dark grey to black) of the Kunga Formation, south side of Maude Island, Skidegate Inlet, Queen Charlotte Islands.

From: Report of Activities, Part A;
Geol. Surv. Can., Paper 77-1A (1977)

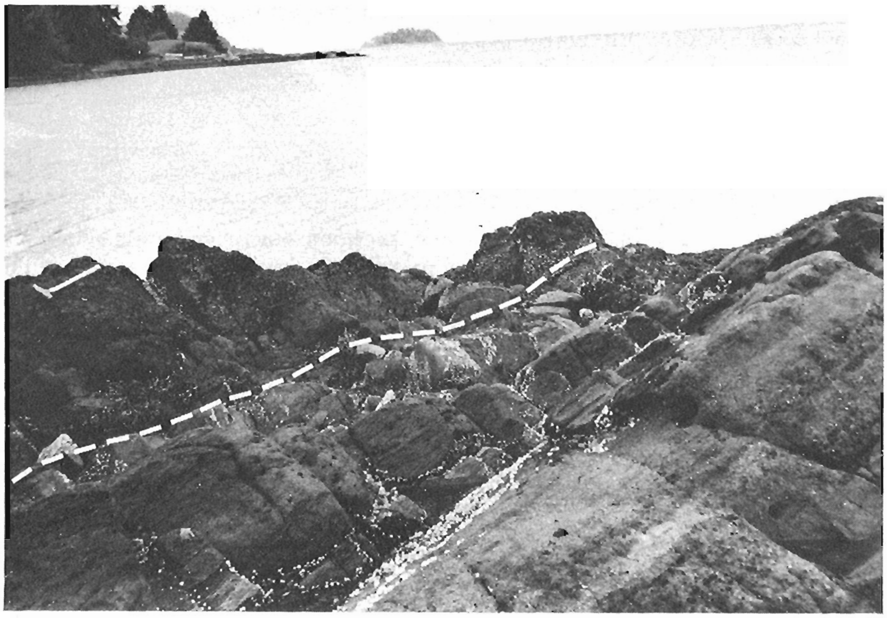

Figure 51. 3

Disconformable contact between the Lower Callovian upper Yakoun sediments and the volcaniclastic Middle Bajocian; near east end of Maude Island, Skidegate Inlet, Queen Charlotte Islands.

The Yakoun Formation comprises volcanic and sedimentary rocks of Middle Jurassic age (Sutherland Brown, 1968, p. 66-76). The lower part is mainly volcanic and is Middle Bajocian in age. The upper part is entirely sedimentary, except for one bed of lapilli tuff, above a prominent pebble conglomerate. Lower Callovian ammonites were found in the sediments near Robber Point, Maude Island. No Bathonian or Upper Bajocian fossils were found hence the Lower Callovian upper Yakoun rests disconformably on Middle Bajocian lower Yakoun volcanics (Fig. 51. 3).

Harbledown Island

Crickmay (1928) studied the geology of Harbledown Island and introduced the name Harbledown Formation for a sequence of Sinemurian shales and argillites. In 1976 these Lower Jurassic sediments were restudied and an extensive fossil fauna collected.

The Harbledown Formation in the type area (Fig. 51. 4) comprises pelitic sediments with a few sandy interbeds. No good section exists but the lower beds with *Arniotites* are interbedded black to grey argillite and light grey to grey tuff in beds a few centimetres thick or less. In places the tuffs comprise as much as 50 per cent of the rock but decrease in thickness and abundance upward. The highest beds with *Melanhippites* are black to dark grey shale and argillite without discernible tuff beds. In the lower argillite beds, grey limy concretions to 15 cm are scattered irregularly. The sediments have been altered to hornfels and baked by Tertiary (?) intrusions that underlie most of the island. The sediments remain as small remnants on the margins of the plutons.

The area is cut by a myriad of near vertical faults. The formation has a general northeasterly shallow dip but the fracturing is so pervasive that a hypothetical section can only be obtained by stacking many short sections controlled by the contained fauna. The older Upper Triassic Parson Bay Formation is separated from the Harbledown Formation by a fault; the contact is not gradational as previously suggested (Jeletzky, 1970, p. 9).

The biostratigraphic succession proposed by Crickmay is essentially correct. However a more varied fauna was collected and the presence of four or possibly five faunal zones of the standard European zonation are documented. The fauna collected were submitted to Dr. Hans Frebold and the results of his study are summarized in Table 51. 1. This succession is confirmed by field work in Queen Charlotte Islands and Tyaughton Creek.

Cowichan Lake Area

A brief visit was made to the Cowichan Lake area, Vancouver Island with J. E. Muller to collect fossils from sedimentary members in the Bonanza Volcanics. At one locality a varied fauna from tuffaceous greywacke and siltstone indicated the probable presence of the Upper Pliensbachian Margaritatus zone. Another locality with a poorly preserved fauna suggested a Sinemurian age.

Tyaughton Creek Area

Intensely thrust faulted Lower and Middle Jurassic sediments occur at the headwaters of the south branch of Relay Creek. No section is well exposed but by piecing together the faunal zones from different thrust sheets a reasonable section can be assembled. The Lower Sinemurian to Middle Bajocian sediments are composed mainly of black to grey shale and siltstone with minor grey to black limestone bands or lenses and minor fine greywacke or coarse siltstone. Conglomerate is unknown in the section. Grey to light grey concretions to 1. 5 m diameter, usually 10 cm or less, are common.

A varied but sparse fauna was collected and preliminary field identifications suggest that nearly all Lower Sinemurian to Middle Bajocian zones known elsewhere in British Columbia are represented in the Tyaughton Creek area. This, coupled with previously obtained evidence, confirms that the Tyaughton Creek rocks of Late Triassic (Late Norian) to Middle Jurassic (Middle Bajocian) age are entirely sedimentary, are essentially a conformable succession and are probably representative of all stages and substages of this interval.

The Middle Bajocian shales are overlain with erosional (possibly angular) unconformity by Lower Callovian coarse sandstone, pebble conglomerate and tuffaceous sediments. At no place in this area are Upper Bajocian or Bathonian strata found.

Yalakom Fault and Tyaughton Trough

The Tyaughton Trough was a basin of deposition during Late Jurassic and Early Cretaceous time in which 3000 to 5000 m of sediments accumulated (Jeletzky and Tipper, 1968). The northeast margin of the basin was delineated throughout most of its history by the Yalakom Fault, a northwest trending, right-lateral transcurrent fault. It is readily apparent that older rocks are offset by a greater amount than younger strata but the precise amount of offset of any sequence is still difficult to determine precisely.

Tectonic activity persisted in the Tyaughton Trough during sedimentation. At least four unconformities have been recognized. The oldest, possibly older than the Trough, is pre-Early Callovian in which coarse

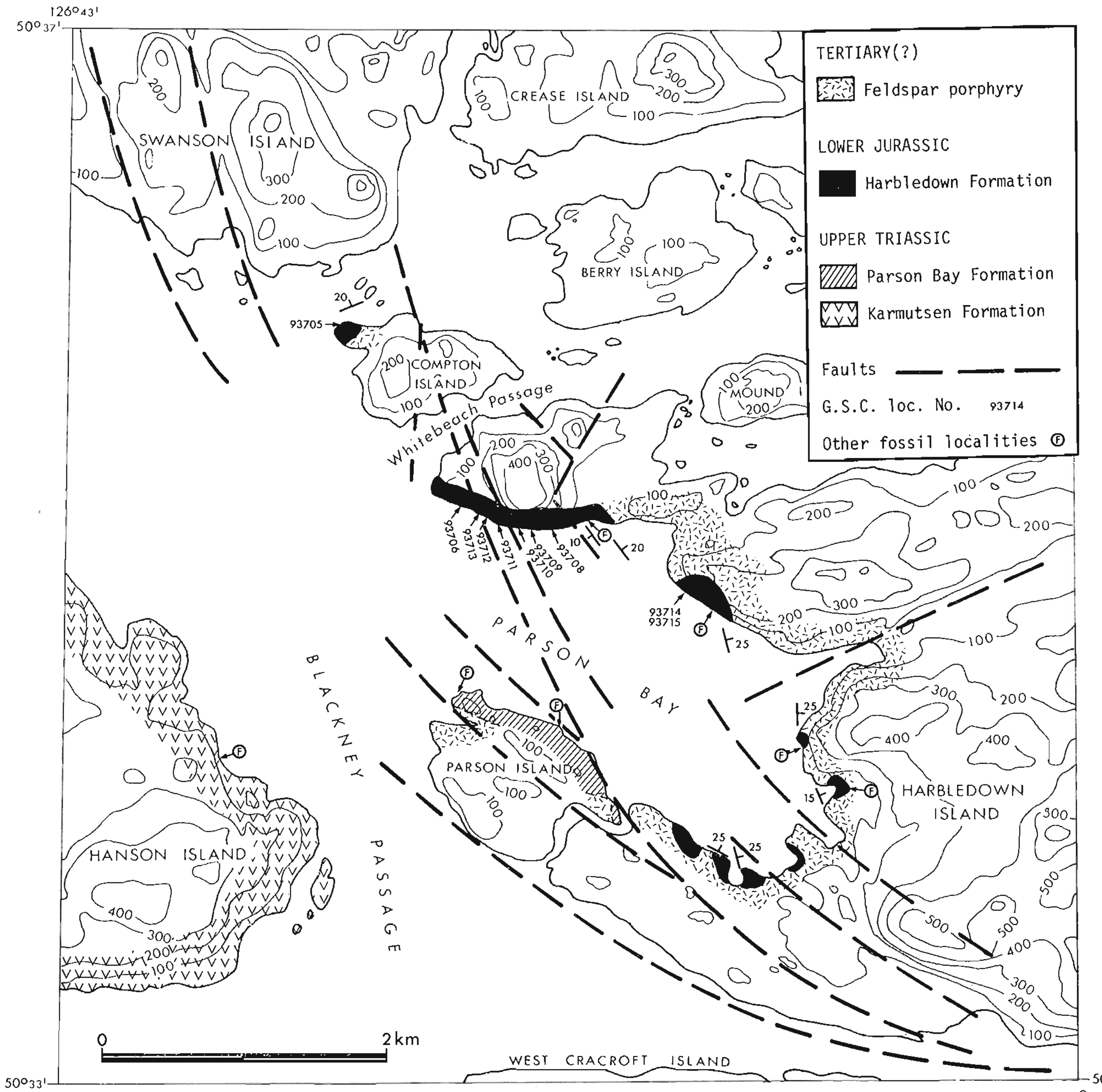

Figure 51.4. Geology of Harbledown Island.

SUB-STAGES	ZONES	AMMONITES	GSC LOCS.
Upper Sinemurian	Raricostatum Zone	*Melanhippites* (= *Paltechioceras*) *harbledownensis* *Melanhippites* (= *Paltechioceras*) sp. nov.?	93708 93714 93709 93715
	Oxynotum Zone	*Oxynoticeras*? *Gleviceras*? *Crucilobiceras* sp. indet. *Melanhippites* (= *Paltechioceras* sp. nov.?)	93710
	Obtusum Zone	*Asteroceras* cf. *A. stellare* *Asteroceras* sp. *Arniotites* sp. *Arniotites kwakiutlanus*	93705 93706
Lower Sinemurian	Turneri Zone Upper part of Semicostatum Zone	*Arniotites* cf. *A. kwakiutlanus* *Arniotites begbiei* Ammonites gen et sp. indet.	93711 93712 93713
	Lower part of Semicostatum Zone Bucklandi Zone	Not identified	

Table 51.1. Biostratigraphy of the Harbledown Formation.

Callovian (late Middle Jurassic) sediments rest on Middle Bajocian (early Middle Jurassic) sediments and on Upper Triassic (?) augite porphyry breccias. Mid-Hauterivian (mid-Early Cretaceous) sediments rest directly on Callovian sediments and on Valanginian (early Cretaceous) sediments. Barremian (Early Cretaceous) sediments were observed at one locality resting with angular (about 15 degrees) discordance on Hauterivian sediments. Albian and Aptian sediments of the Taylor Creek Group rest on Middle Triassic (?), on Upper Triassic, on Middle Jurassic, on Upper Jurassic, and on earlier Cretaceous strata over a wide area and include, in conglomerates, fossiliferous pebbles of earlier Trough sediments. These unconformities are recognized near the margins of the Trough but in the axial parts the succession may be nearly conformable throughout. The structure is further complicated by numerous felsite and feldspar porphyry intrusions of Tertiary age and by several northeasterly directed high-angle reverse faults related to the emplacement of Eocene (?) felsitic plutons near Mount Sheba and Mount Warner and in the Coast Mountains to the southwest.

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From: Report of Activities, Part A;
Geol. Surv. Can., Paper 77-1A (1977)

Project 720041

J.W.H. Monger
Regional and Economic Geology Division, Vancouver

This project is designed to determine the stratigraphy, facies, structural style and external relationships of upper Paleozoic rocks in the western Cordillera. During the 1976 field season Monger re-examined previously-mapped, critical areas of the upper Paleozoic Cache Creek Group near Atlin, British Columbia (Monger, 1975), and also studied small inliers of Permian and older strata around the Mesozoic Bowser Basin (Fig. 52.1). Terry (this pub., rep. 53) continued studies on the Nahlin ultramafic body, thought by Monger (1975, p. 31) to represent part of the basement on which the Cache Creek Group was deposited. Werner (this pub., rep. 54) examined probable pre-Upper Triassic metamorphic rocks southwest of Atlin Lake to determine the original nature and age of the metamorphosed strata and time(s) of metamorphism.

Cache Creek Group near Atlin, British Columbia

The Sentinel Mountain, Nakina Lake and Mount Farnsworth areas of the Atlin Terrane were re-studied. Particular attention was paid to relationships between carbonate and chert in an attempt to resolve the apparent paradox, so common in the Cache Creek Group, of shallow water carbonate bodies enclosed by or overlying supposedly deep water, thin bedded, radiolarian chert.

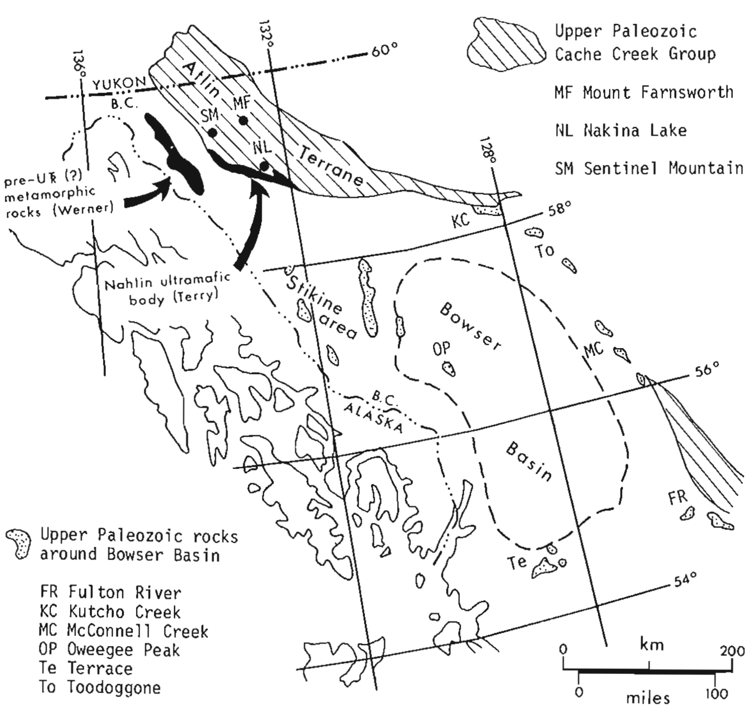

Figure 52.1. Upper Paleozoic rocks, Northwestern British Columbia.

(a) Sentinel Mountain area (SM)

The Sentinel Mountain area consists of a core of basic volcanic and intrusive rock and breccia, all previously included in the Nakina Formation, and associated chert and minor ultramafics. The core is rimmed on three sides by chert, argillite, minor greywacke and small carbonate pods, correlated with the Kedahda Formation, and massive carbonate equivalent to the Horsefeed Formation (Fig. 52.2; Monger, 1975, p. 20-25). Fusulinaceans, small foraminifera and conodonts from carbonate clasts and pods in the core breccia range in age from Early Mississippian to Late Pennsylvanian, those in Kedahda carbonates from Early Mississippian to Permian and those in the Horsefeed carbonate from Late Pennsylvanian to mid-Permian. The contacts between the core and rim rocks are steeply dipping or vertical faults, marked locally by serpentinite bodies. The differing character of apparently coeval core and rim rocks suggests that they have been tectonically juxtaposed and that the bounding faults are parts of a folded thrust. Even though the location of many contacts is certain, it is not known whether they are depositional or tectonic. Consequently, some contacts are shown on Figure 52.2 as "location defined, nature unknown".

A large part of the core consists of a sequence composed of rock fragments of a variety of sizes and compositions. Most common is breccia in which most clasts range from a few centimetres to one metre or more in diameter. The breccia contains sand-sized grains and surrounds unbrecciated bodies more than one kilometre in length. Most clasts and the largest bodies are fine grained, locally amygdaloidal, basalt. Fine grained lithic tuff is common. Carbonate is a minor, but conspicuous component and ranges in size from individual fossils, through cobbles and blocks (Fig. 52.3a) to conspicuous pods more than 100 m long. Chert and argillite are present as small fragments and as pods surrounded by breccia. Serpentinite and rare serpentinized peridotite and gabbro are present in places where the rock is highly faulted. The unsheared nature of many parts of this breccia indicates that it was formed by sedimentary or volcanosedimentary processes such as slumping or sliding, as opposed to tectonic brecciation. Some highly faulted parts contain serpentinite, flattened clasts in a sheared matrix (Fig. 52.3b) and local mylonite. These are presumably sedimentary breccia, secondarily tectonized and qualify as sedimentary melange (of Hsu, 1974, p. 331).

Red, green or tan, thin bedded, manganiferous radiolarian chert that locally passes into red, fine grained, thin bedded, graded crystal tuff, is restricted to the core. It contains pods of well-preserved basaltic pillows, pillow breccia and aquagene tuff. Both chert

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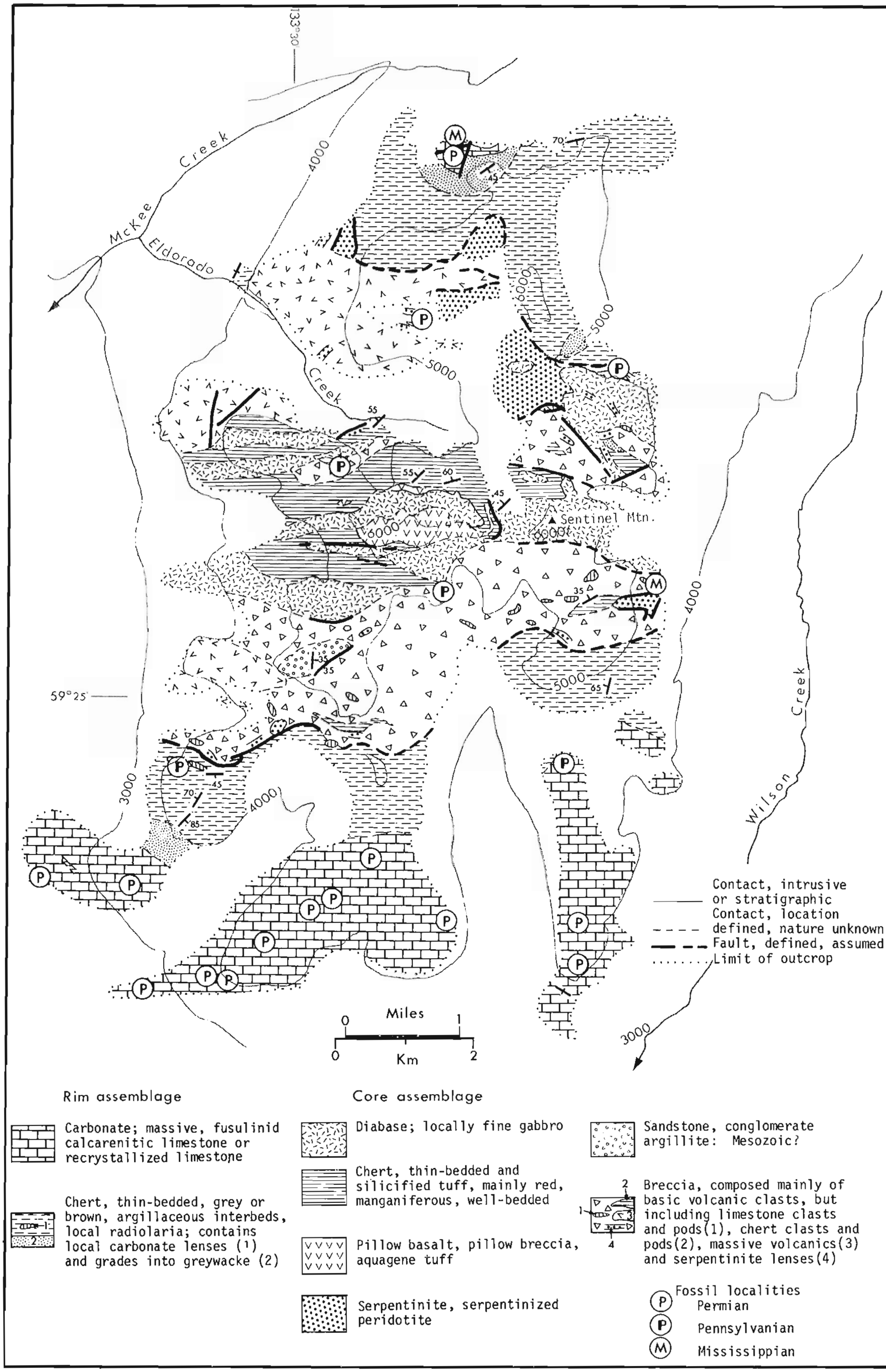


Figure 52.2. Sketch map, Sentinel Mountain area.

and volcanics are intruded by and, respectively, bleached and epidotized, by large diabase and microgabbro sills.

By contrast, chert of the Kedahda Formation of the rim assemblage is invariably grey or greyish brown although it has bedding characteristics similar to the red chert and contains radiolaria. On the west side of the ridge 6 km southwest of the summit of Sentinel Mountain this chert contains a series of aligned carbonate pods that display a depositional relationship with it (Fig. 52.3c). Massive pelletal calcarenitic limestone passes outwards through breccia consisting, successively, of angular unsheared limestone and chert clasts in a calcareous matrix, to carbonate and chert clasts in a cherty argillite matrix to scattered carbonate clasts in chert (Fig. 53.3d). These pods were probably slide blocks that became broken, brecciated and admixed with the cherty matrix during sliding into a chert basin. The contact of the chert with massive carbonate of the Horsefeed Formation is poorly exposed, but appears to be sharp and is probably a fault.

(b) Nakina Lake area (NL)

The writer was accompanied in Nakina Lake area by C.A. Ross, Western Washington State College, who collected 'Tethyan' fusulinaceans, ranging in age from Middle Pennsylvanian to Late Permian, from the Horsefeed Formation to provide a North American reference section for these fossils. The writer concentrated on the relationship between the basal, Upper Mississippian Horsefeed Formation and underlying volcanic breccia, tuff, amygdaloidal basalt and local limestone pods of Upper Mississippian age that were included in the Kedahda Formation (Monger, 1975, p. 14-20). The lower sequence forms the core of a tight, north-northwesterly trending, possibly faulted anticline (Fig. 52.3e). Basal beds of the Horsefeed Formation, a distinctive dark grey limestone containing large brachiopods up to 10 cm across and overlain by crinoidal calcarenite, follow the contact around. The contact is commonly sharp and where tuff underlies the contact it is locally foliated, with foliation parallel to the bedding in the overlying limestone. Rarely, chloritic tuffaceous material is present in the basal limestone. In several places lenses of limestone within the volcanics yield fossils of the same age as the basal carbonate. The contact is apparently a normal depositional relationship along which minor movement has taken place during tight folding. In this place the bank-reef-lagoonal carbonate of the Horsefeed Formation appears to lie on a volcanic basement.

(c) Mount Farnsworth area (MF)

In this area a series of aligned pods of pelletal fusulinid calcarenitic limestone extend over about 5 km in a north-northeasterly direction and are enclosed by thin bedded radiolarian chert, greywacke with abundant chert chips, minor graphitic argillite and pods of brecciated basic volcanics. Where seen, the contact of the limestone pods with broken, sheared chert and argillite is sharp, with no intermixing of carbonate and chert, and is probably tectonic. Thin bedded, brown, ankeritic, carbonate is interlayered with bedded chert in a few localities (Fig. 52.3f). These are perhaps the only unequivocal examples observed in this area of interbedded chert and carbonate in primary depositional contact.

Depositional relationships of Cache Creek carbonates near Atlin

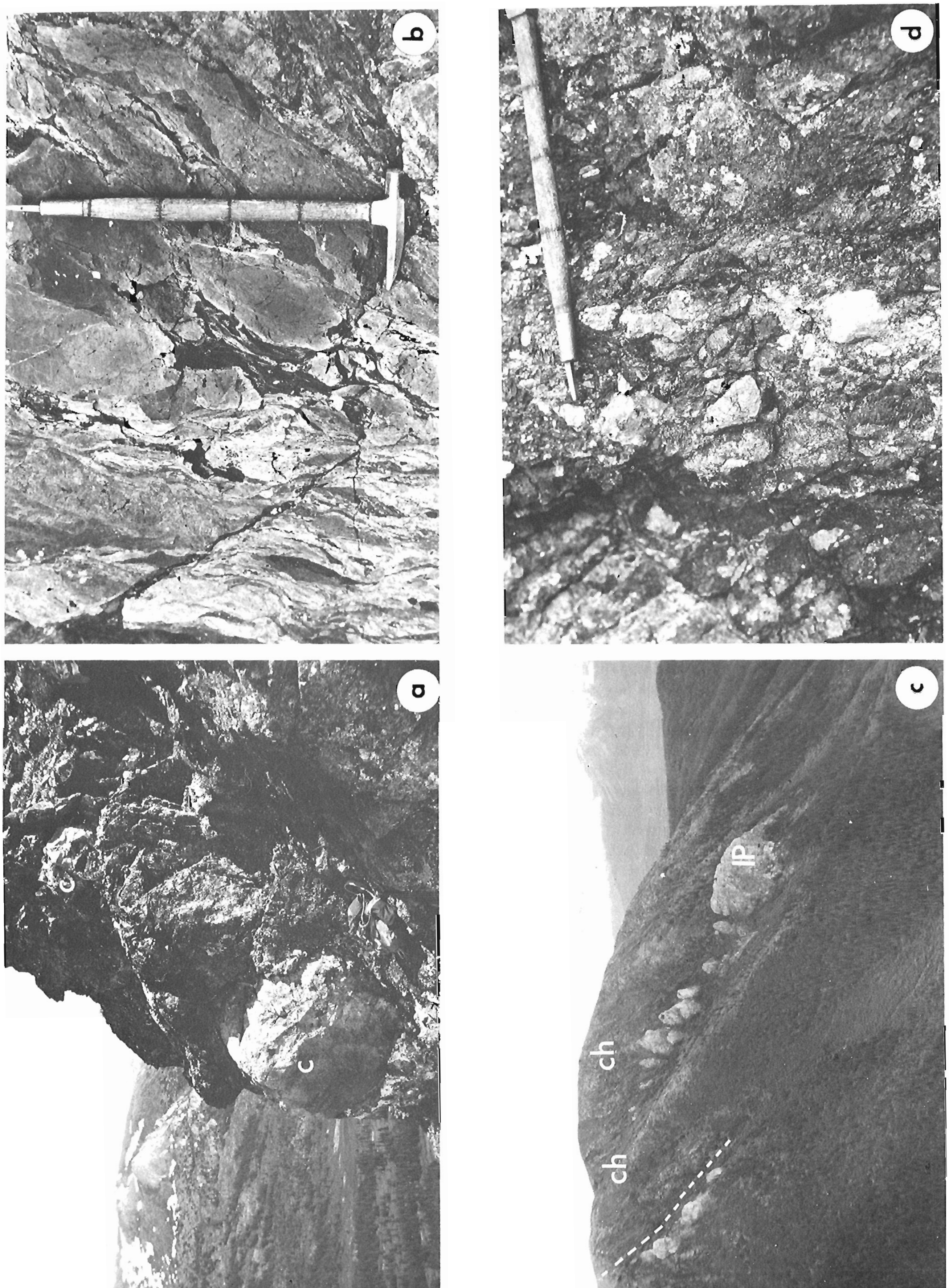
It is difficult to find clear-cut depositional relationships between carbonates of the Cache Creek Group and surrounding rocks. Observed relationships near Atlin suggest that such large, long-lived (Upper Mississippian to Upper Permian) carbonate bodies as the Horsefeed Formation north of Nakina Lake, are founded on volcanic pediments; modern analogues would be atolls. Smaller bodies, such as pods in chert or in breccia are probably allochthonous and arrived in their present settings by submarine sliding. Accepting that thin bedded radiolarian chert represents deep water deposition, then truly deep water carbonate appears to be extremely rare in the Cache Creek Group and is represented only by thin carbonate beds interlayered with chert.

Permian and Older Strata Around the Bowser Basin

An examination of upper Paleozoic rocks near the periphery of the Upper Jurassic-Lower Cretaceous Bowser Basin was undertaken in an attempt to show whether or not they can be correlated on the basis of lithology. Rocks on the western and southern sides in the Stikine, Oweegee Peak, Terrace and probably Fulton River areas can be correlated. Those on the northeast side, extending from Kutcho Creek to south of McConnell Creek map-area, show similar characteristics to one another, some of which are imposed by later metamorphism and deformation.

(a) Kutcho Creek (KC) and Toodoggone (To) areas

The Kutcho Creek area, currently of economic interest because of copper-zinc showings in metamorphosed intermediate to acidic volcanics of uncertain age, was visited with A. Panteleyev and D.E. Pearson, British Columbia Department of Mines and Petroleum Resources, who previously had mapped the area in detail (Panteleyev, 1974, p. 343-346; Pearson and Panteleyev, 1975, p. 86-92). Above the mineralized zone crystalline dolomitic limestone (unit 6 of Pearson and Panteleyev, 1975), contains small corals and brachiopods, and underlies dark grey argillite and argillaceous siltstone (unit 7). The corals are probable scleractinians and therefore post-Paleozoic (E.W. Bamber, pers. comm.). These units closely resemble Upper Triassic Sinwa carbonate and Lower Jurassic Inklin shale and greywacke farther west in Cry and Dease Lake map-areas. The underlying foliated intermediate to acidic volcanics, conglomerate and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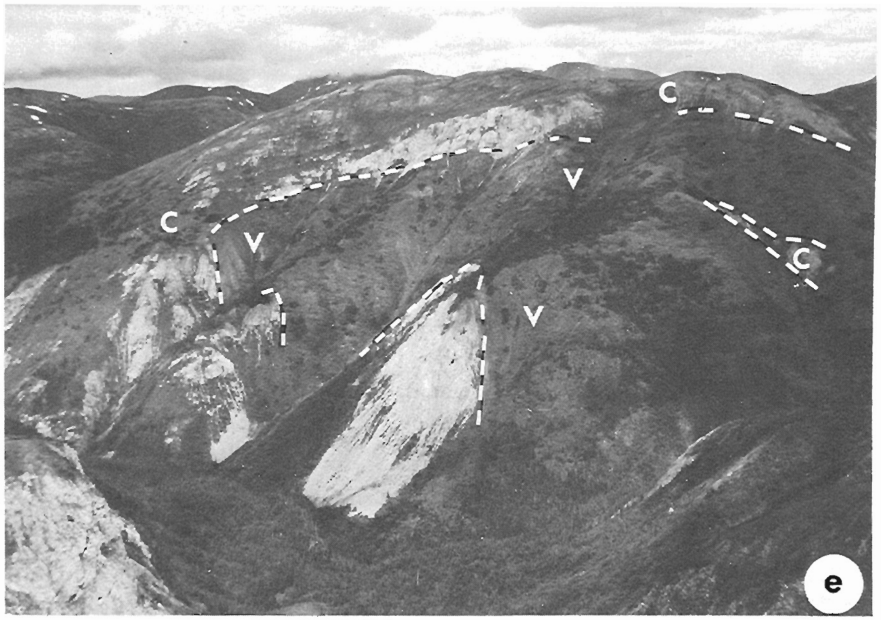

Figure 52. 3. Cache Creek Group near Atlin, B. C.

a: Large (up to 2 m) diameter carbonate clasts (c) in unfoliated breccia composed mainly of volcanic fragments; 2 km southeast of summit of Sentinel Mountain, west side Wilson Creek valley.
b: Foliated volcanic breccia, with a matrix that is locally mylonitic; on ridge 3. 5 km west northwest of summit of Sentinel Mountain.
c: Large pods of Middle Pennsylvanian (P) carbonate, in chert (ch) on west side of ridge, 6 km southwest of summit of Sentinel Mountain. Dashed line indicates approximate location of fault bounding south side of core.
d: Breccia of carbonate and chert clasts in a cherty argillite matrix; uppermost Pennsylvanian carbonate pod in Figure 52. 3c.
e: Base of carbonate of Horsefeed Formation, approximately 9 km northwest of southeast end of Nakina Lake, southeastern Atlin map-area. Upper Mississippian carbonate (c) overlies basalt, volcanic breccia, local chert and argillite (r) forming the core of an anticline.
f: Thin bedded brown carbonate (c) interbedded with grey chert, in creek, 1 km east northeast Mount Farnsworth, Central Atlin map-area.

sericite and chlorite schist resemble the upper Paleozoic in northwestern Toodoggone map-area (Gabrielse *et al.*, 1976). Identical sericite and chlorite schist, phyllite and semischist are involved in thrust faults west and south of the Sikanni Range, McConnell Creek map-area (MC), and can be traced directly into little-metamorphosed Permian Asitka, Triassic Takla and Jurassic Hazelton groups (Monger, 1976; Richards, 1976). Farther south, the probably correlative Sitlika assemblage of Paterson (1974), extends as far as Takla Lake in Fort St. James map-area.

These metamorphic rocks appear to be imbricate slices in a major east- and northeast-dipping thrust system that lies along the west side of the Omineca and

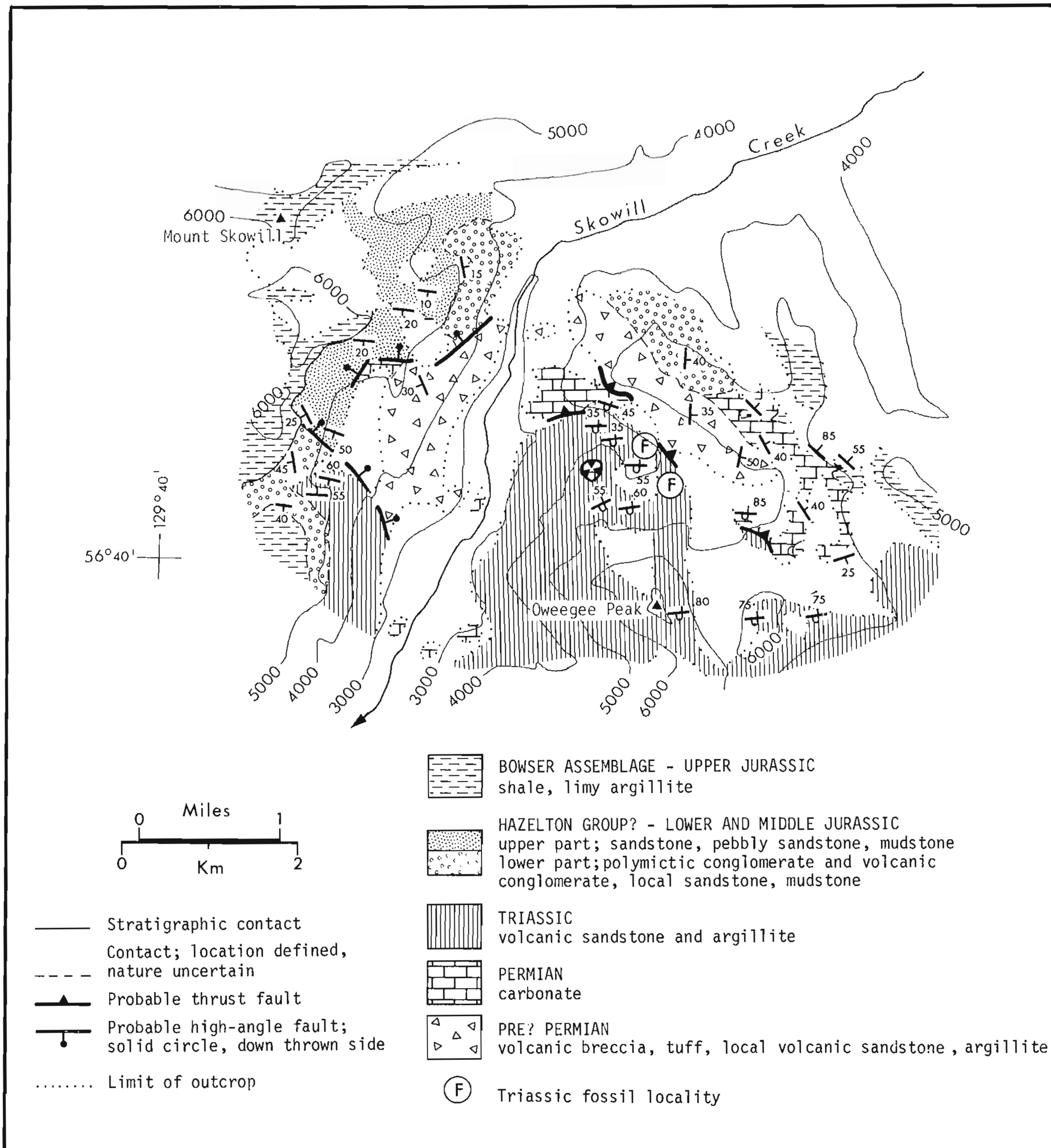

Figure 52.4. Geology near Oweegee Peak.

Cassiar mountains and extends for 700 km from Takla Lake (lat. 55°) in the south to Atlin Lake (lat. 59°) in the north. Structurally above these rocks on the east and northeast is the Cache Creek Group; below them are essentially unmetamorphosed Triassic through Upper Jurassic strata. Stratigraphic and structural relationships and a preliminary isotopic date on metamorphic mica of 110 m.y. (Richards, pers. comm.) indicate the thrusting is a Lower Cretaceous event.

(b) Oweegee Peak area (OP)

This area, studied previously by petroleum companies (J.K. Rigby, pers. comm.; Koch, 1973), consists of a succession of possible pre-Permian through Jurassic strata exposed in a dome surrounded by the Upper Jurassic Bowser assemblage (Fig. 52.4). Oldest rocks are purple and green basaltic breccia, local feldspar and quartz feldspar porphyry, and crystal tuff grading into green chert and minor argillite. These form the core of a large, southwards overturned anticline, outlined by the overlying Permian carbonate. The carbonate consists of approximately 250 m of medium bedded, dark to medium grey, calcarenitic limestone, locally dolomitic with fusulinaceans, corals, bryozoa and brachiopods. The lower half of the section is argillaceous with shale beds between carbonates; the top 30 m is massive, pelletal crinoidal calcarenite. This unit is lithologically similar to and coeval with the lower part of the Permian limestone in the Stikine area (Monger, 1970). Structurally underlying the overturned lower limb of the fold outlined by Permian limestone is a thick succession of grey-green volcanic sandstone, siltstone and dark grey to black argillite with Upper Triassic ribbed belemnites. The upper limb of the fold is apparently unconformably overlain by massive red to grey conglomerate and breccia, containing abundant volcanic clasts and local Permian limestone and granitic clasts. A major episode of deformation thus appears to have followed Upper Triassic deformation and preceded deposition of the conglomerate. On the north side of Skowill Creek the conglomerate is at least 400 m thick, contains abundant granitic boulders and local mud-cracked red mudstone lenses. It closely resembles the polymictic conglomerate at the base of the Hazelton Group in McConnell Creek map-area (Monger, 1976, p. 54). It is overlain by up to 300 m of green, regularly bedded, locally red volcanic sandstone and pebbly sandstone. All of these rocks are possibly correlative with the Hazelton Group. They are unconformably overlain by Upper Jurassic shales of the Bowser assemblage. An episode of block faulting apparently preceded Bowser deposition, for the Bowser is not interrupted above a fault between older polymictic conglomerate and green sandstone northwest of Skowill Creek.

(c) Terrace area (Te)

Weakly to unmetamorphosed Paleozoic rocks are best preserved on both sides of the Zymoetz (Copper) River in possibly faulted septa along the eastern margin of the Coast Mountains granitic rocks. The oldest rocks are grey-green cherty volcanics. These are apparently overlain by at least 200 m of brown weathering, dark grey argillite, siltstone, and calcareous argillite, and thin- to medium-bedded argillaceous limestone with scattered worm tubes, bryozoa, fusulinaceans, corals, bryozoa, brachiopods and crinoids. Above this are up to 300 m of medium to massively bedded, pale grey to buff, locally dolomitic or cherty, calcarenitic limestone with abundant fossils similar to those below. South of Zymoetz River, red tuff laminae and red, tuffaceous limestone horizons are common in the upper part of the section, and north of the river as much as 100 m of green and red lithic tuff are intercalated with the base of the upper carbonate. This calcareous sequence resembles that in the Oweegee and Stikine areas in bedding characteristics, fossils, the argillaceous lower part, and the nonargillaceous, more massive upper part, but differs because of the presence of volcanics in the upper part. The limestone may be folded into large recumbent folds although none were mapped; an apparent fold is visible in the cliff on the south side of Zymoetz River valley, 5 km northeast of Mount Attree. The limestone is overlain by a distinctive conglomerate, with abundant Permian limestone and rare chert clasts in a reddish matrix. The conglomerate appears to fill channels cut into the limestone, and grades upwards through tuffaceous, quartz-eye, pebbly sandstone into red and green volcanic breccia composed mainly of fine grained feldspar porphyry clasts. These rocks were mapped as Triassic (?) by Duffell and Souther (1964) but are identical to and conterminous with rocks mapped by them as Jurassic Hazelton Group, and it seems likely that they are the basal part of the Hazelton Group. A Triassic fossil was collected from an unknown locality along the Zymoetz River by Hanson (1926), so that possibly here, as in the Oweegee Peak area, Permian and Triassic rocks were folded together and are overlain unconformably by the Lower Jurassic strata.

(d) Fulton River area (FR)

Outcrops exposed near the dam on lower Fulton River consist of, from west to east, calcareous phyllite containing large, flattened horn corals, laminated marble, with small isoclinal folds and local crinoid columnals, dark grey cherty argillite and chert, and red-grey to green tuff, and tuffaceous limestone with abundant crinoids, bryozoa, and small corals. Preliminary identifications of conodonts from the phyllitic carbonate are Carboniferous and from the tuff, Permian (B.E.B. Cameron, pers. comm.).

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GEOLOGY OF THE NAHLIN ULTRAMAFIC BODY, ATLIN AND TULSEQUAH MAP-AREAS, NORTHWESTERN BRITISH COLUMBIA

Project 720041

J. Terry[1]
Regional and Economic Geology Division

In order to make a comparison between the Nahlin ultramafic body of northwestern British Columbia (map-areas 104 E, N) and the Pindos ophiolites of Greece, rocks earlier mapped by Aitken (1959) and Souther (1971) were re-examined during the field seasons of 1972 and 1976. Field work in 1972 was funded by a National Research Council Grant obtained by K.C. McTaggart (U.B.C.) and was supported by the Geological Survey in 1976. Travel between Canada and France was paid by the French "Ministere des Affaires Etrangers".

The Nahlin body is 100 km long and up to 8 km wide and is the largest alpine-type ultramafic in the Canadian Cordillera. On the southwest it is faulted against Triassic volcanic and Jurassic sedimentary rocks, and on the northeast against upper Paleozoic rocks of the Cache Creek Group (Fig. 53.1). Largely included within it is an elongated quartz diorite pluton, the Nakina River stock, that intrudes both the ultramafic body and possible Triassic sedimentary and volcanic rocks.

The Nahlin body is composed of (a) well foliated ultramafic rocks (unit 1, Fig. 53.1), (b) ultramafic and mafic rocks with cumulate textures (unit 2), (c) a composite group of rock types transitional between cumulates and volcanic rocks (unit 4, in part) and (d) dyke rocks that intrude the ultramafics but whose genetic relationship to the other groups is not yet certain (unit 3). Though partly disrupted, the Nahlin ultramafic body thus has lithologies that are characteristic of ophiolitic assemblages.

The Nakina Ophiolite Suite
(units 1, 2, 3, 4)

Unit 1: Foliated ultramafic rocks

Buff or slightly khaki-weathering peridotites are by far the most common lithology. They are mainly light or dark green, fresh or only slightly serpentinized rocks that range from dunite almost devoid of pyroxene to peridotites with a high (40 per cent) pyroxene content. Though several differing types may be distinguished in the field by the pyroxene content or by grain size, they rarely make up units of mappable extent but rather grade from one type to the other within a few metres. All types appear to be tectonites with a variable degree of deformation. Olivine and pyroxene grains are strained, kink-banded and sometimes highly elongated. Field observations suggest a direct relation between the pyroxene content of the peridotite and the degree of deformation in that the elongation and granulation of the crystals appear to increase systematically from coarse, pyroxene-rich peridotite to the usually finer grained, dunitic peridotite. This feature is probably complicated by diverse degrees of annealing and recrystallization and needs further microscopic examination. Chrome spinel is a widespread accessory mineral and is most common in the dunitic rocks where it may form schlieren up to a few centimetres thick that are grossly conformable with banding of the host peridotite.

Pyroxenites appear as sharply defined layers commonly a few centimetres thick (and rarely up to 25 cm) throughout the entire unit, and also as later thin veins and dykes rarely more than 30 cm thick that cross both the pyroxenite banding and foliation. The pyroxenite dykes are generally coarser grained than the layered pyroxenites and the crystals, almost undeformed, frequently attain several centimetres in lenth. Preliminary examinations show that the pyroxenites are orthopyroxene or orthopyroxene + clinopyroxene rocks with or without olivine and minor chrome spinel. Plagioclase appears in some of the later dykes, and these possibly grade into true pegmatoid gabbro.

Dunite occurs most abundantly as stringers, diffuse bands and elongated patches that are intimately associated with the pyroxene layers. A lesser but still important volume of dunite is represented by tabular or amoeboid bodies a few metres wide and rarely exceeding hundreds of metres in length that seem to have developed *in situ* without any appreciable distortion of the early structures. The third type is, by comparison, relatively insignificant in volume. Narrow, steeply dipping dunite zones cut the main structural trends and contain a secondary foliation almost normal to the predominant one. In a few cases evidence of displacement was found across these zones, suggesting that the dunite bodies are recrystallized mylonite.

Internal structures. Earliest recognizable features within the peridotite are the narrow pyroxenite layers described above. The layers are most common in peridotites containing less than 20 per cent pyroxene crystals; the thickest layers are closely associated with pyroxene-poor peridotite and dunite, and are very rare in the coarse grained, pyroxene-rich peridotite. The common thickness of the layers is 3 to 4 cm but in tightly folded areas they generally are thicker. In many places they are sliced into very thin stringers by later foliation planes and elsewhere they are boudinaged.

The layering was subsequently deformed by pervasive plastic deformation that includes the development of a strong foliation and at least two episodes of folding. These deformations predate all

[1]Sciences de la Terre, Université des Sciences et Techniques de Lille, France.

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Geol. Surv. Can., Paper 77-1A (1977).

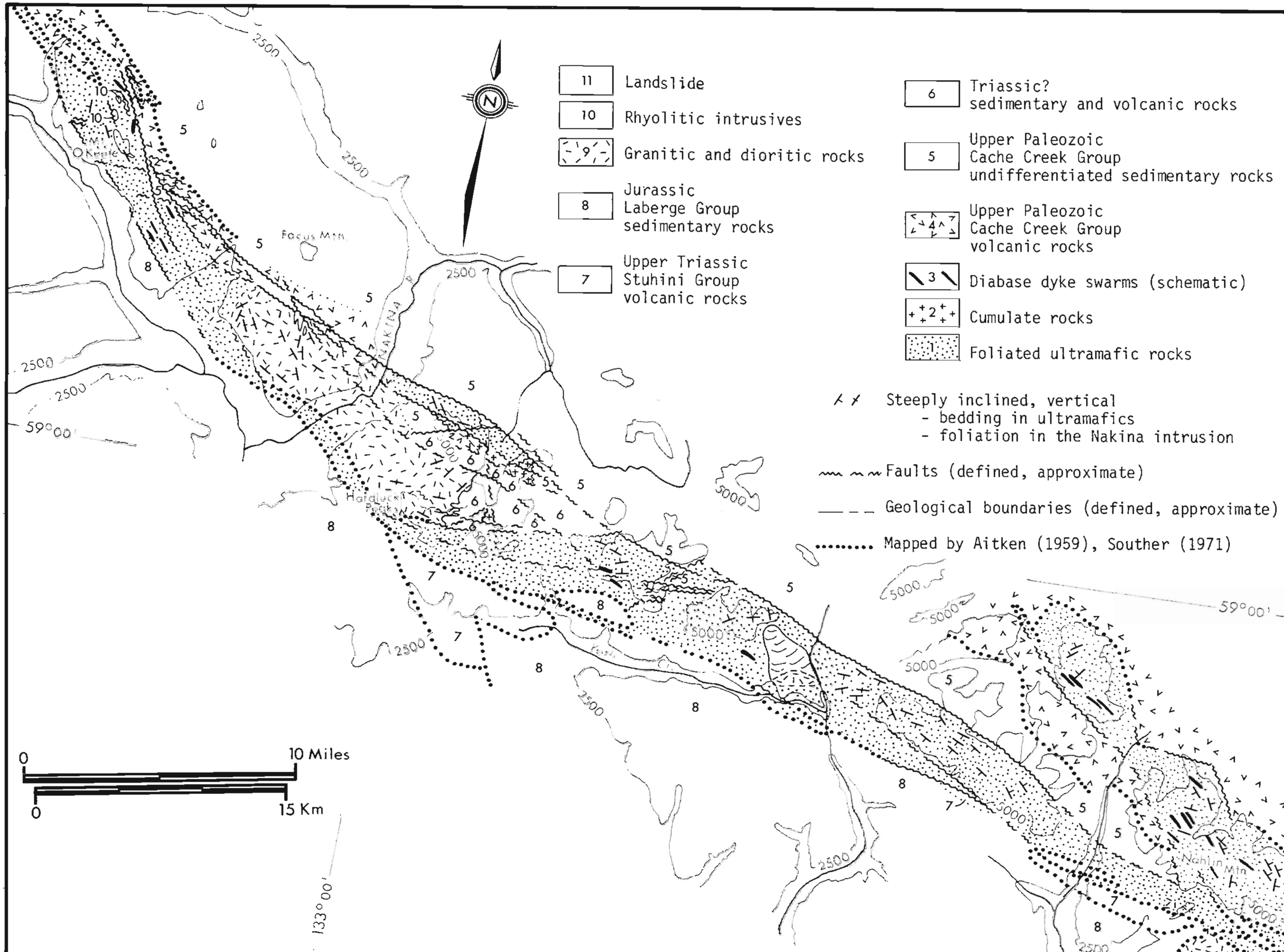

Figure 53.1. Geological map of the Nahlin ultramafic body northwestern British Columbia.

minor intrusions and the foliations cut by pyroxenite dykes and veins that seem to have originated within the ultramafic body itself. The earliest folds that affect the pyroxenite layering are tight, similar, and in many cases, isoclinal. The foliation of the ultramafics is parallel to the axial surfaces of these folds. Some folds have moderately curved hinges, whilst others show sharp, stretched hinges with numerous offsets of the layering along the axial cleavage surfaces, thus generating swarms of narrower layers. It is possible that these structures are the result of several superimposed episodes of isoclinal folding.

The latest folds are large scale, regular, open undulations that fold the earlier foliation and banding. They are southerly-dipping, monocline-like folds tens of metres in wavelength, with a pronounced thinning of the least inclined limbs, in which thick, cross-cutting, pyroxenite veins are commonly developed. These structures are particularly well displayed between the southeastern end of the Nakina stock and Nahlin Mountain.

Unit 2: Ultramafic and mafic cumulates

These rocks range from peridotite cumulates to gabbro. The peridotite cumulates consist of well sorted and rounded olivine crystals a few millimetres across enclosed by pyroxene that locally forms large poikilitic crystals up to 2 cm across. Minor interstitial plagioclase is sometimes present. Pyroxenite cumulates consist of tabular, undeformed crystals, a few millimetres to up to 2 cm in length, that generally display a platy fabric and are locally size graded. The gabbroic cumulates are fine grained, greyish plagioclase-pyroxene rocks with a strong preferred orientation of the crystals.

Cumulates occur in several scattered localities. One of the best sections is on the ridge 4 km northeast of Hardluck Peaks, where the cumulates are faulted against foliated ultramafics on one side and grade into coarse gabbro and trondhjemitic rocks that intrude volcanics on the other. They consist of steeply dipping, repetitive sequences of layers that form ultramafic units 1 to 2 m thick and gabbro units 1 to 15 m thick. Individual layers have either sharp or gradational contacts and are locally unconformable. The total observed section is about 500 m thick, but was probably thicker originally, for the contact with the foliated ultramafics partly truncates the layering. Other cumulate localities are (a) 3 km southeast of Mount O'Keefe where they form sheared outcrops of limited extent on both sides of a sequence of pillow basalt and agglomerate, (b) on the highest cliff of Hardluck Peaks where they grade into trondhjemite, and (c) as numerous blocks associated with dislocated, rodingitized pegmatoid dykes in highly sheared serpentinite approximately 12 km west of Nahlin Mountain.

The cumulates show a classical intrusive relationship with volcanic rocks lithologically similar to and correlated with volcanic rocks of the Cache Creek Group (unit 4) in the area north of Hardluck Peaks discussed above. Fine grained, uniform gabbro grades into coarse gabbro with highly variable grain size. This in turn ranges from gabbro with coarse diabasic texture that is locally crowded with angular to rounded, partly recrystallized fine grained, dark green xenoliths, through an agmatite-like rock with gabbro dykes and veins into fine grained diabase of volcanic rock. The latter grades progressively into more massive volcanics. Veins and irregular patches of fine grained, leucocratic rocks presumably of trondhjemitic composition occur in the agmatitic rocks.

Typical diabase in the volcanics locally contains internal chilled contacts, but no consistent, repetitive pattern of sheeted dykes was found.

Unit 3: Diabase dyke swarms

Numerous subparallel bodies of basic rocks displaying diabasic texture and with chilled margins intrude the foliated ultramafic rocks with a trend that is very similar to the present trend of the Nahlin ultramafic body. Widths of these intrusions range from 1 to 25 m and many of the thickest bodies are multiple intrusions. Northwest of Nahlin Mountain there is a swarm of more than fifteen dykes that cut both sheared serpentinite and fresh unaltered peridotite. They are also abundant near Mount O'Keefe but many of them are completely disrupted in the enclosing sheared serpentinite.

All diabasic intrusions predate observed faults, they are not present in surrounding sedimentary strata, and their relationship with the volcanics of the Cache Creek Group need to be tested by further petrographic and chemical studies.

Late Intrusions

(a) Minor dioritic and acidic intrusions

A variety of minor elongated intrusive bodies, rarely more than 20 m wide, parallel the main regional trend along the length of the Nahlin ultramafic body. Most of them are dislocated by late faults and their original form is obscure. They range from rhyolitic plugs near Mount O'Keefe, through quartz porphyry with large feldspar phenocrysts, hornblende and biotite, to dark hornblende diorite containing numerous mafic xenoliths.

(b) Nakina River Stock

This stock was examined in detail because it is largely within the Nahlin ultramafic body. Although its present elongated shape is accentuated by truncation due to later faulting the internal primary foliation parallel with the boundaries of the intrusion indicates that its shape was mainly acquired at the time of emplacement.

The stock is a homogeneous, medium to coarse grained quartz-diorite that contains about 25 per cent hornblende and biotite. Near contacts the rock is locally crowded with xenoliths and has a much higher colour index. Near the eastern contact clusters of hornblende crystals give the rock a distinctive spotted appearance and sparse orbicules are present. Minor rock types are aplite and dark grey, hornblende microdiorite dykes in the outer parts of the intrusion.

A planar arrangement of mafics which can be observed in most parts of the intrusion becomes pronounced towards the contacts. Most foliations dip towards the centre of the stock and become almost vertical near the middle, indicating an elongate, funnel-like shape for the intrusion.

The northeastern and southeastern contacts are largely faulted and exposed primary contacts are of limited extent. South of Focus Mountain hornfelsed and foliated country rocks are almost completely surrounded by quartz diorite and represent either pendants or collapsed blocks. Foliation in the hornfelsed metabasalt and pelite is concordant with that in the enclosing quartz diorite. In the Hardluck Peaks area, quartz diorite clearly intrudes and thermally metamorphoses the Triassic (?) sedimentary and volcanic assemblage (unit 6) and the ultramafics are injected by long tongues of quartz diorite.

Structure of the Nahlin ultramafic body

High angle faults are the major regional structures. The Nahlin fault zone is complex and appears to result from several episodes of deformation and is a series of high-angle, northeasterly-dipping or almost vertical faults. Carbonatized serpentinite masses along most of the length of the body are particularly well-developed in the Mount O'Keefe area. These retain an older foliation that dips generally less than 50° northeasterly, an attitude that may support the suggestion by Monger (1975) that the body is emplaced along a major, northeastward dipping thrust zone. The high-angle faults near Mount O'Keefe locally affect small acidic plugs (unit 10) that are tentatively correlated with the Upper Cretaceous and Early Tertiary Sloko Group; thus a post-Early Tertiary age for the faults is indicated.

Numerous late minor faults, transverse to the Nahlin trend, display conspicuous horizontal striations in carbonatized rocks near Mount O'Keefe.

The anomalous trends of foliation and faults in Triassic (?) sedimentary and volcanic rocks (unit 6) are parallel with the contacts of the Nakina River stock and are believed by Aitken (1959, p. 57) to be due to forceful emplacement of the pluton. This foliation is concordant with the internal deformation of the stock.

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54. METAMORPHIC TERRANE, NORTHERN COAST MOUNTAINS WEST OF ATLIN LAKE, BRITISH COLUMBIA

Project 720041

L. J. Werner[1]
Regional and Economic Geology Division

Parts of the metamorphic terrane west of Atlin, British Columbia shown in Figure 54. 1, were studied in detail. Marble forms 25 per cent of the total area, quartz-muscovite to quartz-biotite schist 40 per cent, quartz-amphibolite gneiss 20 per cent, and hornblende amphibolite 15 per cent. The complex nature of folding makes reconstruction of stratigraphy difficult, and the upper-greenschist to mid-amphibolite grades of metamorphism have destroyed any fossils. Thus the age of the terrane is not known; it was mapped previously

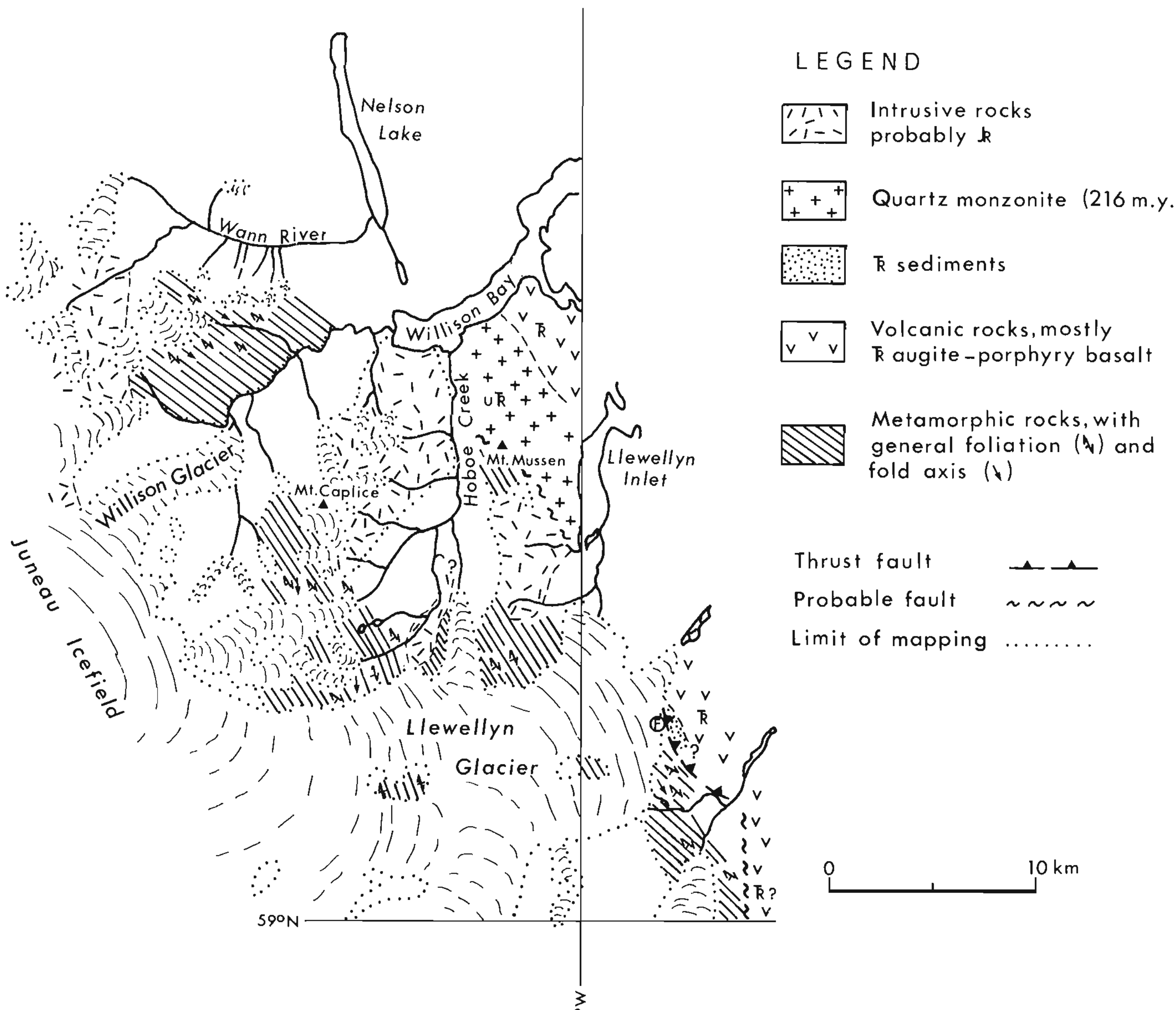

Figure 54. 1. Metamorphic terrane, west of Atlin, British Columbia.

[1]University of British Columbia, Vancouver B. C.

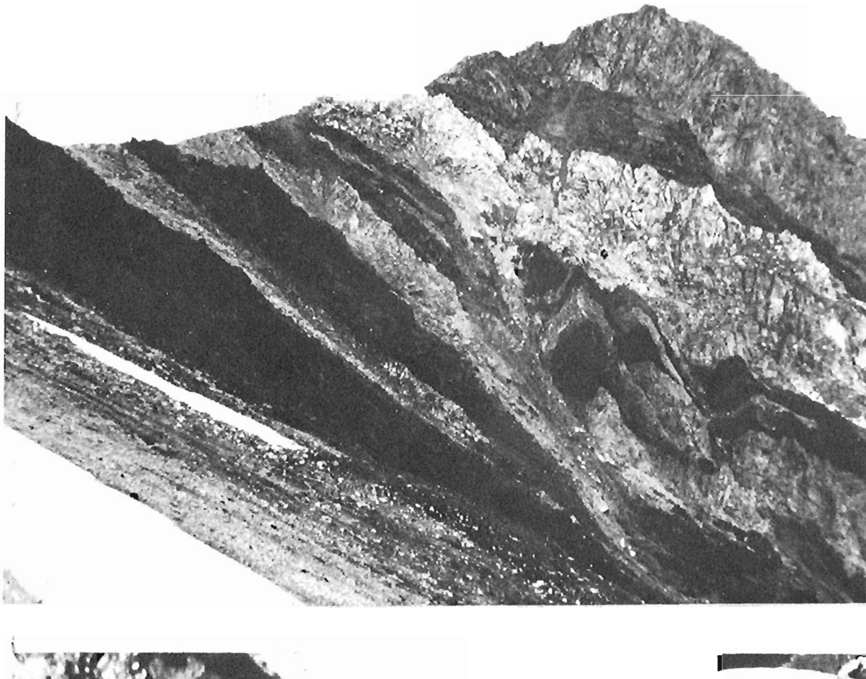

Figure 54. 2.

Interlayered marble and amphibolitic schist forming ridge north of Willison Creek. All layers are internally isoclinally folded.

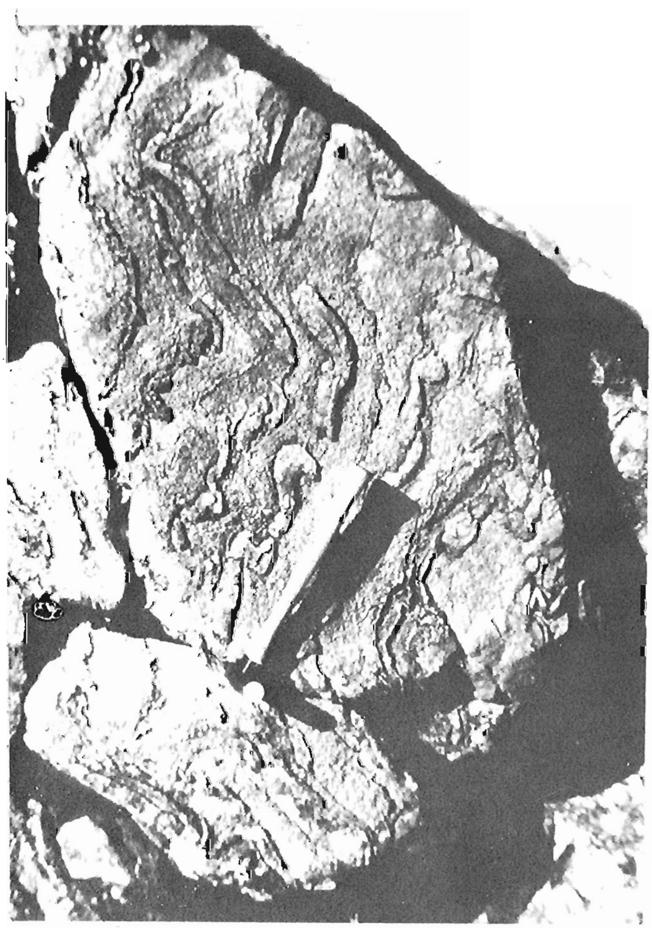

Figure 54. 3. Rootless folds outlined by argillaceous bands in marble.

Figure 54. 4. Polyphase folding in quartz-amphibolite gneiss. Note the layer indicated by the arrows shows closure at both ends.

as Yukon Group of Proterozoic age (Christie, 1957), and no evidence yet indicates that this interpretation is wrong.

Deformation is intense, and at least two and possibly three phases of folding are apparent. Large-scale tight and isoclinal folds, outlined by marble bands, have generally southwesterly dipping axial planes, and variable, southeast plunging axes (Fig. 54.2). These are deformed by nearly coaxial irregular open to light wrinkles on a smaller scale. In some areas a third phase with generally north-south axes is superimposed (Fig. 54.4).

Age of the deformation is not known. An unaltered, undeformed quartz monzonite pluton, dated at 216 m.y. by T.E. Boltmann (pers. comm.) forms the eastern boundary of the terrane on Mount Mussen; the two are probably in tectonic contact. An upper Triassic conglomeratic turbidite sequence on the south side of the toe of Llewellyn Glacier (shown by Aitken, 1959, as being of upper Paleozoic age) contains both metamorphic and granitic clasts. This sequence structurally overlies the metamorphic terrane on an east-dipping thrust fault. A minimum age for the terrane is given by the probable upper Mesozoic Coast Intrusions.

Extensive intrusion of granodiorite-quartz monzonite rocks occurs nearly everywhere, mainly as veins and small plutons. These are most probably related to the Coast Intrusions, which form the western contact of the terrane.

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COAST MOUNTAINS PROJECT

Project 630016

J. A. Roddick and G. J. Woodsworth
Regional and Economic Geology Division, Vancouver

Snow, unmelted because of the worst spring in memory (Fig. 55.1), followed by the worst summer weather ever recorded on the coast severely hampered field operations in the Coast Mountains. Three weeks were spent re-examining critical areas in Bute Inlet map-area, and about five weeks were spent in unmapped parts of the Mount Waddington map-area.

In the southeast corner of Bute Inlet map-area (92K) near Beartooth Mountain an assemblage of basaltic to dacitic volcanic rocks are preserved in a small cluster of pendants. The lower part of the sequence consists of basalt flows, some of which show good pillow structure (Fig. 55.2) and are indistinguishable from the Upper Triassic Karmutsen Formation. If correctly correlated, this represents the farthest from the coast (40 km) that the formation has been recognized in the western Coast Mountains. The more acid overlying volcanic breccias and flows might be equivalent to the Bonanza Formation on Vancouver Island. A dioritic complex underlies the intervening areas among the

Figure 55.1.

View west across Bute Inlet from Mount Sir Francis Drake, Bute Inlet map-area. July 30, 1976.

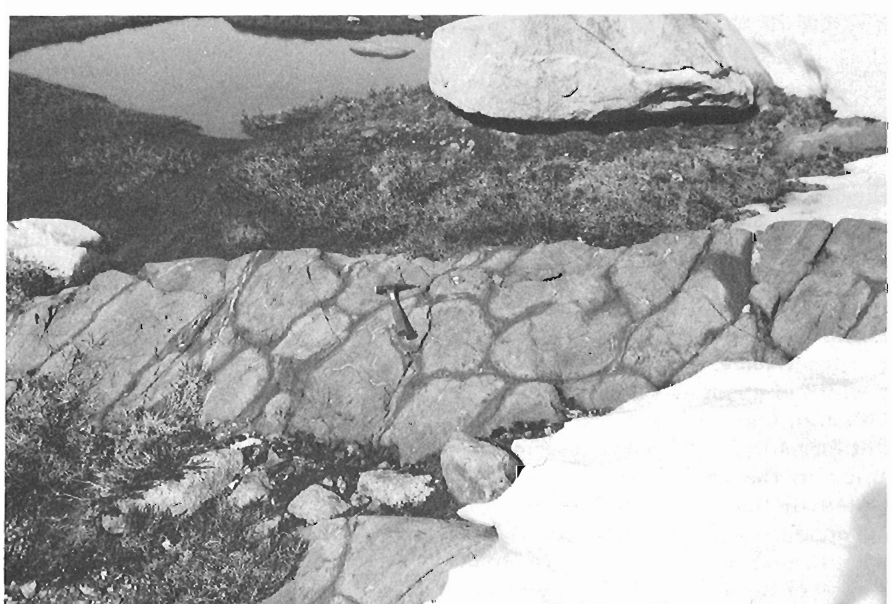

Figure 55.2.

Pillow lava 9 km southeast of head of Powell Lake, Bute Inlet map-area. Rocks are only slightly deformed and metamorphosed.

From: Report of Activities, Part A;
Geol. Surv. Can., Paper 77-1A (1977)

Figure 55. 3.

Granitoid gneiss 6 km southwest of Mount Grenville, Bute Inlet map-area. Dark layers are amphibolite; lighter material is granodiorite.

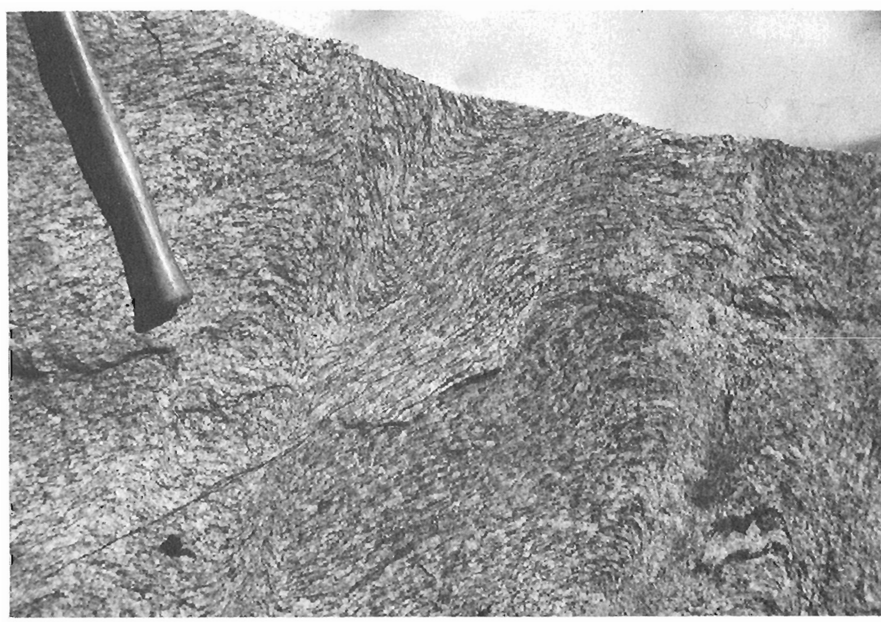

Figure 55. 4.

Folds in gneissic granodiorite of Central Gneiss Complex, 6 km southwest of Mount Grenville, Bute Inlet map-area.

pendants and extends somewhat beyond them. Parts of the complex may represent pre-Karmutsen terrane but the critical relationships have not yet been established.

In Mount Waddington map-area the main unit is a broad band of gneiss and migmatite (Figs. 55. 3, 55. 4), extending from near the southeast corner at least 100 km to the northwest and reaching a maximum width of about 20 km northwest of Mount Waddington. Although the northwestern limits are not defined the unit forms the largest exposure of Central Gneiss Complex in the southern Coast Mountains. It is interrupted in the Homathko Icefield by a large intrusion of granodiorite and quartz monzonite, but then continues southwest to the Mount Raleigh area in the northeast corner of Bute Inlet map-area where it was found to be unconformably overlain by less deformed Lower Cretaceous strata. An examination of a few exposures of the gneisses in the vicinity of Tatla Lake (which is in the Intermontane Belt well beyond the northeast front of the Coast Mountains) indicated that they may be correlative with the Central Gneiss Complex. The extent in the Intermontane Belt of this partly mobilized, possible Precambrian, gneissic terrane is not known.

In 1967 a 230-km traverse across the Coast Mountains was begun on the west coast of Calvert Island and carried across Rivers Inlet map-area into Mount Waddington map-area as far as Knot Creek. In 1976 the final 35 km of the traverse, which extended it into the Intermontane Belt, was completed. The most notable feature encountered by this segment was a broad (2 km) northwest-trending shear zone best exposed about 4 km northeast of Knot Creek.

STRUCTURE AND STRATIGRAPHY OF THE BIG BEND AREA, BRITISH COLUMBIA

E.M.R. Research Agreement 1135-D13-4-107/76

R.L. Brown[1], M.J. Perkins[1], and C.R. Tippett[1]
Regional and Economic Geology Division

Detail mapping in the Big Bend region was initiated in 1973 in order to further unravel stratigraphic and structural problems brought to light by reconnaissance mapping (Wheeler 1963, 1965). Brown was assisted in the summer of 1973 by J. Franzen (M.Sc. thesis, 1974), in 1974 by D. Shaw, M. Perkins, and J. Van der Leeden (M.Sc. thesis, 1976), in 1975 by D. Shaw, M. Perkins, and C. Tippett (M.Sc. thesis, 1976), and in 1976 by D. Shaw, M. Perkins, and L. Lane. The following is a brief outline of the results.

The stratigraphy of the Horsethief Creek Group has been established within the region outlined in Figure 56.1. Three distinctive map units are recognized which, despite facies changes and complex structures, can be traced from at least north of the Trans-Canada Highway northwestward to the Big Bend of the Columbia River.

Thicknesses have been greatly altered by deformation; the total exposed section is approximately 4000 m with an unknown depth to the base. The lowest exposed unit,

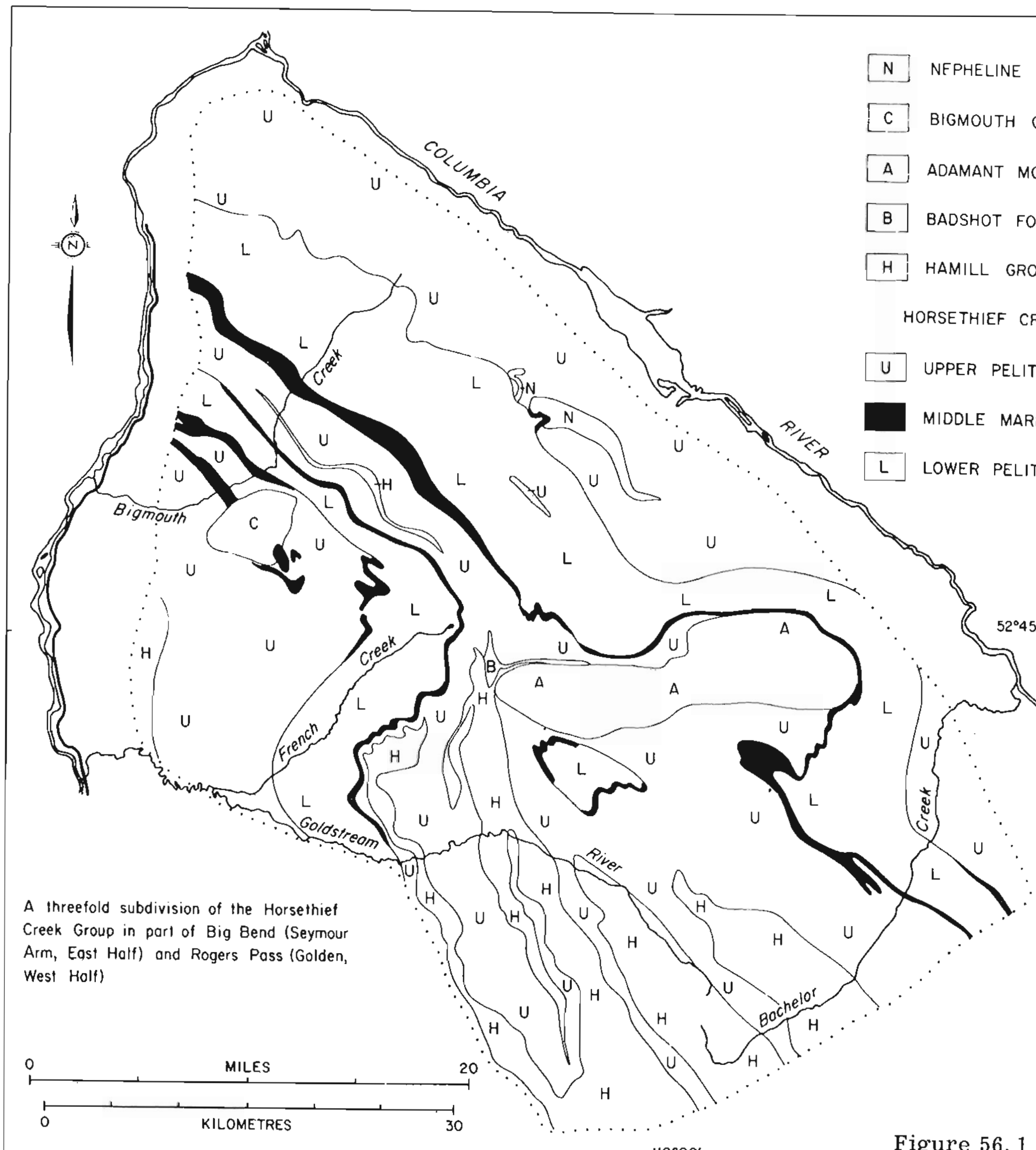

Figure 56.1

[1]Department of Geology, Carleton University, Ottawa, Ontario.

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Geol. Surv. Can., Paper 77-1A (1977).

informally named Lower Pelitic member, is characterized by the presence of amphibolite bands and lenses in thinly interbedded pelite and semipelite together with some thin, discontinuous marble bands. The amphibolites are considered to be primarily metasediments, but thick, massive and locally discordant units appear to be of intrusive origin. Up to 900 m of exposed Lower member has been observed, above which lies the Marble member varying from a prominent carbonate sequence (~ 600 m) with massive grey marbles in the southwestern part of the region to thin discontinuous impure and dolomitic marble bands in the northeast. The Upper Pelitic member characteristically is dark, finely laminated pelite to rusty calcareous pelite with locally thick beds of feldspathic grit and conglomerate. There is an increase in carbonate from northeast to southwest which is accompanied by a decrease in abundance of coarse clastic material. Thin, discontinuous marble bands occur locally in the calcareous pelitic assemblage. The Upper Pelitic member attains a thickness of up to 2500 m, and is conformably overlain by the dominantly psammitic rocks of the Hamill Group. Extensive sliding has occurred along the Hamill-Horsethief Creek boundary locally cutting out parts of the succession.

The Selkirk Fan Axis has evolved by superposition of two distinct phases of deformation upon strata previously involved in nappe formation (Phase I).

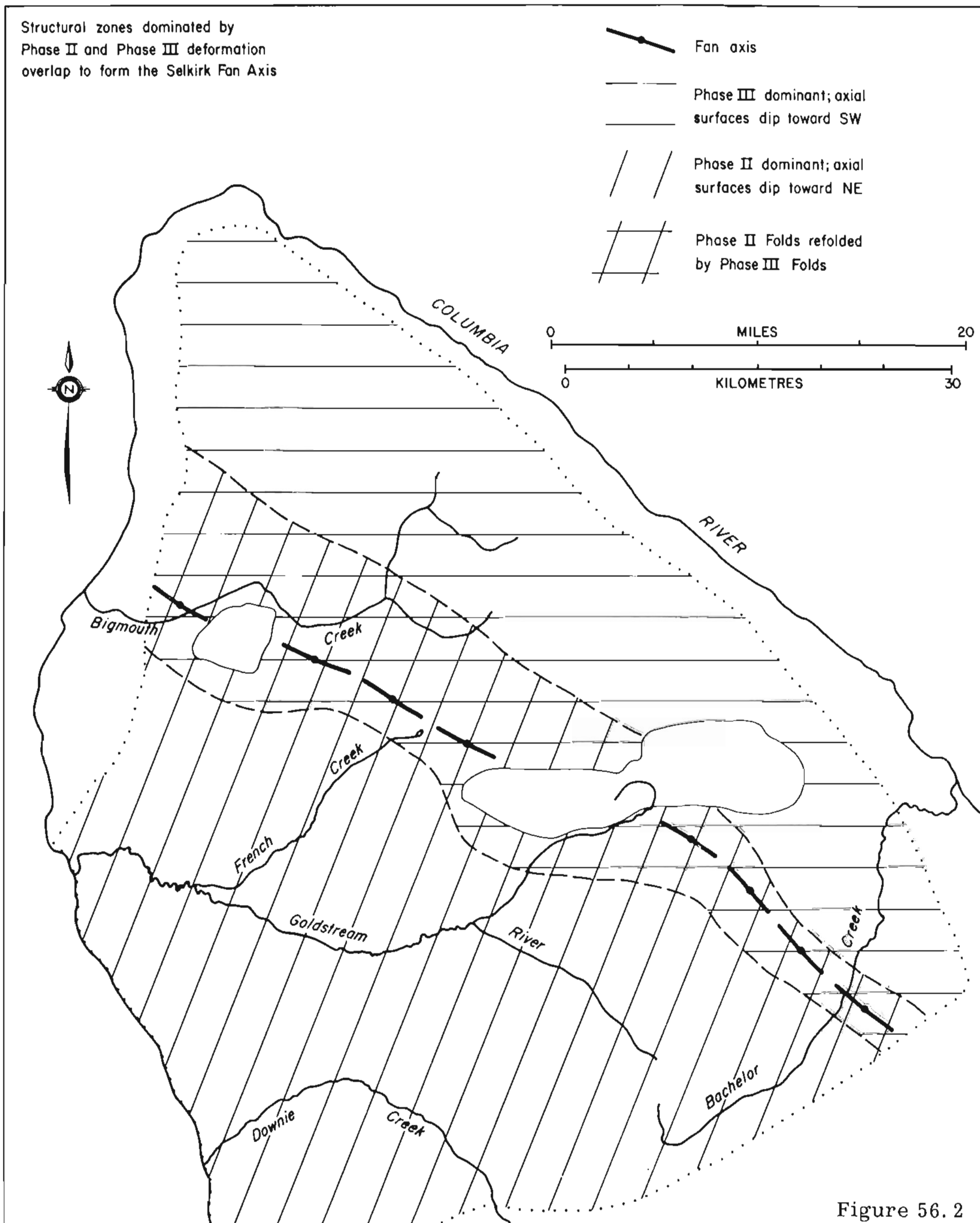

Figure 56.2

Phase II folds are dominant on the southwestern flank of the Selkirks where they are strongly overturned toward the southwest. Phase III folds are dominant on the northeastern flank where they are overturned toward the northeast. The axis is located where northeastern dipping Phase II axial surfaces are overprinted and transposed by steeply dipping to vertical Phase III axial surfaces (Fig. 56.2).

Inversion of stratigraphy occurred in the western part of the area during the early nappe phase, and superimposition of Phase II folds formed tight to isoclinal antiformal synclines and synformal anticlines. There is extreme variation in attitude of axial surfaces and hinge lines of folds throughout the area with hinge line pitch in axial surfaces varying from 0 to 90°. Phase II and Phase III hinge lines generally have similar attitudes, but axial surfaces of Phase III folds are regionally steeper than those of Phase II. Phase I hinge lines vary from 90° to nearly coaxial with adjacent Phase II or Phase III folds.

Metamorphism which culminated between Phase II and Phase III achieved sillimanite grade without muscovite breakdown, in the central part of the area, and was accompanied by granitic plutonism.

The Adamant Pluton (Fox, 1969) was initially emplaced before Phase II and it behaved as a semirigid body during the regional deformation (Shaw, Ph.D. thesis in progress).

The structural evolution of the fan may be explained in terms of initial underthrusting of the southwestern flank of Selkirk terrain by the Shuswap-Monashee complex followed by underthrusting of the northeastern flank by basement rocks of the Rocky Mountain Foreland. The structures appear to have evolved by episodic compression without significant uplift, rather than by gravitational upwelling.

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Project 730067

Peter B. Read[1]
Regional and Economic Geology Division

Nearly 65 years ago hot springs were discovered along Meager Creek and Lillooet River west of their confluence in the west half of map-sheet 92 J (Robertson, 1911). During 1974 and 1975 a program consisting of water geochemistry, resistivity and self potential surveys, and diamond drilling totalling 2523 feet has contributed data for an assessment of the geothermal potential of the area. In the summer of 1976, nearly two months were devoted to mapping of the Pliocene and Recent volcanic rocks outcropping west of the junction of Meager Creek and Lillooet River. L. Hammerstrom, Department of Geological Sciences, University of British Columbia, is engaged in a detailed study of spring water geochemistry.

Nine volcanic assemblages of the Garibaldi Group form the Meager Creek Volcanic Complex. Andesite and dacite flows predominate and range in age from Pliocene and(?) Miocene to postglacial. The complex overlies a basement composed of Mesozoic plutonic and metamorphic rocks which are unconformably overlain by low grade metavolcanic rocks all of which are intruded locally by quartz monzonite stocks of presumed Miocene age. The hot springs issue from basement.

Meager Creek Volcanic Complex

In the complex, widespread andesite constitutes most of the older part which is best exposed in the south. In the northern half of the complex, younger dacite flows and lava domes overlie and intrude remnants of the formerly more extensive andesite flows. Approximately one third of the complex, most of it dacite, is postglacial in age. Nine volcanic assemblages comprising the complex and depicted in Figure 57.1 will be described in order of decreasing age.

1. Basal Breccia: Locally preserved remnants of breccia up to 300 m thick overlie basement on the south side of the complex. Clasts of granitic, grey or green aphanitic volcanic, and minor metamorphic rocks lie in a tuffaceous matrix. South of Pylon Peak, where the breccia is thickest, clasts less than 0.5 m long increase in size downwards to jumbled blocks of quartz diorite up to 20 m long with less than 10 per cent matrix. This area, where basement is lowest, may represent a partly exhumed vent and its southern edge apparently coincides in part with the northern margin of an area of anomalously low resistivity (Shore, 1975).

2. Porphyritic Quartz Dacite: In the southwest corner of the map-area, a grey-green dacite with sparse phenocrysts of quartz, plagioclase and hornblende forms a remnant of subhorizontal flows up to 200 m thick. Gently dipping acid tuff and breccia overlap the older dacite along a subvertical eastern contact.

3. Acid Tuff and Breccia: On the south and west flanks of Pylon Peak and The Devastator is a cream to yellow ochre weathering assemblage up to 500 m thick of acid volcanic rocks. They are hydrothermally altered quartz latite with locally preserved quartz, plagioclase and biotite phenocrysts. Silicification, pyritization and the development of ubiquitous clay minerals and sporadic carbonates characterize this unit. Crudely layered tuff and breccia, dipping gently northeastward, compose all but the eastern end of the unit. Here the quartz latite is massive and may represent either flows and/or hypabyssal intrusions of a partly preserved vent.

4. Aphanitic Flows and Minor Intrusions: Medium to dark grey aphanitic flows here and there overlie the basal breccia and acid volcanic units and a few dykes less than 50 m thick cut both units. On the south-southeast ridge of The Devastator a lens of conglomerate composed of subrounded pebbles and cobbles of this lithology overlies the acid volcanic unit.

5. Porphyritic Plagioclase Andesite: Porphyritic plagioclase andesite, the most extensive unit of the complex, forms most of the southern and western parts of the complex. Best outcrops are on Pylon Peak and The Devastator. Gently dipping flows are more extensive than basal and intercalated breccia and tuff, and dykes and plugs are restricted to The Devastator and possibly Peak 7927' at the head of Job Glacier. The maximum thickness may exceed 1200 m of flows south of Capricorn Creek. Flows are commonly flow-layered or have a subparallel platy jointing and thin reddened breccia and tuff lenses may separate flows up to 20 m thick. Monomictic breccias up to a few hundred metres thick of porphyritic plagioclase andesite clasts lie at or within a hundred metres of the base of this sequence. The monomictic composition and differential weathering of the clasts distinguish this breccia from the basal breccia unit. Close to The Devastator, angular clasts up to several metres long are common in breccia. The concentration of hypabyssal intrusions and coarse volcanic breccia in the vicinity of The Devastator favour it as a major andesite vent. Potassium argon dates of 4.2 ± 0.3 m.y. and 2.1 ± 0.2 m.y. (Anderson, 1975) indicate a long period of andesite volcanism spanned by this unit.

6. Hornblende-Biotite Rhyodacite: Surrounding Mount Job in the centre of the complex, are ochre-yellow weathering flows of porphyritic hornblende-biotite quartz rhyodacite. They are prominently flow-layered and locally have columnar jointing. At the head of

[1]Consulting Geologist: Geotex Consultants Ltd., no. 1103 - 100 West Pender St., Vancouver, B.C. V6B 1R8

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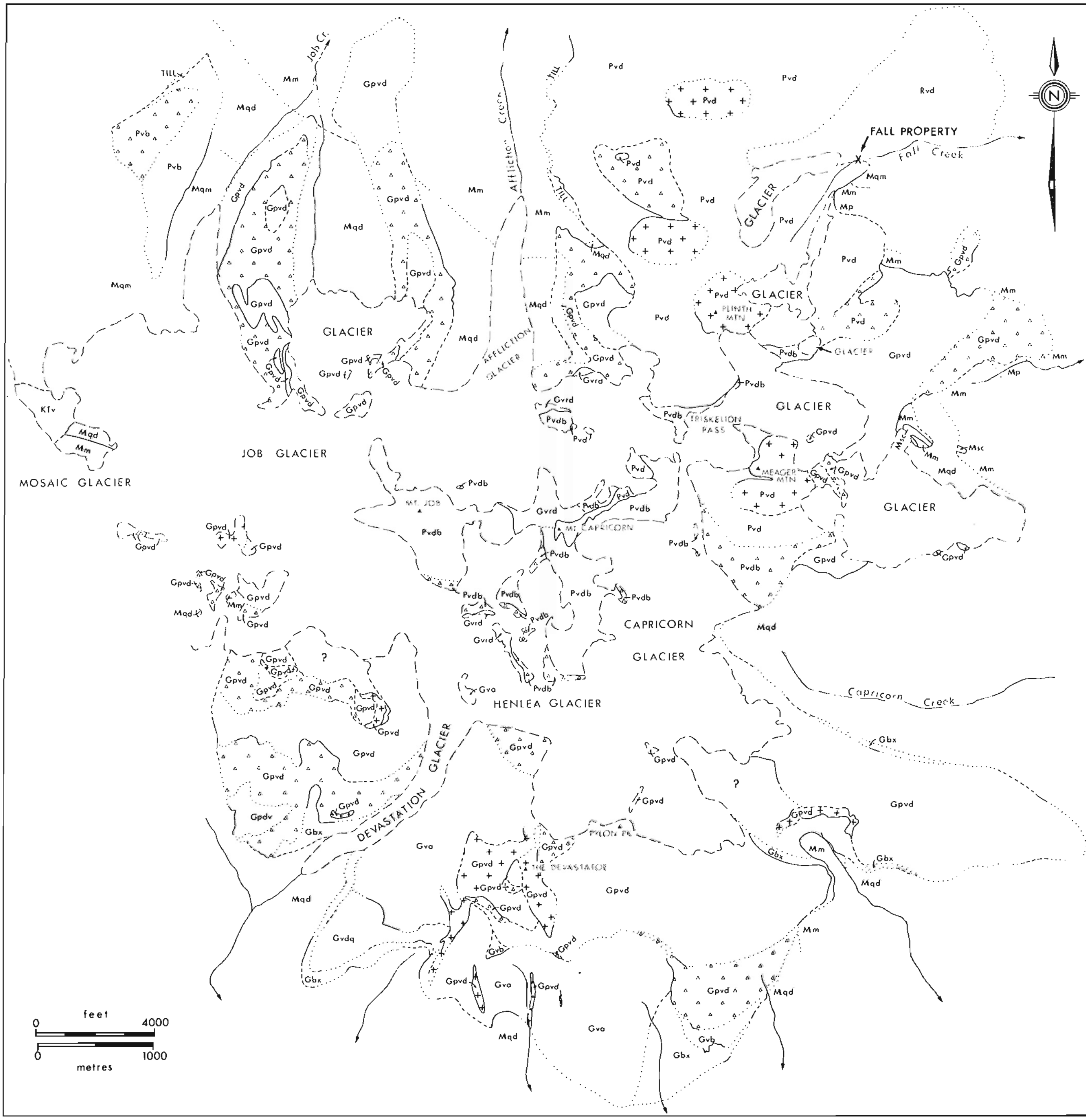

VOLCANIC COMPLEX

- Rvd Scoriaceous dacite
- Pvd Porphyritic dacite
  - breccia and tuff
  - hypabyssal intrusion
- Pvb Porphyritic (plagioclase, olivine) basalt
  - scoriaceous bombs, breccia and tuff
- Pvdb Porphyritic (biotite) dacite
  - breccia and tuff
- Gvrd Porphyritic (hornblende, biotite) rhyodacite
- Gpvd Porphyritic (plagioclase) andesite
  - breccia and tuff
  - hypobyssal intrusion
- Gvb Aphanitic flows
- Gva Acid tuff and breccia
- Gvdq Porphyritic (quartz) dacite
- Gbx Basal volcanic breccia

BASEMENT

- Mqm Biotite or biotite-hornblende quartz monzonite
- KTv Aphanitic flows and breccia
- Mqd Quartz diorite, diorite and quartz monzonite
- Mm Amphibolite
- Mp Phyllite and rusty schist
- Msc Marble

Figure 57.1. Simplified geological map of the Meager Creek Volcanic Complex.

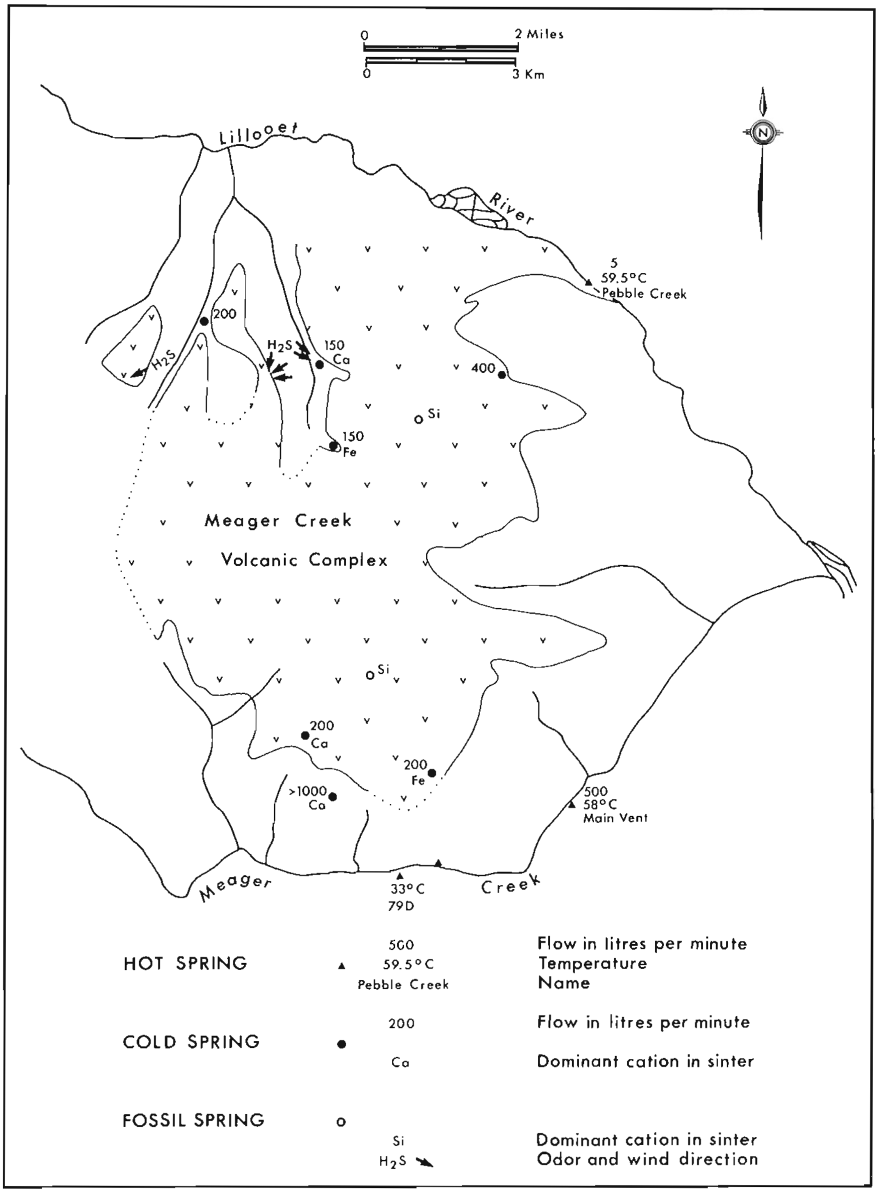

Figure 57.2. Springs of the Meager Creek Volcanic Complex.

Affliction Glacier, the unit attains a maximum thickness of 500 m. On the east side of the glacier, it unconformably overlies porphyritic andesite and at the head of Affliction and Capricorn glaciers it is truncated by porphyritic biotite dacite of Mount Capricorn.

7. Porphyritic Biotite Dacite of Mounts Capricorn and Job: The final 600 vertical m of mounts Capricorn and Job are brick red to maroon-grey weathering dacite. Coarse phenocrysts (5 mm) of plagioclase, quartz, and biotite characterize this vesicular dacite. Angular clasts of dacite up to 2 m long form a basal breccia up to 100 m thick. Similar breccia is interspersed throughout the dacite. On Mount Job local platy and columnar jointing and layering suggest that flows form the bulk of the massif, but their absence on Mount Capricorn may favour this as a source of the eruptive rocks.

8. Porphyritic Dacite of Plinth and Meager Mountains: The top 600 m of Meager Mountain and the bulk of Plinth consists of a light grey porphyritic dacite with medium grained (2 - 4 mm) phenocrysts of plagioclase, quartz, minor biotite, and rare hornblende. The dacite is commonly vesicular, has a glassy matrix, and is distinguished from other dacites by scattered, rounded inclusions of fine grained hornblende andesite. On Meager Mountain, the absence of flows or breccia, and development of steeply inclined flow layering suggest that it is a plug or lava dome. In contrast, Plinth Mountain consists of prominent columnar- or platy-jointed flows and widespread breccia and ash on its northern flank. Only three areas on the north ridge and the flat-topped summit show steep to vertical flow layering and subhorizontal columnar jointing of possible plugs or lavas domes. On Plinth, attitudes of flows commonly are subparallel to topography and dips range from 30 to 60 degrees. East of Affliction Glacier and Creek, basal flows overlie a probable till.

The Bridge River ash incompletely blankets the area between the north and east ridges of Plinth. Within this area, crudely stratified breccia and ash deposits are up to 20 m deep on some ridges. Over 90 per cent of the clasts are cream-weathering, porphyritic (plagioclase, hornblende, pyroxene) dacite pumice. They range in maximum size from 10 cm on the summit of Plinth Mountain (Nasmith *et al.*, 1967) through 1 m at the 6500 foot level on the north ridge crest to 4 m blocks on the north side of the creek crossing the Fall Property at 4965'. Two per cent of the clasts are subrounded pebbles and cobbles of a porphyritic quartz monzonite exposed along the creek. These data strongly indicate the lower part of the valley as the source of the Bridge River ash.

The creek flows down the southern margin of a scoriaceous dacite flow which floors the present valley. Because Bridge River ash, which should thickly blanket this flow is absent, the flow must be younger than the ash (2440 ± 140 years B.P.) and probably covers the ash vent. Much of the edifice of Plinth Mountain is probably postglacial and that of Meager Mountain may be as well.

9. Olivine Basalt: A sparsely porphyritic plagioclase and olivine basalt underlies part of the ridge separating Job and Mosaic creeks. Flat-lying to southeasterly dipping flows parallel the present topography. On the northwest side of the ridge, basalt scoria and bombs comprise a breccia which overlies the flows and till.

Basement of the Meager Creek Volcanic Complex

Metamorphic and plutonic rocks form a highly irregular base to the complex with up to 1200 m of relief. Four units ranging in age from Mesozoic to Miocene are described in order of decreasing age.

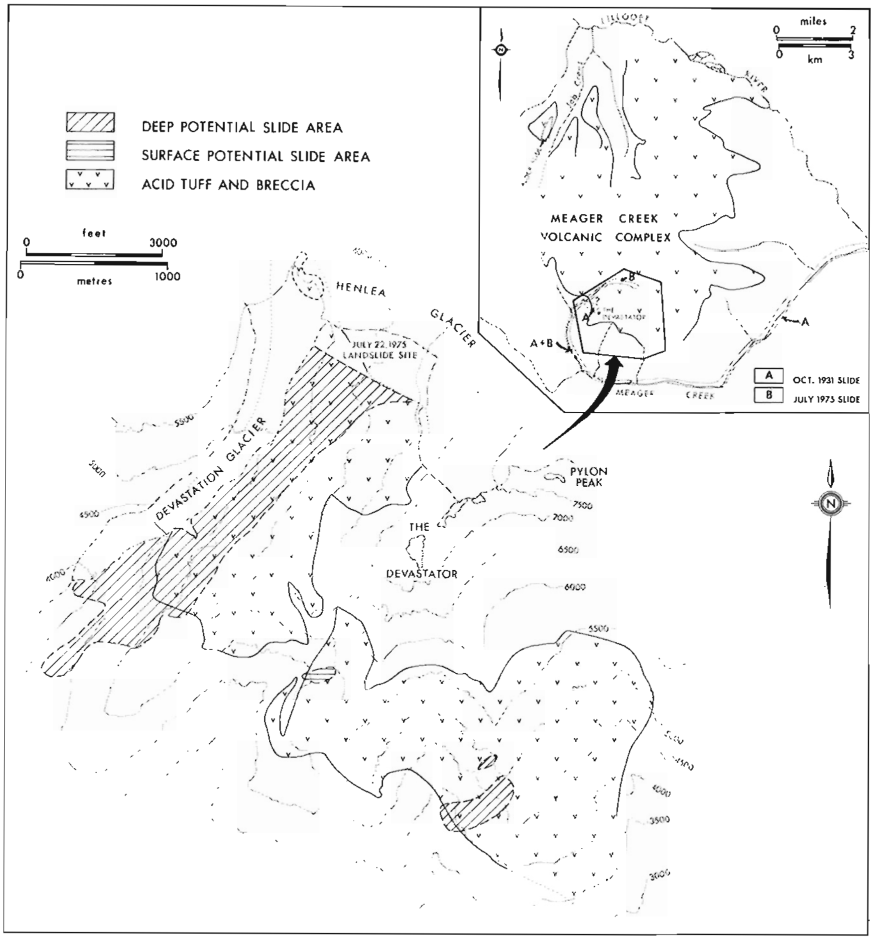

Figure 57. 3. Landslide potential of the Meager Creek Volcanic Complex.

1. Amphibolite, Grey Phyllite, and Marble: Mainly amphibolite, some grey phyllite, and rare marble underlie the northern part of the complex. Plutonic sills and dykes extensively intrude the strongly deformed and steeply dipping succession. The metamorphic rocks are part of a large septum of Upper Triassic to Lower Cretaceous rocks which extends 75 km along the Lillooet River from Lillooet Lake (Roddick and Hutchison, 1973) to Lillooet Glacier (Roddick and Woodsworth, 1976).

2. Quartz Diorite, Diorite, and Quartz Monzonite: Biotite-bearing hornblende quartz diorite and diorite and biotite quartz monzonite underlie most of the complex. Sporadically developed foliation strikes west-northwesterly with a steep dip and parallels trends of the older metamorphic rocks.

3. Volcanic Breccia and Flows: Weakly deformed and metamorphosed flows and breccia underlie the west side of the complex. A basal breccia composed of plutonic, metamorphic and volcanic detritus ranges in thickness from a few tens to hundreds of metres. Dark grey to light grey-green aphanitic flows with intercalated volcanic breccias containing plutonic clasts comprise a sequence over 500 m thick. The rocks are unfoliated and metamorphosed in the subgreenschist facies. Because the sequence nonconformably overlies the previously described plutonic rocks and rocks of possible Miocene age intrude it, a Cretaceous or Early Tertiary age is likely.

4. Miocene(?) Stocks: From the east end of the complex, Fall Creek stock extends eastward to and beyond the limits of the map-area. The biotite (3%) leucoquartz monzonite stock has fine grained and pegmatite phases. Locally, sulphides including molybdenite fill closely spaced fractures as on the Fall Property.

From Job Glacier to beyond the western limit of the map-area is the yellow to ochre weathering quartz monzonite of Job Glacier stock. On the west side of Job Glacier, the medium grained (1 to 2 mm) hornblende biotite (8%) leucoquartz monzonite stock has closely spaced fractures filled with pyrite. Adjacent to the fractures, intense hydrothermal alteration produces a porous rock from the quartz monzonite, and elsewhere propylitic alteration is common.

Because these stocks are lithologically similar to Salal Creek stock of Miocene age and because Job Glacier stock is the youngest of the basement units, Job Glacier and Fall Creek stocks are probably of Miocene age.

Springs

Included within springs are hot springs which issue from basement, cold springs flowing from basement or the complex, and fossil spring areas found only in the complex (Figure 57.2). Of the several odorless cold springs found, most have flows in the 50 to 250 litres per minute range and deposit carbonate or hydrated iron oxide sinters. No new hot springs were discovered, but during calm, dull days a strong H_2S odor drifts up the valley of Affliction Creek. Because known hot springs in the area are nearly odorless (Nevin, pers. comm., 1976), the odor may indicate an undiscovered hot spring within the lower course of Affliction Creek or the immediately adjacent Lillooet River. Fossil spring areas are characterized by opaline sinter encrusting joint surfaces over areas less than a 100 m square.

Landslide Potential

Within the last 45 years, three slides of rock, mud, water and ice mixtures have flowed down Meager Creek (Patton, 1976). The earliest one in October 1931 originated on the west flank of The Devastator and flowed down the length of Meager valley (Carter, 1932) (Figure 57.3). Little is known of a slide which occurred in 1947. The recent slide of July 22, 1975, originated between the 5300 and 6600 foot levels on the southern edge of Henlea Glacier above its confluence with Devastation Glacier. Failure of a water-saturated slope of the acid tuff and breccia unit between the 5300 and 6000 foot levels removed support from the overlying basal breccia of the porphyritic plagiocalse andesite unit. Bedrock of these units plus ground moraine and ice flowed down Devastation Glacier, destabilized the base of slopes underlain by the acid tuff and breccia unit on the east side of Devastation Glacier, and caused massive slumping of this unit which was incorporated in the slide mass. The slide path was 7 km long. The acid tuff and breccia unit underlies most potential slide areas, of which the largest potential slide masses lie on the western flanks of The Devastator overlooking Devastation Glacier which is receding and removing its support from the base of the slopes. On the western valley wall at the head of the west fork of Job Creek is a potential block slide area.

Economic Geology

Because of the proximity of the Salal Creek deposit (Mo) and Fall Property to Job Glacier stock, this locally pyritized and hydrothermally altered stock of possible Miocene age requires additional investigation. The hydrothermally altered and pyritized acid tuff and breccia unit, known to be older than 4.2 m.y., may be an eruptive equivalent of the Miocene stocks and deserves further attention.

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58. GEOCHEMISTRY OF THERMAL WATERS IN THE MOUNT MEAGER HOTSPRINGS AREA, BRITISH COLUMBIA

Project 730067

L. T. Hammerstrom[1] and T. H. Brown[1]
Regional and Economic Geology Division

Introduction

The Mount Meager hotsprings area, a potential geothermal energy source, has been the object of separate studies for the past three years by the Department of Energy, Mines and Resources (Souther, in press) and by the British Columbia Hydro and Power Authority (Nevin *et al.*). Temperature gradients, water flow, and rock types have been determined from six shallow exploratory wells (maximum depth of 347 m) drilled in the Meager Creek valley. Substantial geophysical work has also been undertaken in an attempt to construct a physical model of the subterranean reservoir.

This project is concerned with the chemistry of the thermal spring waters and surficial fresh waters of the Mount Meager area. Analyses of the waters will provide the data necessary to construct thermodynamic models of the possible reservoir conditions using basic mass transfer principles of water-rock interactions. These models will be applied to determine if alteration products observed in the well cores are the result of the thermal waters reacting with the host rock. In addition, attempts will be made to estimate the maximum water temperature within the reservoir using existing or new geothermometer principles.

The Mount Meager hotsprings area is 160 km north of Vancouver, British Columbia, near the headwaters of the Lillooet River (Fig. 58.1). The hotsprings outflow ranges from seeps in glacio-fluvial deposits and fractured bedrock to substantial artesian flows from the 4 inch drill stems. The springs are found at an approximate elevation of 730 m a. s. l. along the banks of Meager Creek and the Lillooet River.

Access to the area is usually by helicopter either from a base at Alta Lake, British Columbia approximately 60 km to the southeast, or from a rough helipad on the

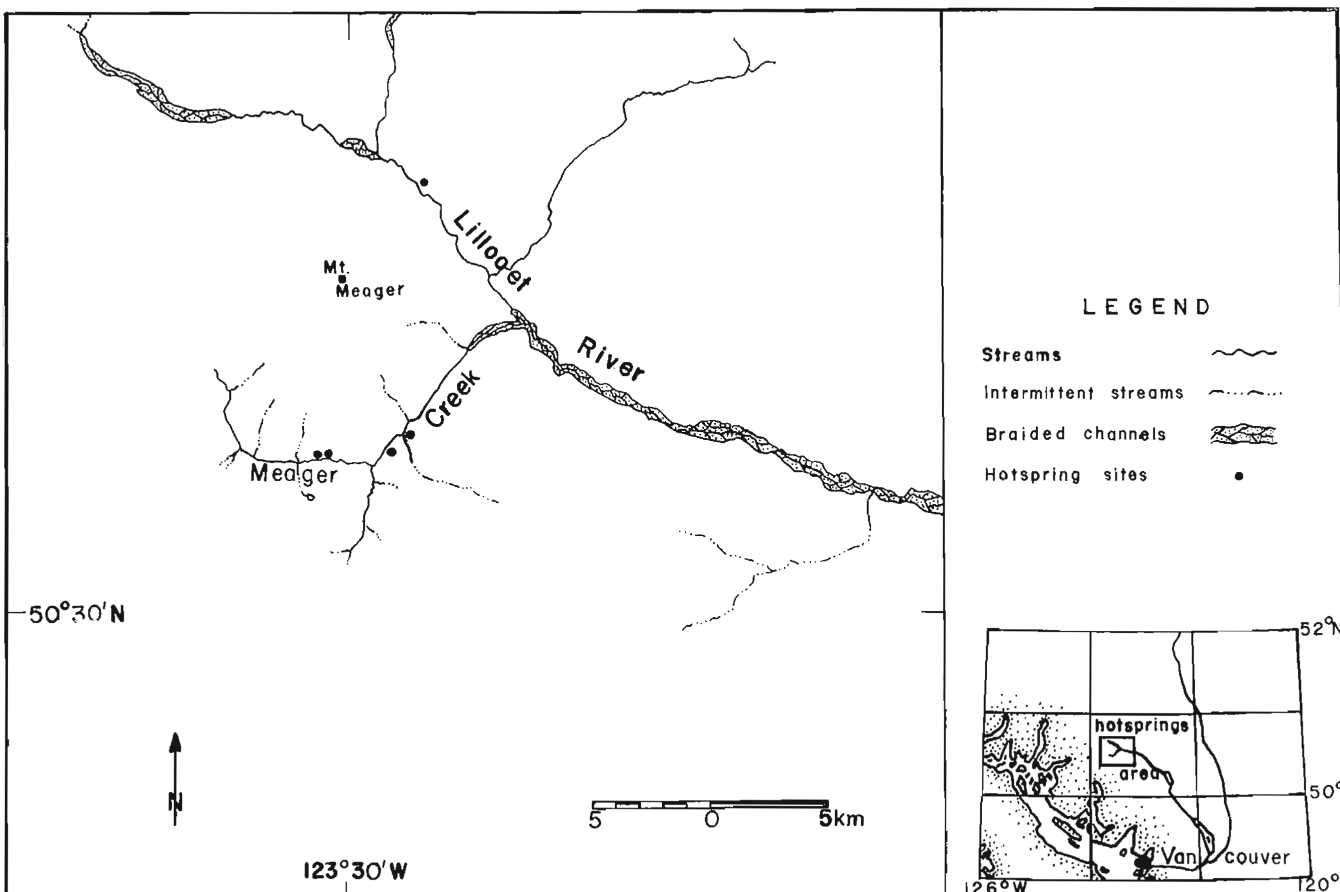

Figure 58. 1. Hotsprings locations (•) in the Mount Meager area, British Columbia.

[1]Department of Geological Sciences, University of British Columbia, Vancouver, British Columbia

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Geol. Surv. Can., Paper 77-1A (1977).

Table 58.1

Analytical results for hotspring waters in ppm.

Site	Location	pH	T°C	SiO_2("reactive")	Cl^-	SO_4^{-2}	Na^+	K^+	Ca^{+2}	Mg^{+2}
01	Meager	6.50	31.4	56.0	133	12.5	165	23.7	19.0	15.4
03	Meager	6.80	30.0	54.0	295	12.5	248	27.0	19.5	17.1
05	Meager	6.40	48.0	80.5	428	28.8	347	44.0	5.6	24.8
06	Meager	6.05	56.0	92.0	466	40.0	377	46.2	13.0	34.1
07	Lillooet	7.70	53.5	40.0	72	45.6	410	13.8	6.0	6.6
08	Lillooet	6.85	59.0	43.0	67	45.6	396	18.2	5.4	6.1

southwest bank of the Lillooet River 20 km downstream from the hotsprings area. Access from the nearest private logging road requires a 10 km hike into either the Meager Creek or Lillooet River hotsprings.

The initial reconnaissance and sampling trip into the Meager Creek area occurred July 30 — August 3, 1976. Three hot springs, one drill site, and two freshwater streams were sampled. A second trip was undertaken from September 1 — September 5, 1976, to the hotspring site on the Lillooet River (Fig. 58.1). Samples were taken from two hotsprings and two freshwater streams. Tentative plans call for continued sampling of the Meager Creek area on a monthly basis throughout the winter of 1976-1977, to determine if temporal changes occur in the hotsprings' chemistry.

Sampling Procedure

The sampling procedure used in this study was developed by Presser and Barnes (1974) of the U.S. Geological Survey and requires 3.4 litres of sample water per site. The sample water is stored in polyethylene bottles which have been cleaned previously with a 10 per cent nitric acid solution.

At each site water temperature and pH were recorded. The pH was determined with a meter capable of temperature compensation. On the first trip a portable Hach model 1975 pH meter was used. On the second trip an Orion model 407A specific ion meter was used and this meter will be used for all subsequent analyses. Both meters used an Orion model 90-01 reference electrode and an Orion model 91-01-00 pH electrode.

The water sampling involves initial filtration with Whatman no. 4 filter papers using a hand operated vacuum pump to provide the suction. Three one litre bottles are collected. Two are acidified to approximately pH 2 by the addition of 10 ml of concentrated nitric acid, and the remaining one litre is left untreated. Three 125 ml bottles are filled with 100 ml of sample. One is acidified by the addition of concentrated hydrochloric acid to approximately pH 2 and two are duplicates for sulfide determination. For silica determination a fourth 125 ml bottle is filled with 90 ml of distilled water and 10 ml of sample, which has been filtered through a 0.01 micron Sartorius membrane filter to remove the colloidal silicic acid. For the winter sampling period the total volume of water will be decreased to 2.1 litres because of weight and volume considerations during transport.

Preliminary Results

Analyses have been completed for "reactive" silica, chloride, sulphate, sodium, potassium, calcium, and magnesium ions. The results in parts per million (ppm) for the six hotspring sites are shown in Table 58.1.

The "reactive" silica and sulphate were determined by spectrophotometric analyses, the chloride by a chloride specific ion electrode, and the four major cations by atomic absorption methods. Other species to be analyzed include sulphide, total carbonate, bicarbonate, aluminum, lithium, strontium, trace metals (Mn, Fe, Cu, Pb, Zn), iodide, bromide, fluoride, phosphate, nitrate, and total silica.

From the concentrations given in Table 58.1 there seem to be significant differences between each hotspring site. When the data is converted from parts per million to molalities and with the use of activity coefficients, to activities, it becomes more thermodynamically useful. The activities for the cations can be placed in a ratio with the pH to represent the chemical potential of the oxide components. For example, the chemical potential of the component MgO can be represented as $\ln a_{Mg^{+2}}/a^2_{H^+}$. Tentative results of the activity ratios indicate little variation among the Meager Creek hotspring sites, which suggests they all could originate from the same thermal reservoir system. However, comparing Meager Creek hotsprings' activity ratios to the Lillooet River hotsprings' activity ratios indicates a significant difference in the chemistry of the two hotsprings areas. This could mean that the Lillooet River hotspring, sites may originate from a different thermal reservoir system than the Meager Creek hotsprings. The Lillooet River hotsprings are far from electrical neutrality when the difference between the total molalities of anions and cations for each site are considered. However, the Meager Creek hotsprings, sites are nearly neutral, which indicates that the major solution components have been accounted for. By inspection of the data the Meager Creek hotsprings appear to be a predominantly NaCl solution, but the Lillooet River hotsprings have the highest Na^+ molality and the lowest Cl^- molality of all the hotsprings. The only common anion not presently analyzed for is bicarbonate which will probably be the major anion at the Lillooet River hotsprings. A large amount of bicarbonate at the Lillooet River hotsprings would satisfy the electrical neutrality of the solution and it would provide an anion for the excessive sodium cation. Therefore the Meager Creek hotsprings area may be part of a thermal reservoir system with a characteristic NaCl solution and the Lillooet River hotsprings may be part of a separate thermal reservoir system with a characteristic Na_2CO_3 solution. It must be stressed that this is a tentative conclusion and may not be borne out by further analyses.

Future work will consist of monthly sample collection, chemical analyses, and data processing. It is hoped that this work and the resultant thermodynamic models may lead to a better understanding of the Mount Meager thermal system.

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METCHOSIN VOLCANICS AND SOOKE INTRUSIONS OF SOUTHERN VANCOUVER ISLAND

Project 730036

J.E. Muller
Regional and Economic Geology Division, Vancouver

Introduction

The writer spent about 6 weeks investigating the Eocene Metchosin Volcanics and Sooke Intrusions that underlie southernmost Vancouver Island, south of Leech River Fault (Fig. 59.1). The geology and petrology of these rocks was studied earlier in considerable detail by Clapp and Cooke (1917) and the geology of related copper deposits by Fyles (1949) and Stevenson (1951). More recently the geology of the Crescent Formation of Olympic Peninsula, correlative with the Metchosin Volcanics, on the opposite side of Strait of Juan de Fuca, has been studied by many workers (*see* Cady, 1975) and its chemistry by Glassley (1974). Metchosin Volcanics and Sooke Intrusions largely determine the geophysical properties of Southern Vancouver Island. Geophysical work recently undertaken in the Strait of Juan de Fuca (McLeod *et al.*, in press) and work in progress by the Geological Survey and the Earth Physics Branch demanded an updating and expansion of detailed knowledge of these rocks. This report combines preliminary results of fieldwork with existing data and attempts some general conclusions on the origin of the formations.

Structural Setting

Most of Vancouver Island is underlain by eugeosynclinal sequences of volcanic and plutonic rocks with minor sediments of Paleozoic and Mesozoic age. These "Island Mountains" form part of the Insular Belt of the Canadian Cordillera that continues northwestward in Queen Charlotte Islands and Wrangell Mountains of Alaska. South of Leech River Fault Metchosin Volcanics and Sooke Intrusions of Eocene (and older ?) age are overlain by minor Neogene clastic sediments. They are geologically part of what might be called the Olympic Mountains block. With one possible exception at Point of Arches on the northwest Washington coast these are not known to rest on older continental rocks and are held to be Eocene oceanic floor. A narrow strip of low to medium grade metasedimentary schist, Leech River Formation, is interposed to the north of Leech River Fault between Island Mountains rocks and Olympic Mountain rocks. It is considered to be the remnant of Late Jurassic to Cretaceous flysch, deposited on the slope and in the trench off the late Mesozoic continental edge.

Metchosin Volcanics

Eocene Metchosin Volcanics are similar in lithological composition and sequence to Upper Triassic Karmutsen volcanics of Vancouver Island. For the latter the sequence originally established by Carlisle (*see* Muller *et al.*, 1974) includes a lower part of pillow lavas about 3000 m thick, a middle part of broken-pillow breccia and well-bedded aquagene tuff about 1000 m thick, and an upper part of basalt flows more than 3000 m thick. It is inferred that the pillows represent deeper submarine extrusion and the breccias and tuffs shallow submarine outflow, followed by explosive disintegration under low water pressure aided by wave turbulence. The flows are presumed to be subaerial extrusions. A similar progression from deep-water pillow lavas to subaerial flows is apparent in Metchosin Volcanics.

Pillow lavas, pillow breccias and aquagene tuffs form the lower part of the formation (Fig. 59.2, to 59.4). They are well exposed in a central east-west belt of the mapped area, from the village of Metchosin to Muir Creek, in a generally homoclinal, northeast dipping sequence. The thickness cannot be determined accurately due to many cross faults and lack of stratigraphic markers, but is estimated to be in the order of 3000 m (9000 feet). Pillows vary in size within single outcrops from irregular spheres 10 to 20 cm in diameter to ellipsoidal masses 50 cm to 1 m in greatest diameter. In some outcrops pillows, more properly described as lava tongues, vary from a minimum 50 cm to several metres in width and up to about 1 m in height. Upper surfaces of pillows are highly convex but lower surfaces tend to be flatter with downward protuberances that have sagged into open spaces between underlying pillows. Phenocrysts of plagioclase and pyroxene and amygdules filled mainly with quartz and epidote are common (Fig. 59.5). The latter are in many instances oriented perpendicular to the pillow rims and tend to be larger and more numerous in the pillow tops. These various features are aids in ascertaining the general attitude of pillow lavas in the field, but the determination of dips and strikes is only possible by averaging many sample-measurements and may be subject to errors of about 15 degrees. The pillow lavas are associated with pillow breccias and aquagene tuffs. Breccias and tuffs do not appear to form a stratigraphically defined unit in Metchosin Volcanics as they do in the Karmutsen Formation, but are interbedded with pillow lavas and layered tuffs. The breccias are composed of small pillows and angular pillow fragments, some of which are oblong with parallel arrangement indicating the bedding of the rock. Tuffs are fine grained, sandy, silty or cherty and either massive with hardly any indication of layering or distinctly laminated. According to Clapp and Cooke (1917) they are composed of angular fragments of labradorite and augite in a matrix of altered material, mainly chlorite, epidote, calcite and limonite. In the finer grained tuffs the feldspar is sodic, probably a result of albitization by seawater.

Layered, commonly amygdaloidal basalt flows form the upper part of Metchosin Volcanics and are exposed in the northeastern part of the area in a narrow belt

From: Report of Activities, Part A;
Geol. Surv. Can., Paper 77-1A (1977).

about 2 km wide, between the pillow lava unit and Leech River Fault. They overlie the pillow lavas, are inclined about 20 degrees to the northwest, and have a minimum thickness of about 1000 m (3000 feet). Layered amygdaloidal lavas also underlie a belt between Kirby Creek and the coast from Sheringham Point to River Jordan. These strike almost perpendicular to the coast and exhibit mainly moderate to steep northwest dips. At present it is impossible to determine if their apparent thickness of many thousands of metres is real or due to repetition by faulting and folding. It is tentatively assumed that they are in part correlative to the flows south of Leech River Fault. That assumption is mainly based on the theoretical argument that the Metchosin sequence was initiated on the sea floor and that flows were only formed in a later stage of subaerial eruption. Individual lava flows are commonly not distinct but in places tops and bottoms of flows can be recognized in concentrations of amygdules and on bulbous flow tops that have been exhumed to form dip slopes. Most flows are 50 cm to 5 m thick, but thinner ones are also present. Like the pillow lavas the flows are composed mainly of calcic plagioclase and diopsidic pyroxene with accessory magnetite, apatite, and alteration products.

Basaltic dykes cut all Metchosin Volcanics but are most numerous in the lower part of pillows, breccia and tuff. This upward decreasing number of dykes is predictable as they were the feeders of successive extrusive layers. They strike generally westerly to northwesterly like the intruded and overlying extrusive rocks, but some strike northerly across the general trend. In exposures at Sooke Bay multiple dykes range from coarse to fine grained diabasic gabbro, through porphyritic basalt with feldspar phenocrysts a few millimetres in diameter, to fine grained basalt. Chilled

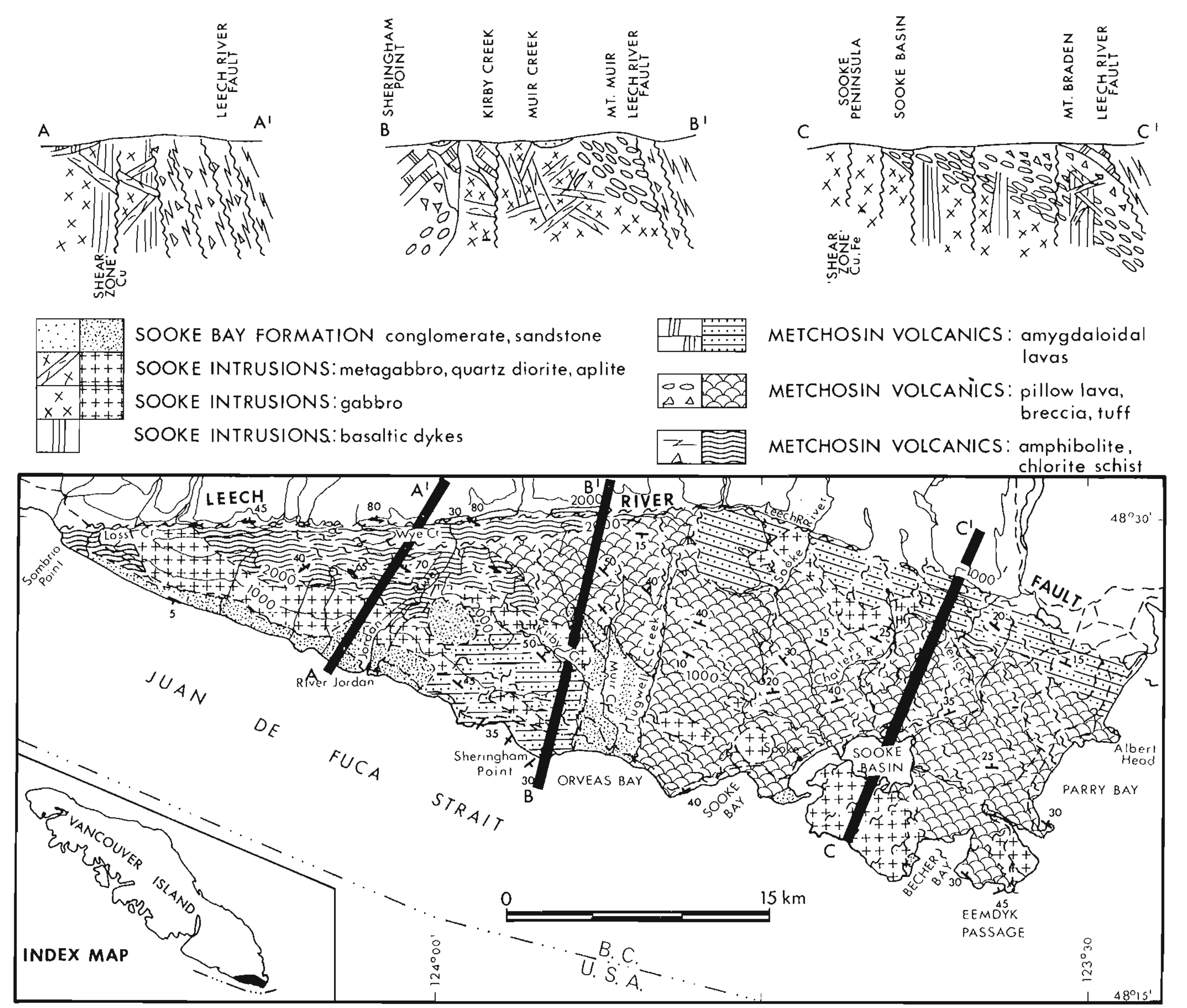

Figure 59.1. (Map and sections). Metchosin Volcanics and Sooke Intrusions.

margins commonly show age relationships. Ten chemical analyses of random samples of Metchosin Volcanics have been made by S. Courville of the Geological Survey. According to the Baragar - Irvine classificaion eight of these are tholeiitic basalt, one is alkali basalt, and one is hawaiite. Glassley (1974) also found mainly tholeiite and minor alkali basalt in Crescent volcanic rocks.

Metchosin Volcanics are metamorphosed in the entire area west of Jordan River and in a smaller area east of that river and south of Leech River Fault. They have been shear folded, converted to chlorite schists and are in places mylonitized. It is possible that these rocks are southwestward overturned limbs of folds involving mainly thinly bedded tuffaceous rocks. Lithologically they resemble shear folded volcanics of the Paleozoic Sicker Group. In the area west of Jordan River and near intrusions east of that river the volcanics are generaly amphibolized. These rocks are dark, massive to more or less foliated microdiorites that have lost most of their primary structures, though relict amygdules, flow and pillow structures are locally preserved.

Sooke Intrusions

The Sooke Intrusions, as defined by Clapp and Cooke (1917), are mainly gabbro, intrusive into Metchosin Volcanics. They contain minor stocks of quartz diorite and tonalite (or trondhjemite) and apophysal phases of hornblendite and aplite. The exposed intrusions are all west-northwestwardly

Figure 59. 2.

Pillow breccia. Fragments show amygdules concentrated towards pillow rims.

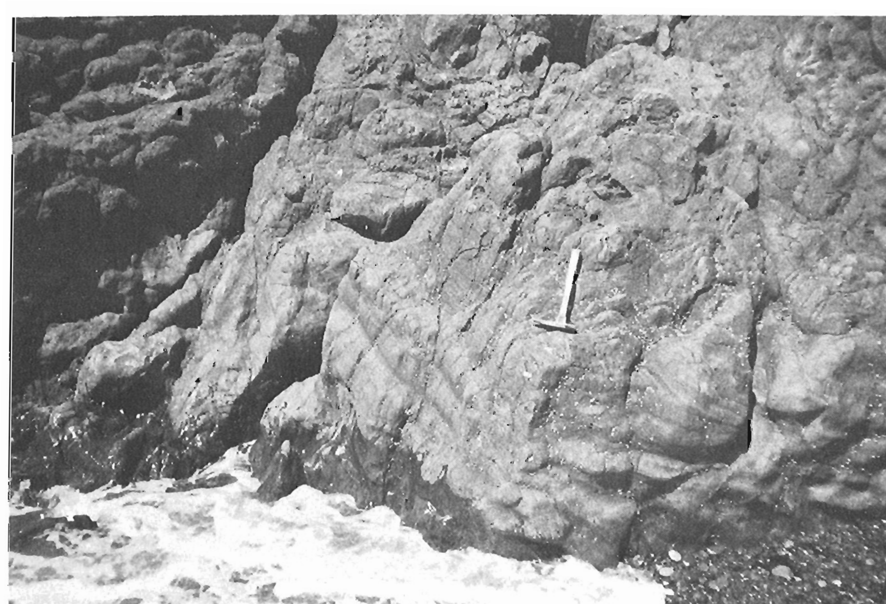

Figure 59. 3.

Tuff overlain by breccia.

elongated bodies with maximum width about 4 km. The body underlying most of East Sooke Peninsula is composed almost entirely of medium to coarse grained gabbro, apparently the deepest level of erosion. The bodies near Leech River Fault reveal the largest masses of quartz diorite and tonalite and represent high-level parts of the plutons. The gabbro of East Sooke was studied in much detail by Cooke (1919 and in Clapp and Cooke, 1917) and recently by Mitchell (1973). Modal analyses show 65 to 45 per cent plagioclase, 20 to 40 per cent pyroxene and 0 to 12 per cent olivine in medium to coarse (1-5 mm) grain sizes. Cooke (1919) reported olivine gabbro with up to 25 per cent olivine and also bytownite-anorthosite with less than 5 per cent mafic minerals (Fig. 59.6) Contacts between these different rock types are gradational or may be intrusive. Hornblende gabbro and hornblendite form late-phase dykes and dykelets. They were injected after consolidation and fracturing of the main intrusions and are composed mainly of hornblende crystals, to several cm long, with subparallel orientation perpendicular to the dyke walls, and of a lesser amount of calcic plagioclase.

Part of the Sooke gabbro is gneissic plagioclase amphibolite. Clapp and Cooke (1917) mention gneissic gabbros showing "flowtextures" with parallel pyroxenes and plagioclase from Sooke Peninsula, commonly at the margin of the intrusion. Gneissic gabbro and amphibolite exposed in several places west of Jordan River are finely foliated rocks with alternating laminae and

Figure 59.4.

Pillow lava.

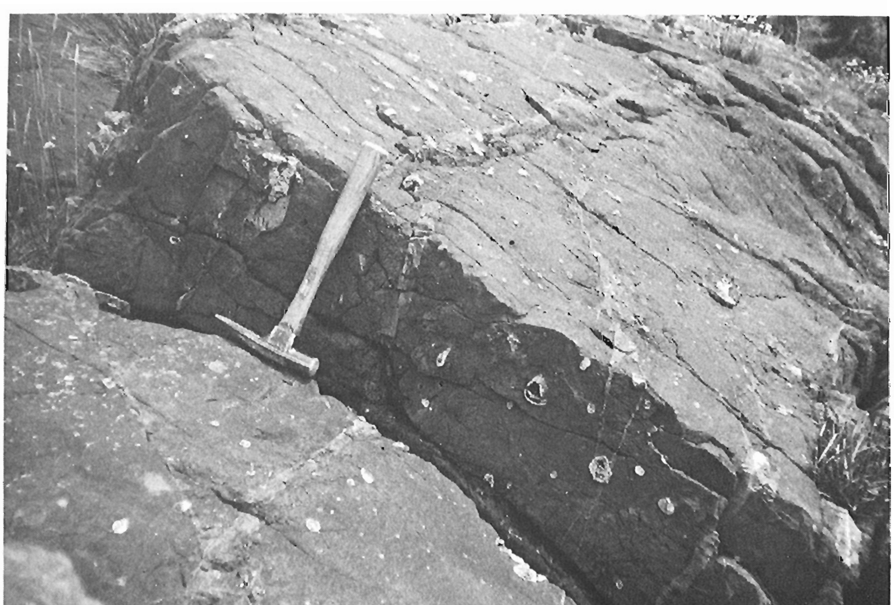

Figure 59.5.

Top and face of basalt flow with large quartz amygdules.

lenticles, no more than a few millimetres thick, of hornblende and plagioclase and appear to have formed by recrystallization coupled with shearing of gabbroic rock (Fig. 59.7).

A small proportion of Sooke Intrusions is composed of quartz diorite and tonalite, commonly dykes and dykelets, ranging in width from less than 1 cm to about 30 m. The larger ones are part of the intrusive bodies near Leech River Fault whereas only narrow felsic dykes, up to a few metres thick, have been seen in the southern intrusions. The felsic material fills single fractures or shatter zones in gabbro and amphibolite and locally forms bodies of agmatite, as in the intrusion between Kirby and Muir creeks (Fig. 59.8). These intrusive breccias show everywhere sharp contacts between angular fragments of mafic rock and the felsic, commonly fine grained matrix. According to petrographic data of Carson (1973) the felsic rocks are composed of 55 to 75 per cent sodic platioclase, 10-30 per cent quartz and about 5-25 per cent mafic minerals, mainly hornblende and/or biotite. They are thus quartz diorite or tonalite (20% quartz) or, if that distinction is made, trondhjemite (with albite or oligoclase).

Mode of Origin

Clapp and Cooke (1917) realized that the Metchosin pillow lavas were submarine. They attributed an explosive origin to tuffs and breccias but considered also fragmentation of volcanics by wave action. It is

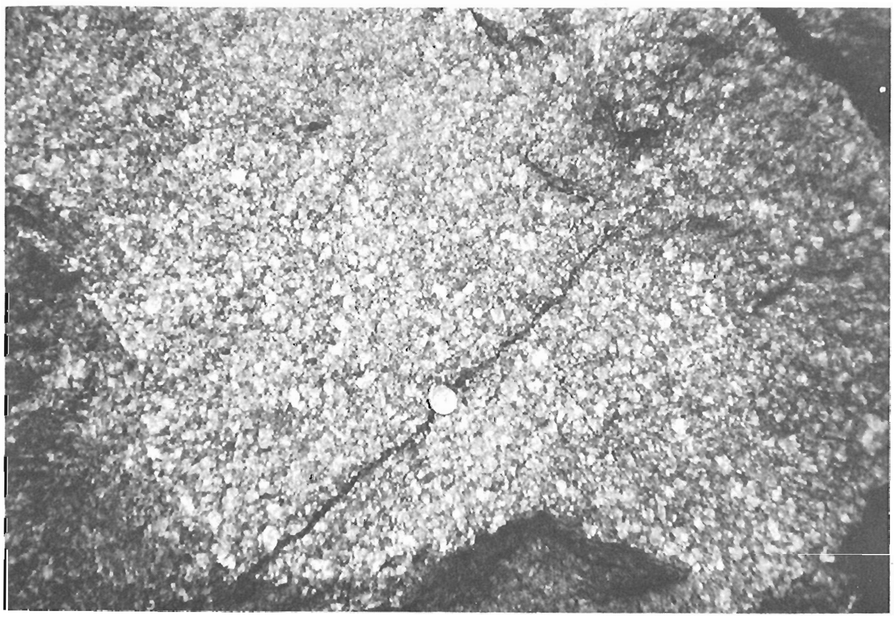

Figure 59.6.

Anorthosite.

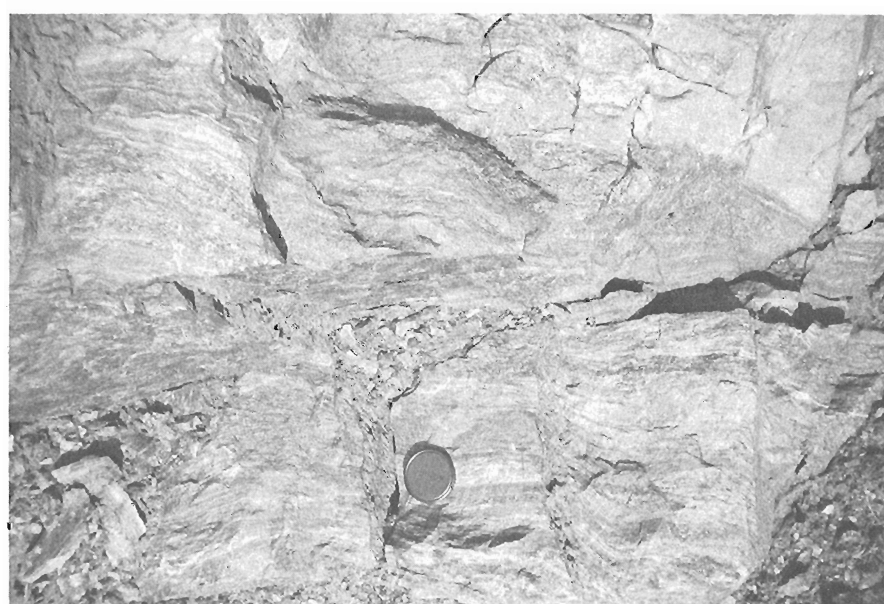

Figure 59.7.

Gneissic plagioclase amphibolite (K/Ar age 46.9 ± 7 m.y.).

now well established that submarine basaltic extrusions commonly form pillows, or under certain conditions (low water pressure and wave action ?) may be shattered into pillow breccia and fine grained aquagene tuff. The thick, in part coarsely amygdaloidal lava flows of the upper Metchosin Volcanics were almost certainly erupted subaerially.

The Sooke Intrusions may be viewed as elongate magma chambers that extended upward into dyke complexes that in their turn were feeders of the volcanics. As younger volcanic layers were extruded the intrusions (magma chambers and fissures) extended upward in the lower parts of the volcanic sequence. Thus, though intrusions and volcanics are essentially of equal age they exhibit intrusive relationships. Of special interest is the relationship of late felsic and hornblendic intrusions to the main gabbroic and basaltic rocks. These relations are displayed to some extent in the canyon of Sooke River, about one km south of Leech River Fault. There coarse grained uralitized gabbro intrudes fine grained volcanics, present in north trending entirely recrystallized septa. Hornblende blades in the amphibolized gabbro are up to 3 cm long. The coarse grained gabbroic rock is intruded by fine grained epidote bearing tonalite aplite dykes which are a few centimetres to 50 cm thick and trend northwest. Youngest are basaltic dykes, cutting all older rocks, trending northerly and distinct from the older dark coloured septa by their content of feldspar phenocrysts. The classical mechanism of rest magma formation, favoured by Clapp and Cooke (1917) is by fractional crystallization and subsequent intrusion of the felsic rest fluid into the surrounding rocks. According to newer petrological concepts granotoid rocks are commonly formed by migmatization and partial melting of existing rocks. Fusion differentiates of basic rocks forming the light material (leucosome) of agmatites are commonly quartz diorite or tonalite. Such agmatite is common in the Mesozoic Westcoast Complex of Vancouver Island. There great volumes of rock are involved and contacts between light and dark material vary from sharp to diffuse, whereas Sooke agmatites are of small volume and consistently show sharp contacts. It does not seem that at present a definite choice can be made between the two processes of granitization in regard to Sooke Intrusions and possibly both were involved to some degree.

In the light of modern oceanic tectonics the volcanics and intrusions apparently represent newly formed ocean floor or, in the case of the lava flows, part of an Iceland-type spreading ridge. It is therefore of interest to compare them with examples of ophiolites that are considered to be oceanic crust, raised above sea level by upthrusting or obduction. Coleman (1971) compiled many such sections from the Mediterranean area, the western Pacific Ocean, and California. Three oceanic layers, established from seismic investigations of the oceanic crust, are recognized in the sections. Layer 1 consists of oceanic sediments and is missing on the Metchosin Volcanics, presumably because they were above sea level during their final stage of extrusion. Oceanic layer 2 of basalt averages 1.4 km in thickness Metchosin Volcanics, though similar in structure and petrology to oceanic basalts, are estimated to be 4 km thick (3 km of pillows and 1 km of flows), and hence are of the same order of magnitude. Oceanic layer 3, 4.7 km thick, of gabbro, diabase and trondhjemite (i.e. leucotonalite) corresponds well with Sooke Intrusions which have the same petrology but are of unknown thickness. The base of oceanic layer 3 is the M discontinuity, underlain by peridotite. The comparison leads to the suggestion that the peridotite layer is or was close to the surface under southern Vancouver Island. This possibliity could perhaps explain a known gravity high but to the writer's knowledge has not yet been explored by geophysicists. In general the volcanic-plutonic complex has indeed the characteristics of oceanic crust formed during Eocene time over a spreading ridge that may have been close to the continental margin like the present Juan de Fuca Ridge.

Age

The age of the volcanics is based on one fossil locality only, located in the bay west of Albert Head where, in a sequence of volcaniclastic sediments overlying pillow lavas, one bed of greywacke about 1 m thick contains a great number of thin, spiral shaped gastropods and a few pelecypods. W.O. Addicott of the United States Geological Survey has determined a small collection from the locality to be mainly *Turritella* cf. *T. uvusana hendoni* Turner indicating probable early Eocene age (P.D. Snavely pers. comm., 1971). The earlier late Eocene age assignment by C.E. Weaver and W.H. Hall (Clapp and Cooke, 1971) is apparently to be revised. The age of the correlative Crescent Formation of the Olympic Peninsula is likewise early (to early middle ?) Eocene. The *Turritella's* and nearby limestone with algal and foraminiferal remains suggest depositon in shallow water, in good agreement with the apparent stratigraphic positon between submarine pillow lavas and subaerial lava flows.

A few published and unpublished K/Ar age determinations by R.K. Wanless of the Geological Survey of Canada are available on Sooke Intrusions. Biotite from trondhjemite east of Jordan River (GSC 65-13, paper 66-17) yielded 39 ± 10 m.y. Hornblende from Jordan River Mine ore (GSC 70-36, paper 71-2) was dated at 44 ± 6 m.y. and hornblende from mineralized hornblendite from East Sooke Peninsula (GSC 72-25, paper 73-2) was 31 ± 15 m.y. Finally, hornblende from fine grained hornblende plagioclase gneiss from the highway, 8 km west of River Jordan (unpublished) was dated 46.9 ± 7 m.y. In this group of ages those with largest reported errors are also the youngest and the average age can probably be taken to be about 45 m.y. or late Eocene. The dating may represent the time of cooling of rest magmas or hydrothermal fluids, intruded as quartz diorite — tonalite and hornblendite dykes and also forming local copper mineralization. Whether this could indeed have occurred several million years after extrusion of basalts and intrusion of the original unmetamorphosed Sooke gabbros is not yet clear.

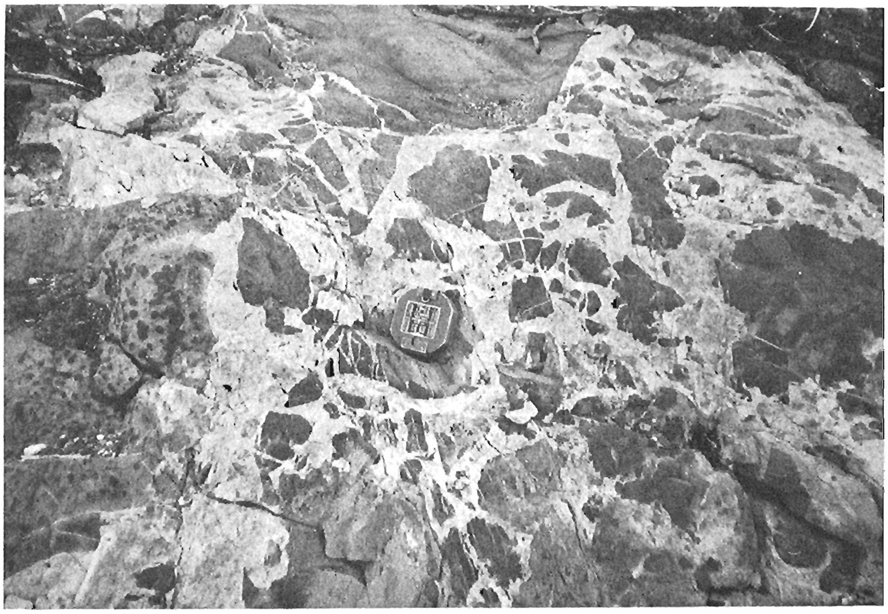

Figure 59.8.

Agmatite (amphibolite and quartz diorite).

Structure

The predominant structural elements of the area are steeply dipping to vertical faults. They are well marked on aerial photographs in the eastern part of the map-area and even though they are commonly marked by debris-filled valleys, their shear zones are exposed in many places. In the western part a thick cover of late Tertiary and Pleistocene deposits hides much of the structure but faults are exposed locally along the larger creeks. The most important Leech River Fault forms the northwest limit of Metchosin Volcanics along its entire distance. Its western half trends almost due east, its eastern half east-southeast. As noted this is a fundamental fault that separates two crustal domains and is marked by major gravity and magnetic discontinuities. Parallel to it is the "Pedder Bay Fault" from Pedder Bay to Sooke Basin. Many other faults are oblique to and terminate against Leech River Fault. One set of faults trending northwesterly may be conjugate to that fault. Several other faults trend northeasterly, e.g. "Tugwell Creek Fault", "Sooke River Fault" and a fault extending from Milnes Landing to Jack Lake. Possibly these faults slightly offset Leech River Fault and may be younger. Offsets in the submarine magnetic pattern suggest that these northeasterly faults extend into the Strait of Juan de Fuca (McLeod *et al.*, in press).

In the eastern part of the area blocks between faults generally are tilted 10 to 30 degrees northeastward toward Leech River Fault and strike roughly parallel to it. About 10 km from the fault, opposite southwesterly dips indicate a gentle fold. West of Tugwell Creek the structure is more complex and poorly understood, due to a veneer of late Tertiary and Pleistocene deposits that cover most of the volcanic-intrusive complex. Steep to vertical dips are common and shear folded volcanics suggest overturning of thick sections. Here also the rocks strike northwesterly at an about 30 degree angle to Leech River Fault. In the coastal strip, steeply dipping northeast striking volcanics are exposed. It is the writer's impression that in this region Leech River Fault obliquely cross-cuts an already folded and faulted complex of volcanics and intrusions.

Nature and timing of Leech River Fault are yet ambiguous in detail. North of the fault the Leech River Formation is deformed in steeply north dipping isoclinal folds with horizontal axes parallel to the fault. South of the fault Metchosin Volcanics strike roughly parallel to it and are inclined about 30 degrees towards it in the area east of Tugwell Creek. Farther west the structure of Metchosin Volcanics strikes at considerable angle into the fault and appears unrelated, although deformation and shearing appears to be most intense near the fault. Thus the Leech River Formation appears to have been thrust southward over the volcanics in a manner consistent with an assumed north-dipping subduction zone in the eastern part. But in the western part the Fault cuts across structures of the Metchosin-Sooke complex in a manner, more compatible with the character of a transcurrent fault. Both near-horizontal and near-vertical dragfolds are found in the Metchosin Volcanics, suggesting that both types of movement occurred.

The time of faulting is bracketed by the early Eocene age of Metchosin Volcanics and K/Ar ages 36 to 41 m.y. (latest Eocene to early Oligocene) on biotite from several samples of Leech River Formation near the fault. If severe post-early Oligocene faulting had occurred, it would most likely have reset these metamorphic ages. Undeformed Miocene Sooke Formation strata lie with angular unconformity on sheared and faulted volcanics near the fault on the coast north of Sombrio Point. Yet the fault seems to have had some influence on later Tertiary sedimentation. Miocene (to Pliocene?) Sooke Formation is limited to the area south of the fault, whereas Eocene to Oligocene Carmanah sediments are present only north of the fault.

Copper Deposits

Economic deposits of copper associated with the Sooke gabbro are the recently abandoned Jordan River mine and the deposits of Sooke Peninsula that were mined 1915-1918 (Stevenson, 1951, Fyles, 1949). The deposits are chalcopyrite with pyrite, pyrrhotite and some molybdenite and occur in vertical fracture zones up to 30 m wide. In the Jordan River deposit the ore is in hornblendized basalt in contact with gabbro, on Sooke Peninsula it occurs in hornblendized gabbro. Part of the ore is also in feldspathized and scapolitized rock. The metasomatic and ore-forming solutions were apparently a late phase of volcanic-intrusive activity and have been dated, as noted, at 45 m.y. or late Eocene. The metal appears to have been extracted from gabbro, as it occurs in or adjacent to that rock. Gabbro is most likely present at some depth below all of the area south of Leech River Fault and additional deposits of copper may be present.

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Project 650016

P.B. Read[1]
Regional and Economic Geology Division

A month of field work in the west half of Lardeau map-area (82 K) was devoted to: (1) completion of a structural cross-section of the north end of the Kootenay Arc, and (b) investigation of the relationship between the Kootenay Arc and the Shuswap Metamorphic Complex.

Tracing metavolcanic rocks of the Jowett Formation (black in Fig. 60.1) permits subdivision of the Lardeau Group into the Broadview Formation which overlies the Jowett Formation and the Index Formation which underlies it. Within the area of detailed mapping, southeasterly trending folds, typical of the north end of the Arc, change to an easterly trend on the east side of the Columbia River and the north end of the Upper Arrow Lake.

Along the western edge of the map-area, a fault system juxtaposes the stratigraphy and structure of the Kootenay Arc on the east against those of the Shuswap Metamorphic Complex on the west (Fig. 60.2). In the

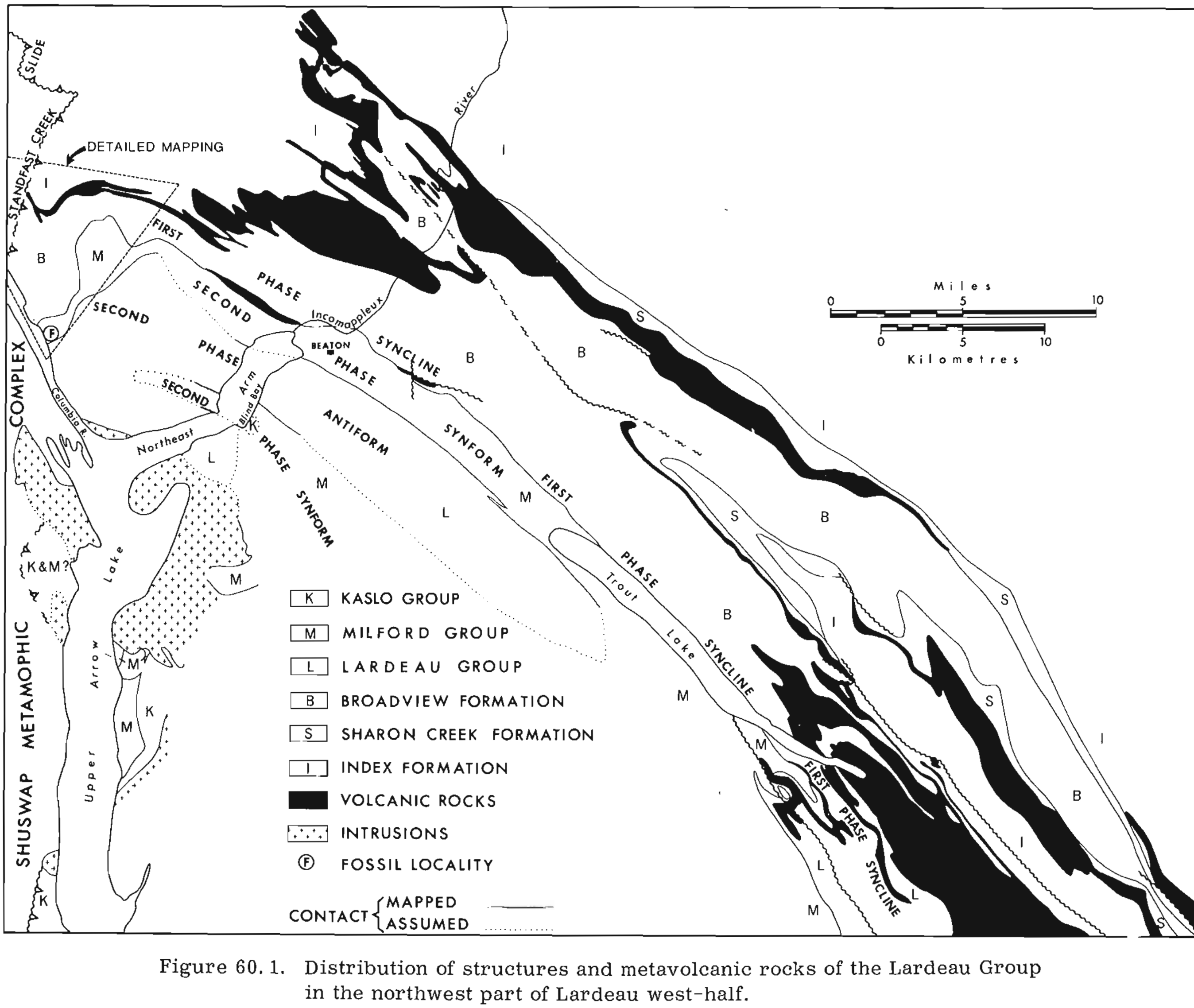

Figure 60.1. Distribution of structures and metavolcanic rocks of the Lardeau Group in the northwest part of Lardeau west-half.

[1]Consulting Geologist, Geotex Consultants Ltd., No. 1103-100 West Pender Street, Vancouver, B.C. V6B 1R8.

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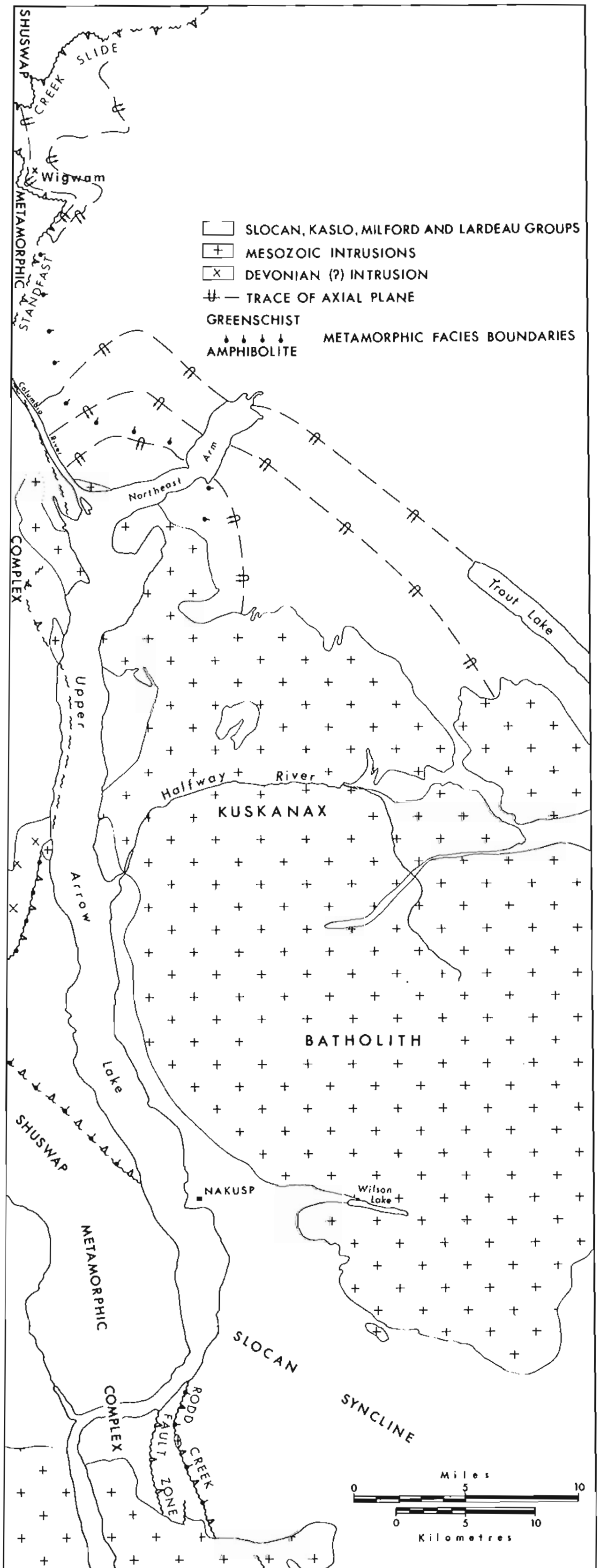

Figure 60. 2. The relationship between the Kootenay Arc and Shuswap metamorphic complex.

north, the Standfast Creek Slide dips gently southerly or easterly (Thompson, 1972), but in the south along the west side of Upper Arrow Lake the dip steepens to moderate easterly. In the southwestern corner of the map-area, Rodd Creek Fault System (Hyndman, 1968) dips 40 to 50 degrees easterly and crosses Upper Arrow Lake. Standfast Creek Slide and Rodd Creek Fault system may join to the west in Vernon map-area (Jones, 1959). The faults truncate structures of the Kootenay Arc and disrupt metamorphic isograds (Fig. 60.2). The normal fault movement probably is dominantly dip-slip but of unknown magnitude. Although metamorphic grade in the Kootenay Arc increases towards the Shuswap Metamorphic Complex, the truncation of second phase structures and later metamorphic isogrades restricts fault movement to the Middle Jurassic-Early Cretaceous interval. South of Nakusp and at the north end of Upper Arrow Lake, Upper Cretaceous or (?) older plutons intrude the fault system. South of the Wigwam Property, the system transects the economically important Badshot Limestone and prevents its reappearance farther south in the Kootenay Arc.

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NEPHRITE JADE OCCURRENCES IN BRITISH COLUMBIA AND YUKON TERRITORY

Project 490038

S.F. Leaming
Regional and Economic Geology Division, Vancouver

During the 1976 field season, three weeks were spent in gathering the latest information on the nephrite industry. The new information permits an upward revision by a factor of two or three, of the estimates of probable and possible reserves (Leaming, 1976) in the Cry Lake map-area and suggests that this area offers good possibilities for many more discoveries.

The area is rather remote (Fig. 61.1) and at present is accessible only by air. This may change in future as two important mineral deposits relatively close to the known nephrite occurrences may justify road construction from Dease Lake on the Steward-Cassiar Highway some 60 air miles to the west. These deposits are the Letain asbestos property north of Letain Lake, and Kutcho Creek copper-zinc deposit of Imperial Oil Ltd.

Cry Lake Jade Mines Limited hold claims and leases on lode and talus deposits in the mountains southeast of Letain Lake (Fig. 61.2, loc. 2). The property was visited on July 17. At that time two employees of the company were core-drilling talus blocks derived from a lode about 500 feet above the floor of the small cirque where most of the blocks had come to rest. The lode has a width of 25 feet and was exposed along a strike length of 60 feet. The full length could be several times this figure; calculations based on minimum dimensions resulted in an estimate of 4000 tons of untested nephrite; there may be two or three times this amount. Talus blocks below the lode aggregate about 200 tons. Because only a small part of the material had been tested by core drilling and no sawing had been done, the proved reserves of commercial nephrite are small.

The material seen in the lode and talus blocks appears sound and some good sections of drill core indicate that a large quantity of good quality nephrite should be available. The company intends to remove selected blocks by crawler tractor during the winter. It is difficult to obtain good hand specimens of nephrite from large boulders or smooth outcrops with a rock hammer, but a specimen was collected for further study. A sawn face on this sample revealed a dark green base with spots and flecks of bright emerald green uvarovite, and some spots and streaks of black magnetite/chromite. There is a weak grain to the rock, but it would qualify as commercial grade although not of the highest quality.

King Mountain Jade Mines Ltd. hold claims and leases adjoining the Cry Lake Jade Mines property on the east (Fig. 61.2, loc. 1). Four major lodes were seen on the property but at the time of the visit these were untested. A number of boulders of unknown quality were seen on a placer lease along the creek leading from the cirque below the lode on the Cry Lake Jade property. Estimates of reserves in the probable category based on measurements of minimum dimensions in the four lodes amount to 2000 tons. Possible reserves might be double this amount. Initial production from one of the placer leases amounted to 1500 pounds, most of which was shipped to the United States of America, the remainder was stockpiled in Dease Lake where it was examined on August 14. The material appears to be of high quality judging from a slab obtained from the company.

Nephro-Jade Ltd. continued to test alluvial deposits along the valley bottom in the vicinity of Provencher Lake, (Fig. 61.2, loc. 6) and expect to remove 200 tons during the winter.

Nephrite deposits in the Cry Lake map-area seem to be concentrated near the southeast end of an elliptical area of ultramafic bodies with a long axis extending from Kutcho Creek to Eagle River. The central and northwestern parts of this area contains some very large masses of serpentinized peridotite; to the southeast the serpentinite bodies are small and numerous.

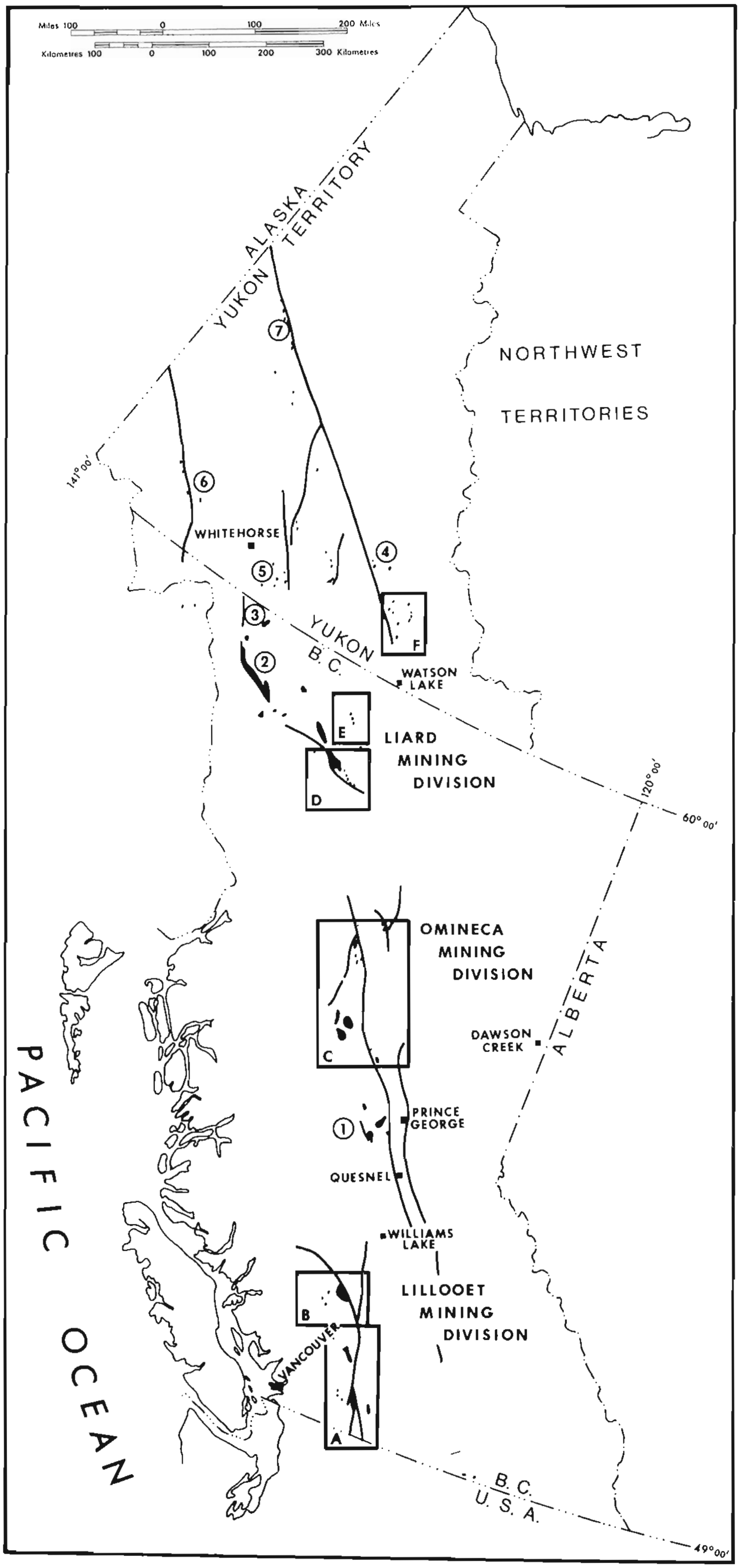
At least 12 separate lodes have been found in the mountains east of Provencher Lake (Fig. 61.2, loc. 1 to 5). In contrast, no nephrite lodes are known anywhere in the serpentinites northwest of Alice Shea Creek (Fig. 61.2, loc. 10). This distribution of nephrite may be more apparent than real as the whole area has not been exhaustively prospected. By analogy with other nephrite areas of British Columbia the apparent distribution may however have a good geological basis. In the Shulaps Range (*see* Fig. 61.1, area B) all the nephrite occurrences are associated with small bodies outside the main mass of the Shulaps ultramafic mass or along its margin. Similarly in the Mount Sidney Williams area, the Jade Queen deposit is associated with a small serpentinite satellite north of the main body.

All other known occurrences in British Columbia, and one in Yukon Territory are associated with small serpentinite bodies, of tabular shape less than a few hundred feet wide.

The reduction of a single large intrusive to a group of spatially related, progressively smaller bodies may be explained as a consequence of later faulting and remobilization. Thus, the initial, intrusive, elliptical mass of the Cry Lake body may have been emplaced in Devono-Mississippian time, with further redistribution along faults developed in Permo-Triassic, and Late Jurassic time. If the distribution of nephrite occurrences as shown on Figure 61.2 is real and not just apparent, then it follows that the nephrite developed in late Jurassic time when the proper environment of temperature, pressure and fluids were present. The theory is very tenuous and may be regarded as a working hypothesis which would have to be discarded on the discovery of nephrite northwest of Turnagain River.

Two days were spent in prospecting serpentinite occurrences near the head of Hoole River in the Finlayson Lake map-area (Geol. Surv. Can., Map 8-1960). This

From: Report of Activities, Part A;
Geol. Surv. Can., Paper 77-1A (1977).



PRELIMINARY ESTIMATES OF NEPHRITE RESERVES IN CANADIAN CORDILLERA

BRITISH COLUMBIA & YUKON

	AREA	SHORT TONS (all grades) Proved	Probable	Possible
A	FRASER	—	200	500
B	YALAKOM	10	200	2,000
C	OMINECA	50	2,000	10,000
D	CRY LAKE	100	20,000	50,000
E	CASSIAR		1,000	10,000
F	FRANCES	—	100	500
	TOTALS	160	23,500	73,000

ADDITIONAL POSSIBILITIES FROM ULTRAMAFIC BODIES WITH NO KNOWN OCCURRENCES

1	Prince George area South of Vanderhoof			300
2	Nahlin ultramafic body			700
3	Atlin			500
4	Campbell range			1,000
5	Teslin area			200
6	Kluane area			200
7	Dawson area			300
	TOTALS			3,200
	GRAND TOTAL	160	23,500	76,200
	Commercial grades	160	2,350	7,620

Inventory of Nephrite in storage at Vancouver, Chilliwack, & Lillooet probably 200 tons May 1, 1976.

LEGEND

- Fault
- Ultramafic bodies
- A — Areas with known occurrences
- 1 — Areas with no known occurrences but where discovery may be postulated from favourable geological indications.

Figure 61. 1.

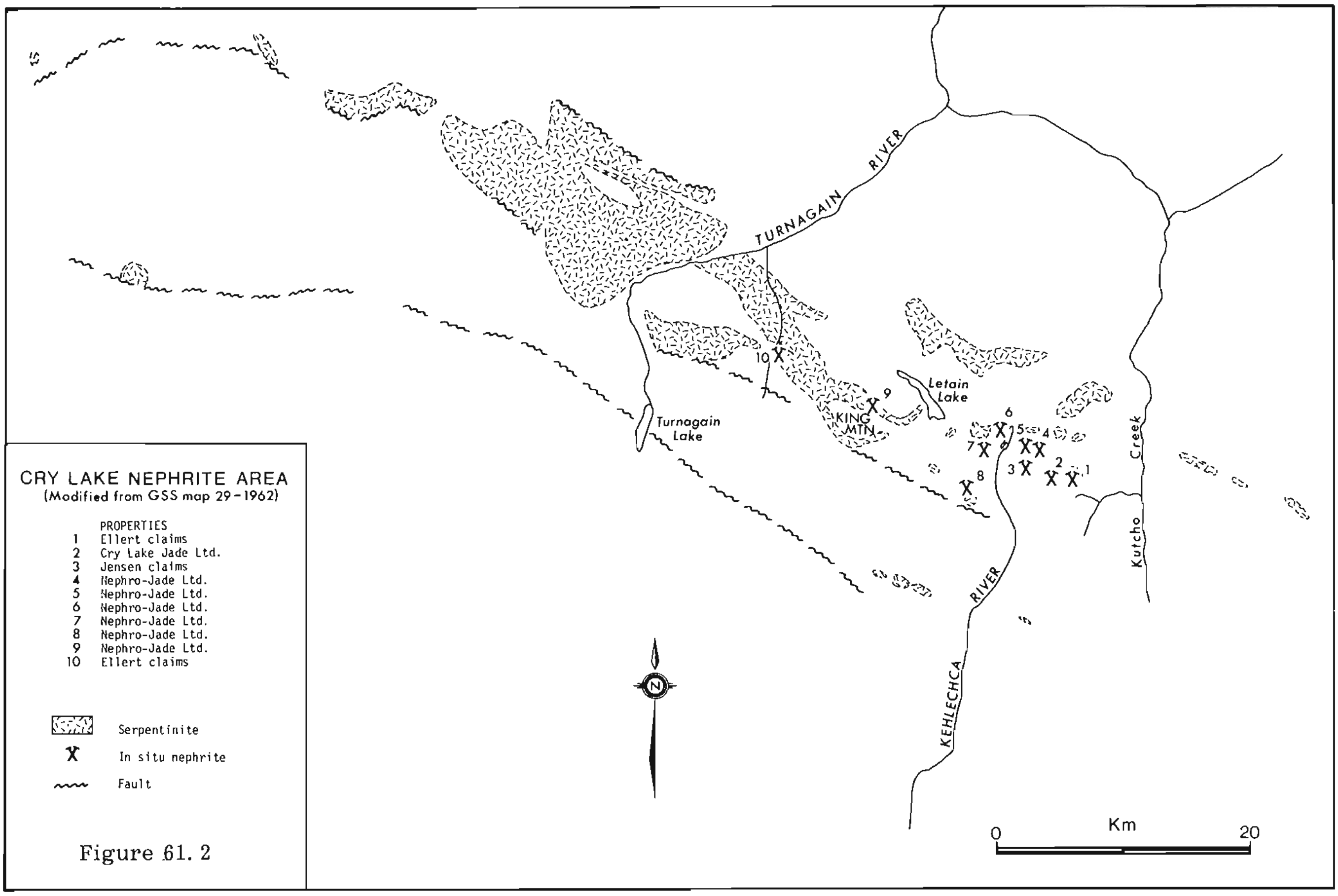


Figure 61. 2

area is a continuation of the geological regime in the Campbell Range where nephrite occurs associated with serpentinite in Devono-Mississippian rocks (Geol. Surv. Can., Map 6-1966). No nephrite was found but a few contact reaction zones with talc and calc-silicates were seen. The area therefore cannot be considered barren. A few other serpentine areas lie between Hoole River and the nephrite-bearing serpentines in the Campbell Range in the southwest corner of Frances Lake map-area. These should be investigated.

The estimates of reserves shown on Figure 61.1 are very speculative and particularly so in the possible category. Probable reserves are shown only for areas with known occurrences and are considered to be of the right order of magnitude. The percentage of commercial grades (10%) may be low because an increasing quantity of poorer grade material is being used in carving.

Reference

Leaming, S.F.
1976: Nephrite Jade Occurrences in British Columbia and Yukon Territory; *in* Report of Activities, Part A, Geol. Surv. Can., Paper 76-1A, p. 127-129.

62. SUBMERSIBLE OPERATION ON THE PACIFIC CONTINENTAL MARGIN

Project 710048

C. J. Yorath, D. L. Tiffin, and B. E. B. Cameron
Regional and Economic Geology Division, Vancouver

Introduction

In June, 1976, two of the authors (C. J. Y. and D. L. T.) participated in five submersible dives on the continental shelf and slope adjacent to the west coast of Vancouver Island, British Columbia (Fig. 62. 1). The submersible used was PISCES IV, owned and operated by the Department of Environment and located at the new Institute of Ocean Sciences at Patricia Bay on Vancouver Island. The support vessel employed was M. V. PANDORA II, a ship specifically modified to maintain, launch and recover this submersible.

The purpose of the program was to evaluate the submersible as a vehicle for conducting stratigraphic and mapping studies in the open waters of the Pacific continental margin. B. R. Pelletier (1968) described a feasibility study conducted in the Arctic during which PISCES I was used in trial dives to make both geological and oceanographic observations. Since then submersibles of various types have been used for geological studies off the Canadian east coast (King and MacLean, 1970; Pelletier, 1970; Drapeau, 1970, 1973; Schafer, 1973), in Hudson Bay (Lewis and Sanford, 1972) and in Beaufort Sea.

The Pacific continental shelf and slope are ideal for geological submersible operations. Extensive seismic and echo-sounding profiles, obtained from surface ships during the past several years, have revealed numerous areas where bedrock is exposed on the sea floor. Dredge hauls have recovered Tertiary rocks from many locations and exploratory wells by Shell Canada Ltd. penetrated several thousand metres of Tertiary strata beneath structural culminations, some of which intersect the shelf surface. Except in a general way, little is presently known of the lithology, facies, stratigraphy, biostratigraphy, inter and intra-stratigraphic relationships, sedimentology and geochemistry of the layered succession beneath the shelf and slope. If PISCES IV could be used in much the same manner as helicopters are used in airborn geological work on land, as suggested by Pelletier (1968), both as an observational vehicle and as a sampling device, then detailed stratigraphic studies could be conducted on the continental slope. Once the stratigraphy and facies are defined, the various units beneath the shelf can be mapped in conjunction with seismic, side-scan sonar and bathymetric data, much of which is presently available.

The Submersible

PISCES IV is a small three-man submarine with a designed depth capability of 2000 m, although its present licensed depth limit is 730 m. Its length overall is 6. 1 m with a beam width (including thrusters) of 3. 0 m and a height of 3. 7 m (Fig. 62. 2). The unit weighs 10 360 kg in air and has a buoyant lift capacity of 455 kg in water when manned with two crew and normal equipment. When all jettisonable equipment is eliminated, including a manually operated drop-weight of 175 kg, its lift capacity is increased to 710 kg. The lift support system, consisting of oxygen tanks and CO_2 scrubbers, has a 200 man-hour capacity.

The submersible is constructed of two large spheres and two smaller ones connected by frames and covered by a removable fiberglass skin. The entire structure, supported by the steel framing, rests upon two skids.

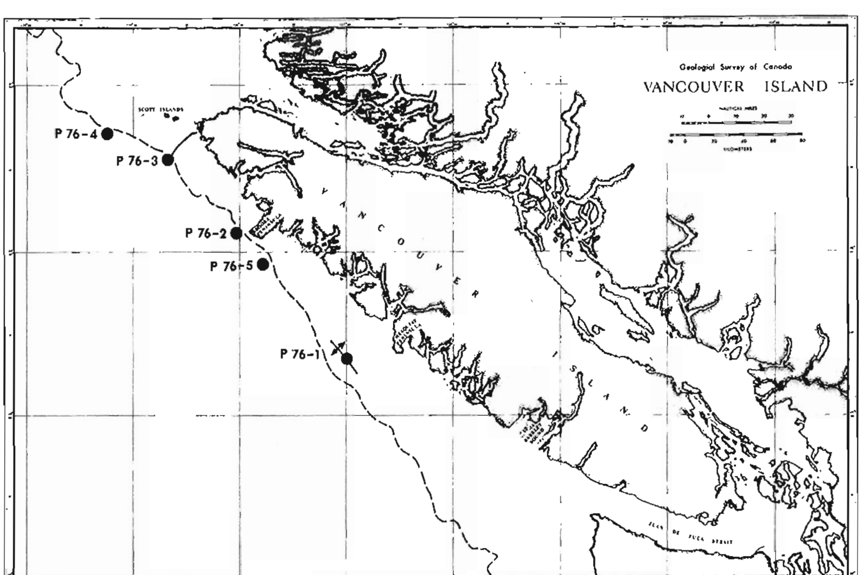


Figure 62. 1

Location map for dive sites on the continental shelf and slope off Vancouver Island. The line leading landward from P76-3 represent the track of seismic line 68-26. The dashed line represents the approximate edge of the continental shelf.

From: Report of Activities, Part A;
Geol. Surv. Can., Paper 77-1A (1977)

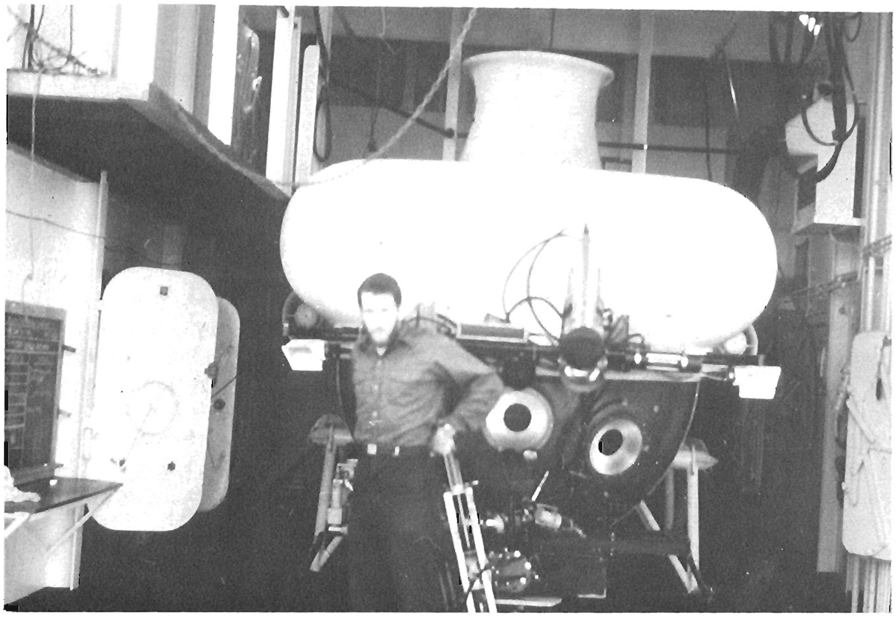


Figure 62. 2

PISCES IV submersible — front view. The entrance hatch is at the base of the "sail" at the top. Also shown are the video TV camera and side mounted arc lamps, two of the three view ports on the personnel sphere, the drill and claw sampler below the pilot's (Ian Sanderson) elbow, the skids and side mounted thrusters.

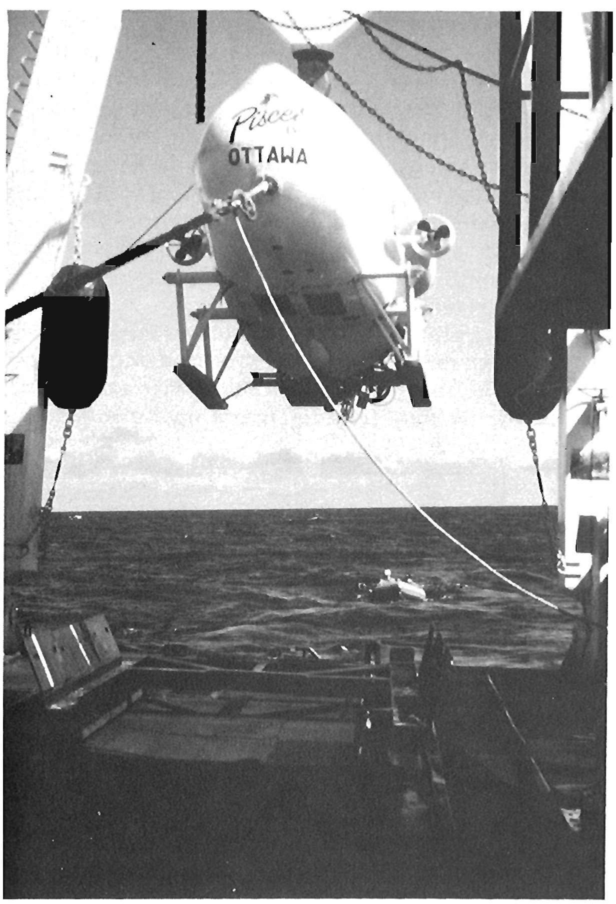


Figure 62. 3. PISCES IV being launched via the aft mounted A-frame on PANDORA II. The small support boat is used to assist in launching and recovering the diver.

The largest sphere, 203 cm in diameter, contains the instrumentation, control systems, personnel and life support system. This sphere has three view ports, each of 15 cm diameter, facing forward and slightly downward. The main aft ballast sphere is divided by a fiberglass partition at its equator. The lower hemisphere can hold up to 410 kg of water and the upper may be used for electronic equipment. Two forward ballast trim spheres, mounted on each forward upper quadrant of the personnel sphere, are 81 cm in diameter and can each hold up to 265 kg of water. They are used to adjust the buoyancy and fore and aft trim by means of a pump connected to the main aft ballast sphere. Additional soft ballast tanks provide the submersible with sufficient buoyancy to float on the surface but are filled with water during dives.

The submarine is provided with two thrusters, mounted on each side amidships which are capable of moving the boat at a speed of 1.5 knots across the bottom. They have complete axial rotation, are controlled independently, and provide the boat with optimum maneuverability.

The external equipment consists of a claw sampler with six degrees of freedom of movement that can lift up to 68 kg, a small rock drill, a recording video TV camera, and a 70-mm still camera with strobe. Two 1000-watt tungsten iodide lamps provide good lighting for visibility. A hydraulically extendable tray is used for storing samples. Another claw sampler with a lift capacity of 910 kg can be substituted for the drill.

Ship to submersible communication on the surface is accomplished by VHF radio and, in the subsurface by underwater telephone (Subcom single side-band transmitter-receiver). Submerged location equipment includes a 45 KHz pinger with a 3300 m range and an acoustic tracking and ranging transponder system that responds to signals from terminal equipment on the mother ship.

Figure 62. 4

The diver is detaching the hoist cable.

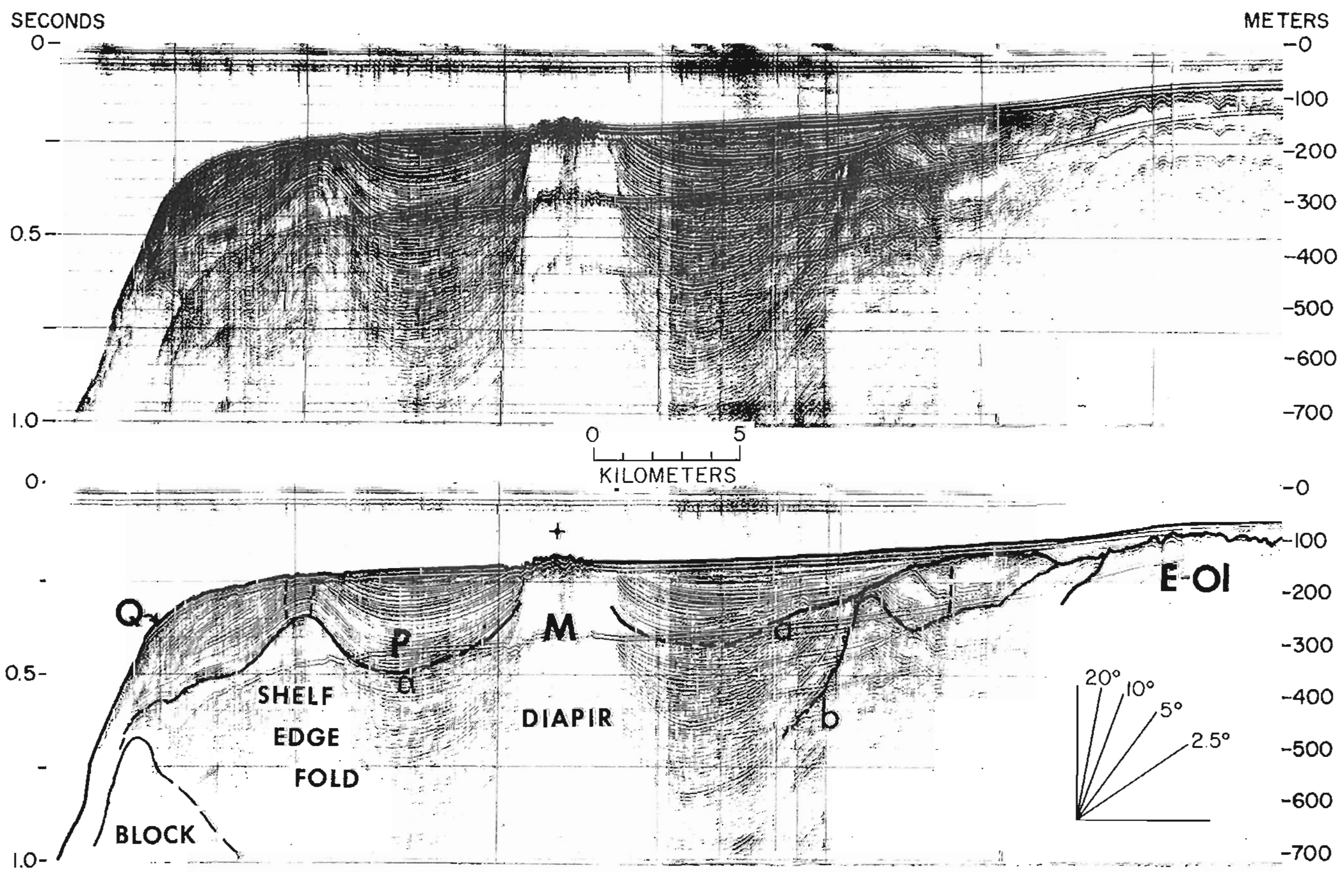


Figure 62. 5. Seismic line across the Nootka (Apollo) structure near dive site P76-1. Figure taken from Tiffin *et al.* (1972) in which

Q = Quaternary
P = Pliocene
M = Miocene
E-O1 = Eocene-Olgocene. *See* text for alternative interpretation.
Vertical Exaggeration = 16x.
a and b = Angular unconformities.

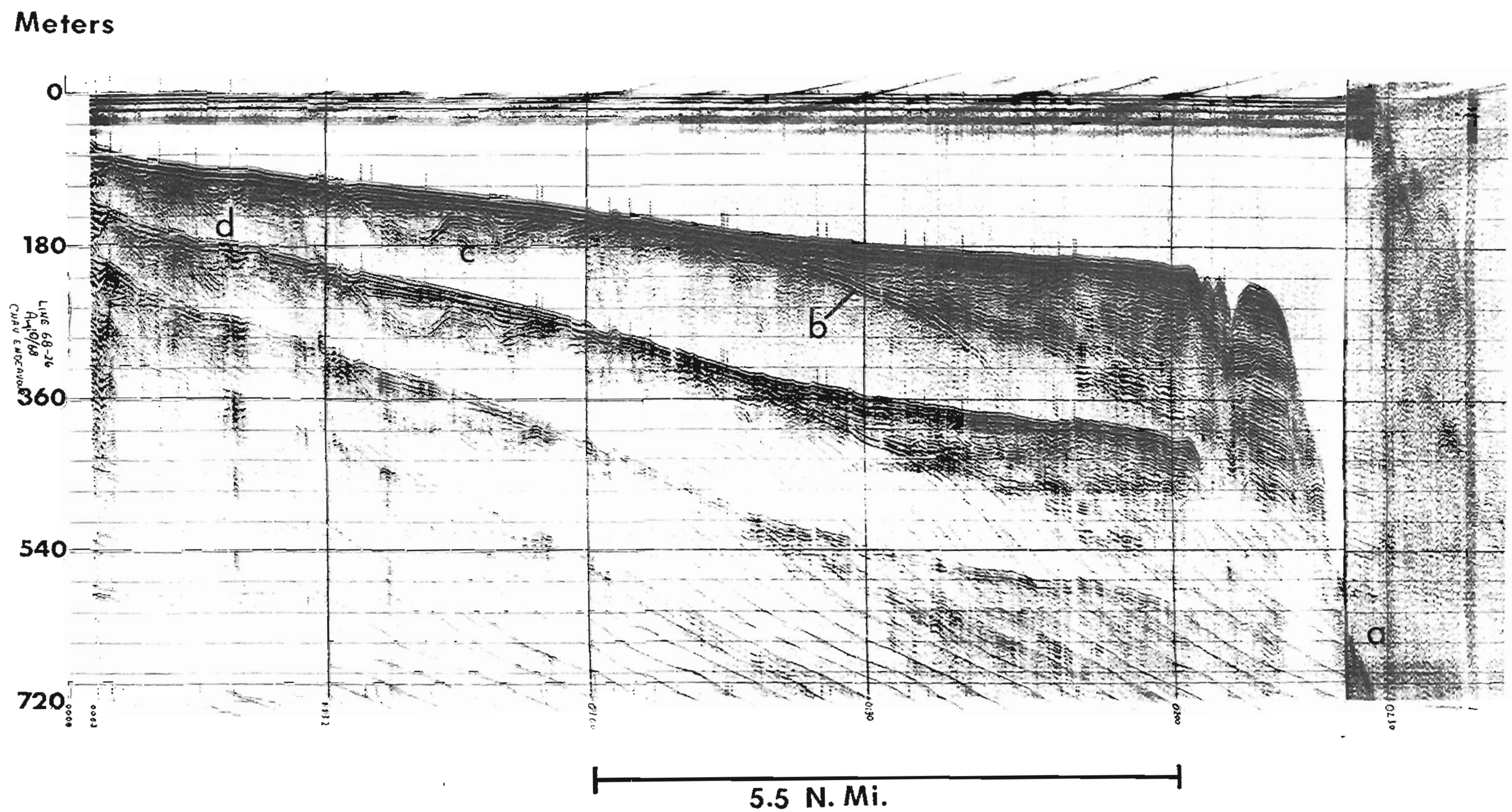


Figure 62.6. Seismic line 68-26 (*see* Fig. 62.1 for location). The locus of dive P76-3 was from position a to the top of the continental slope. Vertical exaggeration = 12x.

In addition to controls which operate the thrusters, ballast pumps, life support systems etc., internal equipment comprises a gyro compass, echo sounder and ranging sonar consol. Controls which operate the video and 70-mm cameras, drill, claw sampler and sample tray are also inside the personnel sphere.

The support vessel, M.V. PANDORA II, is fully equipped for maintaining and servicing the submersible. An aft mounted A-frame is used for launch and recovery. She also carries an 8-m diesel tracking launch equipped with portable VHF, HF and echo sounder.

Present operational limits are: a water depth less than 730 m, surface visibility of at least 3.5 km, wave height less than 1.5 m and a predicted maximum current during the dive of less than 2 knots. Other limitations related to safety and communications were found during these test dives. These must be rectified before further open water dives can be made.

Operational Procedures

The dive sites were preselected on the basis of known bathymetry and subsurface structure as revealed in seismic profiles. With the exception of P76-1, over the Nootka (Apollo) structure on the continental shelf, slope dives were conducted over the steepest known scarps in the given dive regions. Seismic profiles over the steep continental slope adjacent to northern Vancouver Island show many reflectors intersecting the sea floor. For the most part these outcrop areas are scarps thought to be formed by major faults on the shelf edge and slope (Tiffin *et al.*, 1972).

Upon arrival at the dive site the submersible is launched via the aft A-frame (Fig. 62.3) with the dive personnel inside. A diver accompanies the submersible into the water to detach the hoisting cable (Fig. 62.4). The diver is then recovered by dinghy.

Depending upon the degree of negative buoyancy attained the submersible descends at a rate between 10 and 18 m per minute. During descent observations are made on visibility, plankton density, fish populations etc. On the first two dives all observations were recorded in written form but on later dives a cassette tape recorder was used. The observor is in a prone position while looking through the viewports and, on long dives, the increasing discomfort of continually shifting from one elbow to another and straining to look upward at vertical cliffs through generally downward directed viewports renders writing very difficult and tape recording of observations almost necessary. Constant attention to the depth indicator, sonar, video screen and time is maintained in addition to viewport observations.

The slope dives were conducted by direct descent to the target depth, rather than by traversing down the slope. Although initial considerations of time and submersible operation demanded this method it is no longer considered safe. Since the viewports are downward facing an unsuspected, unstable overhanging mass of sediment could conceivably be triggered by the submersible when moving upward over an unknown slope. The results could be disastrous. Future descents will therefore traverse downslope to the target depth in order that the crew can obtain an overview of the geological succession, topography and slope stability prior to the detailed ascent.

Upon reaching the target depth, the submersible was directed up the closest and steepest slope found either visually or by sonar. Observations were made on the nature and structure of the bottom sediments, organic activity, bottom currents etc. Bedrock outcrops, commonly in vertical cliffs, were described in terms of the appearance of the rock type, strata attitudes, bedding thickness, internal sedimentary structures, and external structures such as joints, small faults and sedimentary dykes. On dive P76-5 a continuous video tape was recorded and a number of 70-mm photographs were taken. Samples collected were limited by the container size. The tray had only one divider which made it difficult to identify specific samples upon completion of the dive. To help overcome this problem, the sample was held up to a viewport and described before dropping it into the tray. Most samples were collected with the jaw sampler, the drill being used only once.

The dives were terminated when the edge of the continental shelf was reached. This was determined when, normally at about 200 m, a steady and uniform decline in slope angle was noted with a consequent lack of outcrop. The appearance of erratic pebbles, cobbles and boulders intermixed with surficial muds also signaled the top of the continental slope.

Future Operations

As these reported dives were trials for offshore stratigraphic and mapping studies, what was learned is useful in planning and conducting future programs.

1. As observers become increasingly experienced at making and recording geological observations under water the amount and quality of information will improve. The effects of submarine erosion, weathering, light scattering and biotic activity on the stratigraphic successions were new to the observers resulting in somewhat simplistic observations and incomplete recording. In addition, observations during these first dives in a dark and alien environment were not helped by initial feelings of apprehension which decreased with familiarity with the techniques and methods of operation.
2. The licensed depth of the submersible (presently 730 m) should be extended toward its designed capability of 2000 m. This would allow far greater access to exposed thick stratigraphic successions on the middle and lower slope.
3. A number of modifications to the submersible are required to make the operations safer and more effective:

   a) Increased and improved sample storage capacity will allow more samples (30 or more) to be obtained without the danger of mixing.

Figure 62.7

Bioturbated muds showing burrow holes, feeding trails. Depth 525 m. P76-5.

b) An additional claw sampler would be more helpful than the drill. The problem inherent in using the drill is that the submersible must increase weight by taking on water ballast to prevent the boat from lifting off bottom when pressure is exerted on the drill. The main aft ballast sphere must be filled to capacity, requiring a long period of pumping. Once a core is obtained, it takes even longer to pump the water out before the boat can move.

c) A pan-and-tilt platform for the video TV and 70-mm cameras is necessary. It is often difficult and slow to point the sub at the subject to be photographed.

4. An accurate positioning system (minifix) should be used for precise horizontal control of detailed bathymetric site surveys of proposed dive sites.

5. Bottom mapping will require three component acoustic transponder systems (range and bearing) for horizontal control.

All of these modifications are feasible and some of the equipment is at hand. A major part of the requirement is more experience.

Geology

Dive site P76-1 (49°21.5'N; 127°00.0'W) was chosen on the Nootka (Apollo) structure (Tiffin *et al.*, 1972; Shouldice, 1973) on the core of which Shell Anglo Apollo J-14 was spudded (Fig. 62.1). The well penetrated 3096 m of Plio-Pleistocene to Miocene strata consisting of grey, greenish grey and yellowish brown, calcareous mudstones, minor fine grained sandstones and siltstone. The structure is a northwesterly trending anticline, arcuate (convex toward shore) with dimensions about 21 km long by 3 km wide. The structure is expressed on the sea floor by a series of cuestas, the steeper slopes of which face toward the core. The cuesta relief seen is about 4.5 m with the inner face near vertical and the dip-slope inclined away from the axis at between 10 to 15 degrees. Portions of two opposed sets of ridges were observed and some appear to have been cut by minor extension fractures which transect the cuesta trends at high angles. Two outcrop samples were obtained which consist of dark brownish green (wet), medium brownish grey (dry), blocky, friable, moderately well indurated, well sorted, matrix supported siltstone. The rock is intensely burrowed by modern organisms. A presumed age of Miocene-Pliocene has been determined on the basis of poor

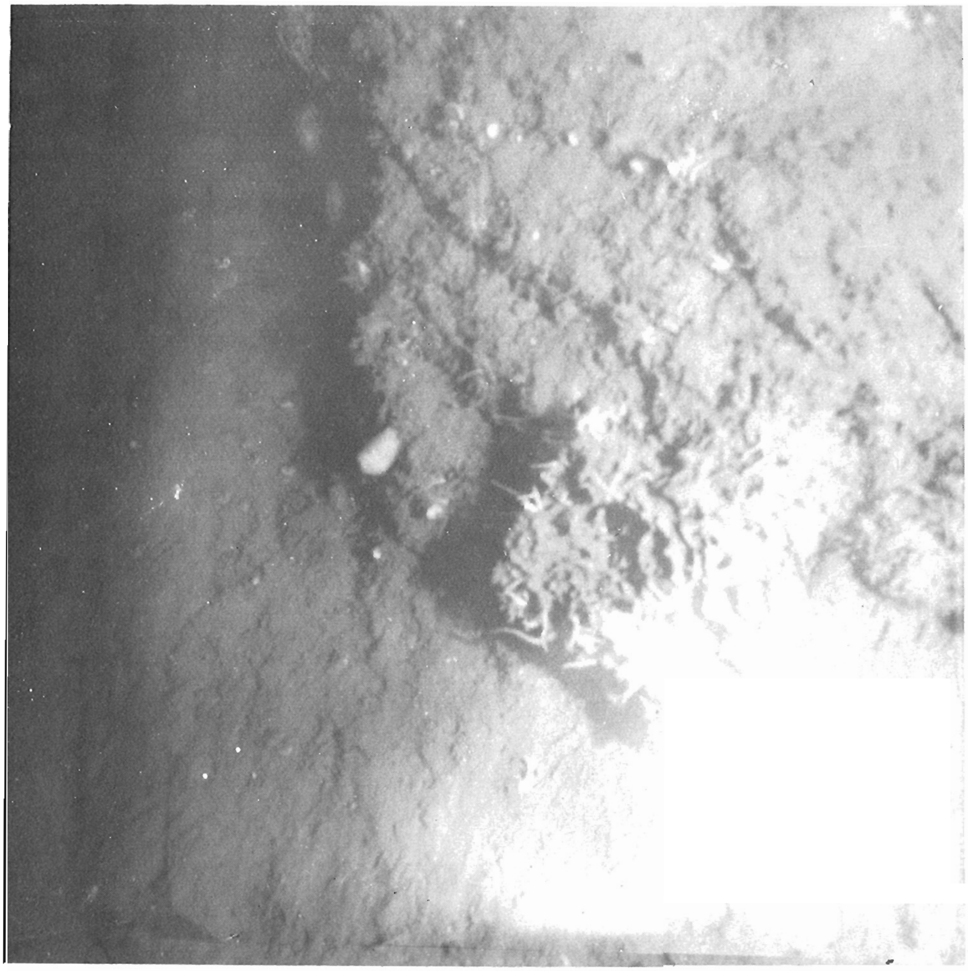


Figure 62. 8

Horizontally layered and mud covered outcrop of the pyroclastic unit. These cliffs are commonly vertical and several metres in height. Depth 405 m. P76-5.

microfossil content of the samples, however, earlier dredge hauls recovered from near this site have yielded sediments of Pliocene age.

A number of additional observations were made. The intervening flat areas between the cuestas were commonly littered with rounded to subrounded cobbles and boulders up to a few metres in maximum dimension. In boulder-free areas the bottom consisted of brownish to greenish grey mud, intensely bioturbated and occasionally rippled into sand waves with about 14 cm of relief and very short (approx. 30 cm) wave length. The trend of the waves was not recorded. Bottom feeding fish were abundant.

Figure 62. 5, reproduced from Tiffin *et al.* (1972), illustrates the subsurface geometry across the Nootka (Apollo) structure. In the lower half of the diagram, the well symbol represents the location of Shell Anglo Apollo J-14. The structure clearly involves the sea floor and, at least in its upper 700 m, is diapiric in origin. Given a Pliocene age for the sea floor samples in the core of the structure then the angular unconformity (a) may represent an inter Pleistocene-Pliocene hiatus with the section above (a) being Pleistocene and Holocene in age. The unconformity (b) would thus represent the hiatus between Pliocene and Middle Miocene sediments described and illustrated by Shouldice (1972). If this interpretation is correct then major Quaternary deformation has occurred on the continental shelf of the Pacific margin, represented by diapirs, compressional folds and faults.

The remaining four dives were conducted over the continental slope adjacent to northern Vancouver Island.

At dive site P76-2 (50°07. 0'N; 128°02. 4'W), immediately west of Cape Cook on Brooks Peninsula, a succession of black, dark purplish brown and dark grey weathering, hard, dense, in part calcareous, grain supported, fine to medium grained greywackes occur in near vertical cliffs observed over a height of 230 m. The succession appears massive, complexly jointed, and, where bedding is recognizable, is deformed with dips ranging from 30° to vertical. At the base of the cliffs are black, blocky, hard, dense, pyritic argillites and at the top, the uppermost sample (290 m) included polymictic diamictite, the clasts of which comprised subangular to rounded gabbro, chert, carbonate and argillite in a fine greywacke matrix.

Cliffs at this site are vertical with intervening narrow, more gently sloping surfaces between. The bases of the cliffed units are strewn with large blocks of the greywacke succession which rest among bottom muds at angles of repose of more than 30°. The muds are easily triggered into turbidity flows that move downslope at 2 to 3 knots.

Figure 62.9

Horizontally layered outcrop of the pyroclastic unit. Depth 298 m. P76-5.

This succession may represent part of the Pacific Rim Complex (Tofino Greywacke) described by Muller (1973) of general Late Jurassic and Early Cretaceous age. A small outcrop of rocks assigned to the Pacific Rim Complex occurs on Brooks Peninsula along the coast south of Cape Cook (Muller *et al.*, 1974). Rock specimens from that locality and others on Vancouver Island compare favourably with those collected on the slope. An alternative possibility is that the slope rocks may represent part of the Longarm Formation – Queen Charlotte Group interval, units containing lithic greywackes of Early Cretaceous age on Vancouver Island (Muller *et al.*, 1974) and Queen Charlotte Islands (Sutherland Brown, 1968).

Dive site P76-4 (50°42.9'N; 129°13.9'W) was about 19 km southwest of Triangle Island, the most westerly of the Scott Islands. Triangle Island is underlain by rocks of the Pacific Rim Complex (Muller *et al.*, 1974). At the base of the observed succession (720 m) rocks very similar to those observed at P76-2 occur in vertical cliffs of unknown height and which extend to unknown depths below. The exposures consist of dark purplish brown, massive, closely and complexly jointed, hard, dense, highly feldspathic siltstone. The matrix is locally calcareous and argillaceous. The exposures are blocky and rubbly; no bedding was observed. Samples yielded a very sparse microfauna of Early Cretaceous age (Barremian-Albian), and thus on this meager evidence, the basal unit of this site is tentatively correlated with the Longarm Formation-Queen Charlotte Group interval.

Unconformably overlying the basal unit at site P76-4 is a somewhat monotonous succession, about 415 m thick, consisting of pale grey weathering, horizontally thin- to medium-bedded, friable, bentonitic mudstone and siltstone. Much of the quartz component appears primary (volcanic) in origin, is well sorted, matrix-supported, and is mixed with varying amounts of finely disseminated micas, organic grains and diatoms. The pyroclastic origin of the succession is supported by the common occurrence of spherulites (dia. 1 mm) which appear to represent devitrified sphaeroidal glass (J.G. Souther, pers. comm.). The succession is uniformly broken by vertical joints which trend orthogonally to the trend of the slope. Along a number of these fractures siltstone dykes have been emplaced which commonly protrude a few centimetres above the surrounding sea floor.

At site P76-3 (50°34.2'N; 128°41.7'W) much the same succession as that studied at P76-4 was observed. There the basal feldspathic siltstone unit (Longarm Formation-Queen Charlotte Group) occurs at a depth of 650 m where beds about 60 to 90 cm thick are horizontal.

Figure 62.10

Horizontally layered outcrop of the pyroclastic unit. Depth 250 m. P76-5.

At a depth of 540 m what appeared to be a massive conglomerate was observed but not confirmed. Overlying the "conglomerate" is the pyroclastic unit, about 434 m thick and extremely well exposed as a series of vertical cliffs and ridges, separated from one another by narrow gently sloping muddy bottom. Towards the top of the succession small vertical faults have about one to two metres of vertical displacement.

Figure 62.6 is seismic line 68-26, used to plan the traverse at P76-3. Point 'a' represents the top of the basal feldspathic siltstone unit which could conceivably be represented by the 'b' reflector, beneath which older rocks are folded at 'c' and 'd'. The location of the seismic line is shown on Figure 62.1. The line trends shoreward toward coastal outcrops of Lower Jurassic Bonanza volcanics and the Longarm Formation (Muller *et al.*, 1974). At the edge of the continental shelf are a series of near vertical faults. The pyroclastic unit is represented by the succession above 'a' and is seen to thin rapidly toward shore to a zero edge at about 1:15 (time marks along lower margin of the record).

At P76-5 (49°57.5'N; 127°47.5'W) the pyroclastic succession was examined but its base was not observed at the deepest point of the dive (530 m). Apart from a 14-cm-thick unit of dark brown weathering, hard, dense, laminated, fossiliferous (*Pecten* sp.) siltstone in the middle part of the sequence, the succession is uniform in lithology and identical in appearance to the pyroclastic unit observed at sites P76-3 and P76-4. Towards the top of the section, a well developed planar cross-bedded unit, about 0.3 m thick occurs within a cliff face about 14 m high. The cross-bedding shows a westerly direction of transport with foreset dip angles averaging about five degrees.

The age of the succession at P76-5 is late Miocene to late Pliocene based on foraminifera. Sediments of similar age have been obtained from dredge hauls in the vicinity of site P76-3. The dominance of *Uvigerina peregina, U. juncea, Epistominella pacifica* and *Cassidulina cushmani* are indicative of an upper bathyal paleodepth.

In view of the limited information on the lithology and lateral distribution of the pyroclastic unit, statements pertaining to source areas are totally speculative. Although volcanics and associated pyroclastic rocks are known from the land areas of the Pacific margin (Sutherland Brown, 1968) and southeast Alaska (Brew *et al.*, 1966; Lathram *et al.*, 1965), most of these are Paleogene to early Neogene in age. Muller *et al.* (1974) have described basic Tertiary volcanic rocks, associated tuffs and volcanic conglomerates which occur within the Brooks Peninsula Fault Zone extending from

Brooks Peninsula northeastwards to Port McNeill on the north coast of Vancouver Island. Preliminary whole-rock potassium-argon dating of the Twin Peaks basalts which occur toward the northeast end of this belt has provided an age of 7.9 and 7.6 m.y., indicating a late Miocene to early Pliocene age. The thickness of the pyroclastic succession beneath the continental shelf and slope (415 to 434 m) suggests a nearby source and it is tempting to speculate that such might be found in volcanic vents within the Brooks Peninsula Fault Zone.

The remaining figures (Figs. 62.7 to 62.10) were taken at site P76-5 from that portion of the slope underlain by the pyroclastic unit. The photos are not of good quality owing to the low visibility and low film speed (ASA 80) relative to the distances between camera and subject. The captions to each figure explain the relevant features.

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63. GRAVITY AND MAGNETIC SURVEY OFF VANCOUVER ISLAND, 1975

Project 710048

D.L. Tiffin and R.P. Riddihough[1]
Regional and Economic Geology Division, Vancouver

During 1975, two contiguous marine gravity and magnetic surveys were carried out from CSS 'Parizeau' by the Earth Physics Branch and the Geological Survey of Canada, in conjunction with the Canadian Hydrographic Service, as part of the Natural Resource charting program (Tiffin and Currie, 1976). The survey, in co-operation with U.S. Geological Survey, covered the area from the coast out to 127°W between 48° and 49°N. A second area extended from the coast out to 128°30' between 49°and 50°N.

Instrumentation and Navigation

Navigation and survey location were provided by the Canadian Hydrographic Service using a Mini-Fix hyperbolic system with a range to 200 km. Lane count was established by a Trisponder system located on high elevations. Absolute accuracy of positioning was estimated at better than ± 100 m. Although positions were recorded every 10 secs, the plotting of geophysical data was based on positions at 5 minute intervals (approximately every 2 km at normal speeds). Track spacing for the gravity surveys varied from 3 to 5 km over the continental shelf and 5 to 8 km in deeper water with check lines every 30 km. Magnetic measurements were recorded over the same tracks but also included measurements along detailed hydrographic sounding lines at spacings of less than 0.5 km covering areas shallower than 400 m. A total of 12 000 km of magnetic data and 8000 km of gravity data was obtained.

For both surveys a Barringer proton precession magnetometer was towed behind the ship at a distance of about 200 m. Data were recorded on tape at 6 sec intervals. Diurnal variations were monitored from a temporary land magnetometer station operated at Ucluelet for the southern survey. These data compared favourably

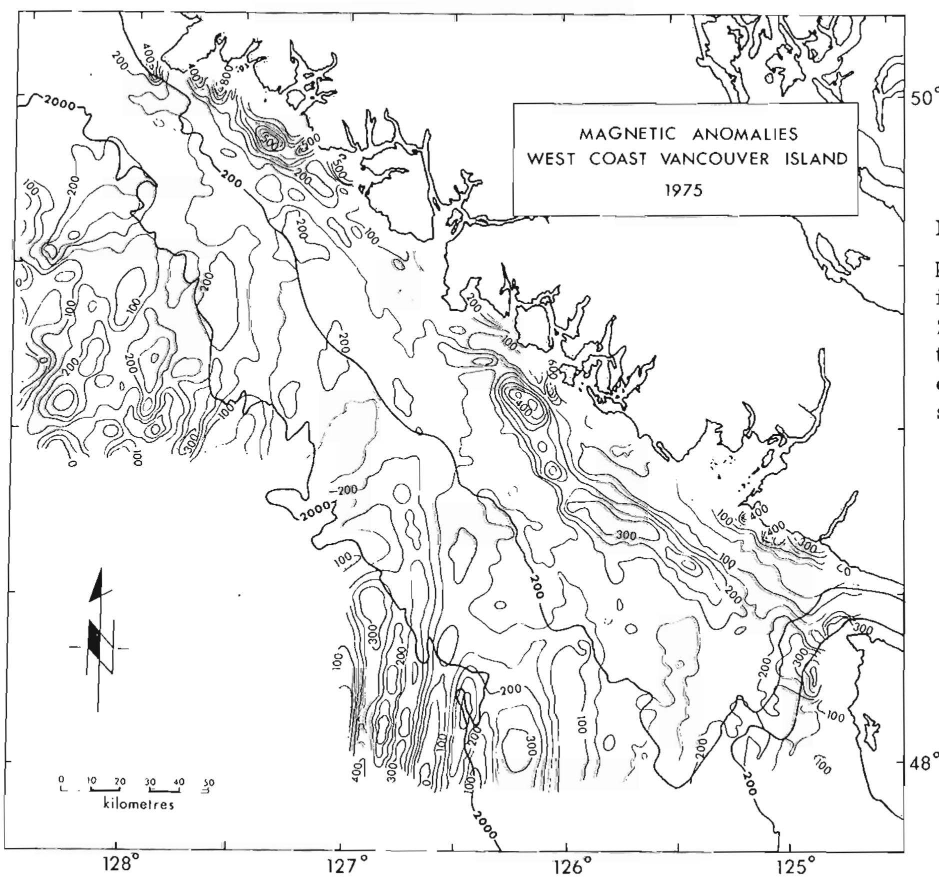


Figure 63.1.

Magnetic anomaly map. Contour interval is 50 gammas. The 200-m and 2000-m bathymetric contours indicate the approximate position of the shelf edge and the base of the continental slope respectively.

[1]Victoria Geophysical Observatory, Earth Physics Branch, Victoria.

Contribution of Earth Physics Branch N°585

From: Report of Activities, Part A;
Geol. Surv. Can., Paper 77-1A (1977).

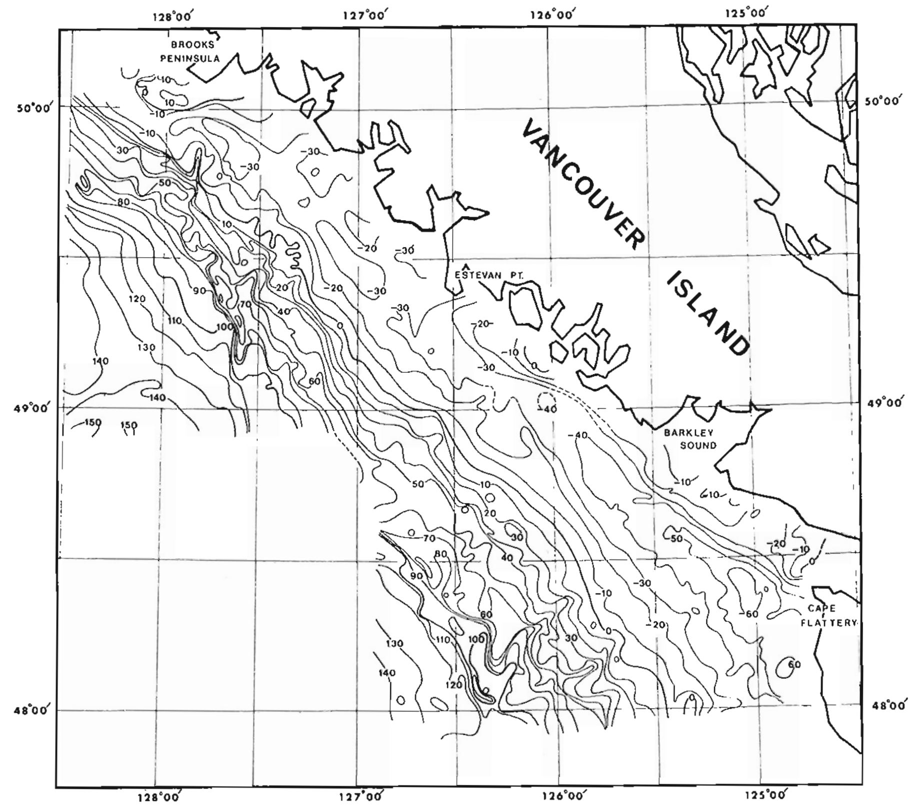


Figure 63. 2.

Bouguer gravity anomaly map. Contours at 10 mgal intervals, reference field, National Gravity Net (Earth Physics Branch, 1974).

with data from Victoria Geophysical Observatory for monitoring purposes, so the Victoria Observatory data were used for correction of diurnal variations for the northern survey.

Gravity was measured for the first survey with a 3-axis LaCoste & Romberg dynamic gravimeter (S56) positioned near the waterline in the centre of the ship. This was calibrated and read at Esquimalt, Victoria at the beginning and end of the survey. For the northern survey the gravity meter was a 2-axis LaCoste & Romberg dynamic gravimeter (S32) which was operated in the same position and similarly calibrated. Data were recorded on magnetic tape and chart records.

Data Reduction

The 6-sec magnetic data were smoothed to produce values at 1 minute intervals and diurnal corrections were made. After digital removal of the IGRF, data was computer contoured by Digitech Ltd. under contract. Mean cross-over error was 8. 1 gamma for the southern survey and 6. 4 gamma for the northern one. The resulting magnetic anomaly field is shown in Figure 63. 1.

Gravity data were reduced by computer to observed gravity and free-air values averaged over 5 minute intervals using Eotvos corrections derived from the 5 minute navigation fixes. After editing for spurious readings at turns, ends of lines, etc., the data sets were adjusted for minimum cross-over errors (total cross-overs, 597) and reduced to the Canadian National Gravity Net 1974 (Earth Physics Branch, 1974). Standard deviations of cross-over errors for the two data sets after adjustment were 1. 78 and 1. 81 mgals. Bouguer anomaly values were calculated using the depth obtained from simultaneous sounding records for the centre of the 5 minute interval concerned. Contouring and final editing were done by hand.

The magnetic map shown is available on open file from the Geological Survey of Canada (Open File 392) and the gravity map from Earth Physics Branch, Gravity Division (Open File Maps 76-2, 77-1). Final coloured versions will be published in the Natural Resource Map Series.

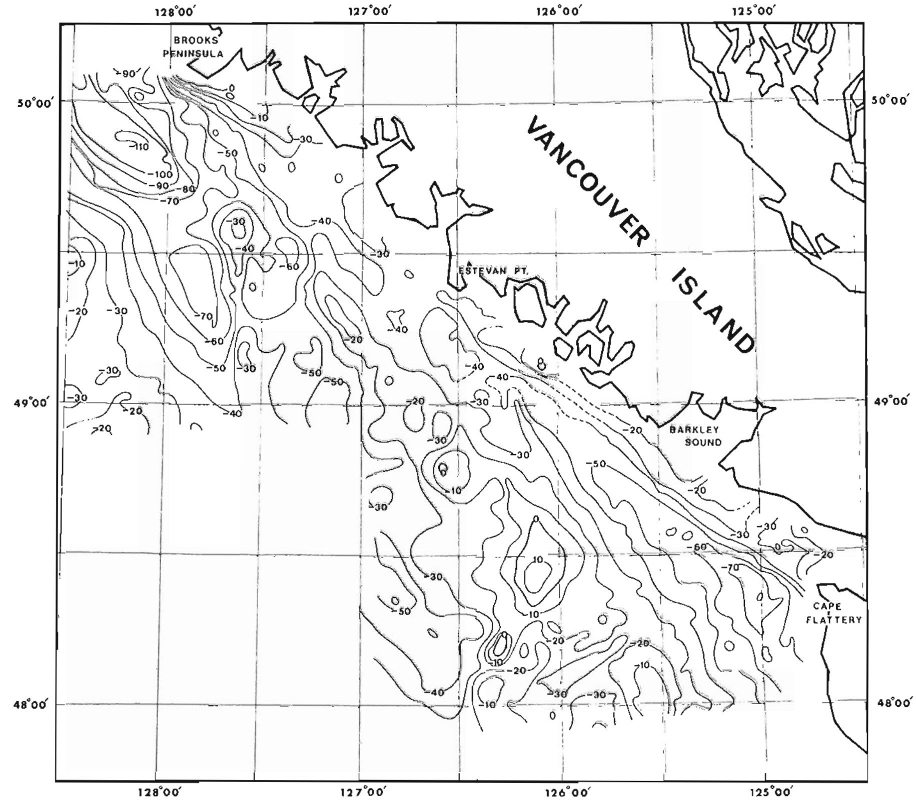
Main Features of the Results

The dominant feature of both the Bouguer gravity and magnetic maps is the contrast between the oceanic and continental areas. The zone intermediate between them (approximately the continental slope) is shown on the Bouguer anomaly map (Fig. 63. 2) by the steep gradient from +10 to +100 mgal, interpretable in general terms as reflecting the rise in the Moho from continent to ocean. On the Free-Air anomaly map of Figure 63. 3, at least south of 49°30', the same line of division is paralleled by a gravity ridge, ranging from -30 to +10 mgals, again explicable as part of the edge effect produced at the continent/ocean boundary.

On the magnetic map of Figure 63. 1, the two areas present a striking contrast with strong, north to northeast trending linear anomalies over the oceanic area and lower relief or northwest trending anomalies

Figure 63.3.

Free air anomaly map. Contours at 10 mgal intervals, reference field, National Gravity Net (Earth Physics Branch, 1974).



over the continental shelf. The linear anomalies in the oceanic area are a continuation of the series of ocean crust magnetic anomalies created at the Juan de Fuca Ridge and first mapped by Raff and Mason (1961). The greater detail of the present survey clearly demonstrates their continuity beneath the continental slope (earlier examined by Barr, 1974) and outer continental shelf. The oldest anomaly identifiable beneath the slope and shelf is about 10 m.y. suggesting that subduction must have occurred since that time.

North of 49°, the anomalies in the oceanic area are more fragmented and exhibit trends generally east of north. They lie in an area largely blank on the map of Raff and Mason and may represent part of the anomaly sequence created at the Explorer Ridge and affected by complex rotations and interactions suggested for this region (Barr and Chase, 1974; Riddihough, in press).

On the continental shelf, the most outstanding features are the extended, northwesterly trending anomalies parallel to the coast of Vancouver Island from Cape Flattery to Estevan Point. The geographic coincidence of the negative Bouguer and Free-Air gravity anomalies with the positive magnetic anomaly, the Prometheus anomaly (Shouldice, 1971; MacLeod, *et al.*, in press), is not perfect but significantly close. Preliminary interpretations (Riddihough, 1976; Tiffin and Currie, 1976) suggest that the gravity anomalies arise from an elongate basin of up to 5-km of sediments (the Tofino Basin) and the Prometheus magnetic high from volcanic rocks within the basin. A Shell Canada test well drilled on the anomaly, reached volcanics at 5850 feet and drilled through 1800 feet of volcanics before being abandoned. The anomaly lies along the same trend as the volcanics of the Crescent Formation on nearby Olympic Peninsula and has a similar magnetic character. Those volcanics are Eocene in age, correlative with the Metchosin volcanics on Vancouver Island (MacLeod, *et al.*, in press, Muller, 1977) and are interpreted to be of oceanic ridge and seamount origin. If the volcanics associated with the Prometheus magnetic anomaly are Eocene, the history of the continental shelf off southern Vancouver Island is probably similar to that of the Olympic Peninsula.

If underthrusting of the oceanic crust is occurring, the younger oceanic rocks must be subducting below the shallower and older volcanics of the Prometheus anomaly. The plane of underthrusting must lie to the west of the Prometheus anomaly marking the westward edge of the continental magnetics.

Both gravity and magnetic anomalies decrease and fade away near 49°15'N and north of this the shelf anomalies appear to assume a different character. The 'edge effect' ridge of the Free Air anomaly map also terminates near 49°30'N and preliminary gravity interpretation of a profile across the continental slope and shelf north of this suggests that the complete shelf and slope area may be constructed of low density sediments. A strong magnetic 'high' extends south-eastward from the Brooks Peninsula parallel to the coast over the area of the Kyuquot Uplift (Tiffin, *et al.*, 1972) and seems to be associated with a gravity 'high' over the same area. The anomalies may be geologically related to basic rocks in the core of the Uplift similar to those exposed on Brooks peninsula (Muller, *et al.*, 1974).

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Ian F. Young[1] and R.L. Chase[1]
Regional and Economic Geology Division

Introduction

A marine geological-geophysical survey of the southwestern portion of Hecate Strait and northwestern Queen Charlotte Sound was carried out aboard C.F.A.V. ENDEAVOUR from June 18 to June 28, 1976, by the Department of Geological Sciences and Institute of Oceanography, University of British Columbia. The aim was to investigate the structure of the western portion of the Queen Charlotte Basin and to map a possible submarine extension of the Sandspit Fault, a major fault traced for 65 km on eastern Queen Charlotte Islands and known to have several hundred metres of vertical displacement and unknown, but large, horizontal movements (Sutherland Brown, 1968). The data collected during the survey will form part of an M.Sc thesis by one of the authors (I.F.Y.) at University of British Columbia. In addition to the offshore work, field work consisting of gravity measurements across the Sandspit Fault, limited geological mapping, and sample collection for radiogenic age determinations, was conducted on the Queen Charlotte Islands. Preliminary results of this work will be published elsewhere (Young and Chase, in press). The gravity profiles near Skidegate Inlet indicate 1500 - 2000 m of vertical displacement on the fault with east side down.

The Queen Charlotte Basin has been the site of extensive marine geophysical surveying by Shell Canada Ltd. in their exploration program on the western Canadian continental shelf (Shouldice, 1971). Eight offshore exploratory wells penetrated a thick succession of nonmarine to marine Pliocene and late Miocene sediments to mid- and lower-Tertiary and late Mesozoic or older volcanics. Few geophysical data are published on the structural and stratigraphic relationships within the basin. A joint program of systematic resource mapping by the Geological Survey of Canada (Vancouver), Earth Physics Branch (Victoria) and Canadian Hydrographic Service has not yet extended to Hecate Strait.

On Graham Island and northern Moresby Island, the Sandspit Fault is at least 65 km long. It forms the western margin of the Neogene Queen Charlotte Basin, separating the basin from Mesozoic and Eocene rocks of the mountainous spine of Queen Charlotte Islands. The basin margin lies beneath the continental shelf east of southern Moresby Island, the area in which this survey was carried out.

[1]Institute of Oceanography and Department of Geological Sciences, University of British Columbia, Vancouver, B.C. V6T 1W5

Data Collection

Twenty-six survey lines on which were obtained continuous seismic reflection profiles (CSP), bathymetric and magnetic data, were run across southwestern Hecate Strait and northwestern Queen Charlotte Sound (Fig. 64.1). The lines were run northeast-southwest, perpendicular to the presumed trend of the marine extension of the Sandspit Fault, at approximately 5 km intervals. Five short lines were also obtained in Skidegate Inlet. In total, 963 km of CSP and 1369 km of magnetic data were collected. Primary navigation for the cruise was by radar, with additional fixes provided by satellite and Loran A.

Total magnetic field intensity, measured by a Varian proton precession magnetometer at ten seconds intervals, was recorded in both analog and digital form. Records from the geomagnetic observatory located at the Victoria Geophysical Observatory will be used for correction for diurnal variations, and data collected during severe magnetic disturbances will be eliminated. The source for CSP was an airgun with 5- or 40-cubic-inch chamber fired at approximately 10 MPa (1500 psi) every 2 seconds. Seismic reflections were received by an 8-m linear array of Geoflex hydrophones with Kennedy preamplifier, fed into a Kennedy ramped amplifier and filter system and recorded in the passband 800 - 5000 Hz by an E.P.C. graphic recorder.

Bathymetric profiles were recorded in analog form from a 12-kHz wide-beam echo sounder. Malfunction of the Edo graphic recorder prevented acquistion of accurate bathymetric profiles along a number of the survey tracks. Eighteen bottom sample stations were occupied within the survey area (Fig. 64.1) where the seismic information indicated little Quaternary cover. Chain bag dredges, Peterson grab sampler, and Phleger and Kullenberg gravity corers were used. Sampling for the most part was unsuccessul: Only two samples of possible outcrop were obtained. Lack of a high-resolution seismic system for site selection and a mantle of Pleistocene sand and gravel over much of the survey area were limiting factors in the sampling operation.

Discussion of Results

Stratigraphic control for the preliminary interpretation of seismic records is provided by the exploratory wells drilled by Shell Canada Ltd. and by dredge samples obtained during this cruise. Magnetic data, not yet processes and mapped, is not discussed.

Three seismic reflection profiles from the southwestern portion of Hecate Strait serve to illustrate the structure of that portion of the Queen Charlotte Basin. From 53°07'N to 52°46'N (Fig. 64.2, profile 14), easterly dipping Upper Tertiary (Pliocene?) sediments

From: Report of Activities, Part A;
Geol. Surv. Can., Paper 77-1A (1977).

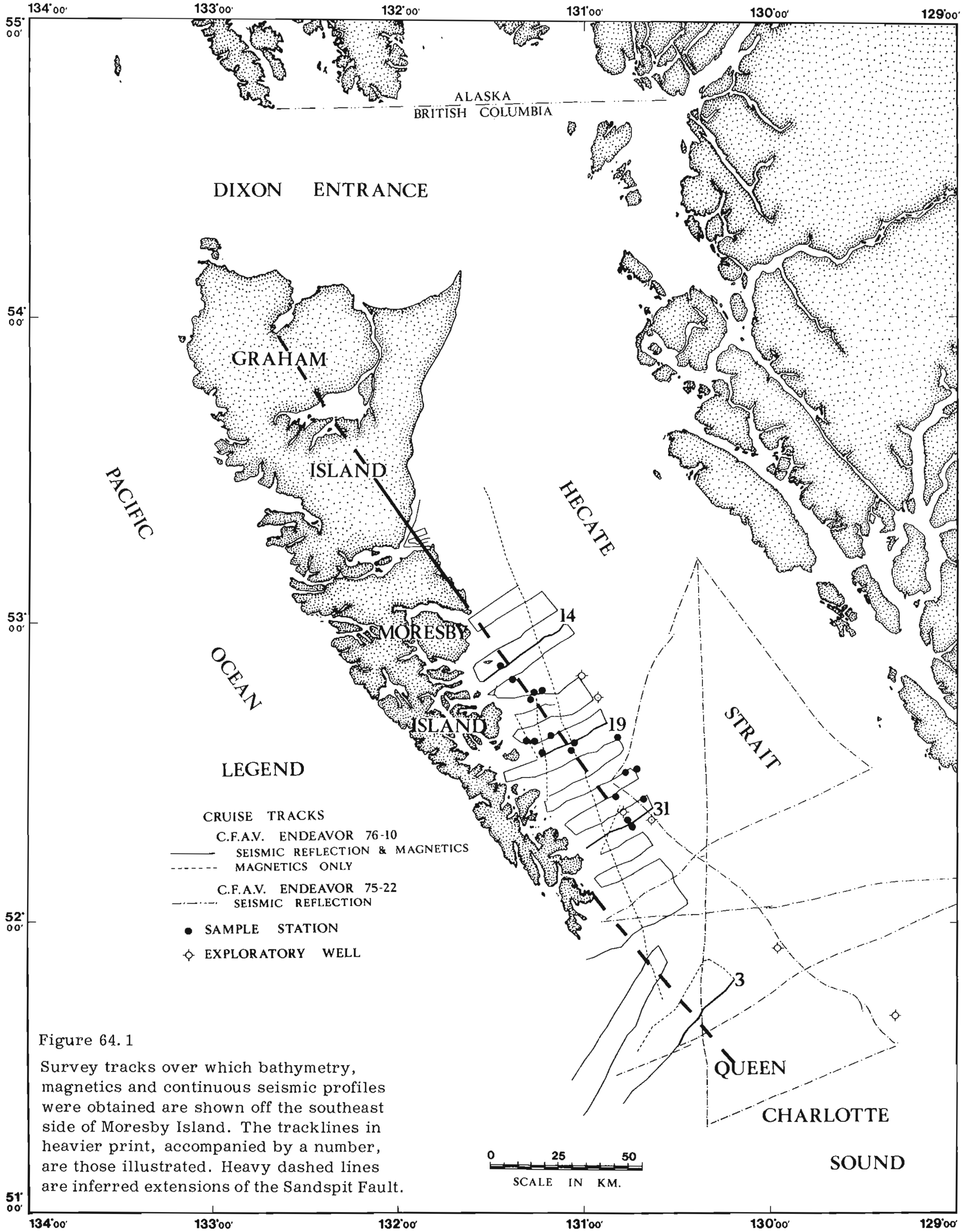

Figure 64.1

Survey tracks over which bathymetry, magnetics and continuous seismic profiles were obtained are shown off the southeast side of Moresby Island. The tracklines in heavier print, accompanied by a number, are those illustrated. Heavy dashed lines are inferred extensions of the Sandspit Fault.

are overlain unconformably by stratified Pleistocene drift. The channelled erosion surface is probably a Quaternary surface though the possibility exists that the unconformity may be Pliocene (Shell Canada Ltd., 1965; Shouldice, 1971). The Quaternary sediments appear unfolded. Volcanic rocks, probably the Eocene Masset Formation, outcrop west of an inferred fault along the western edge of the Queen Charlotte sedimentary basin.

From 52°46'N to 52°35'N (Fig. 64.2, profile 9) gentle folds occupy a zone east of the basin edge. Fold amplitude varies from 100 to 180 m and wavelength from 0.75 to 1.25 km. Normal fault and smaller growth features, recognized by offset reflectors in the seismic profile, are evident. Fold axes and faults trend 140°, parallel with the proposed southeastern extension of the Sandspit Fault. Folding is restricted to the Upper Tertiary sediments: overlying unconformable glacial sediments appear undisturbed.

From 52°35'N to 52°15'N (Fig. 64.2, profile 31) the folds that characterize the zone farther north appear poorly developed. Faults cannot be correlated from profile to profile, hence appear to be local. The basin edge is not defined on profile 31 but may be represented in this area by an onlap of Tertiary sediments on Mesozoic basement rock.

Calcite-cemented sandstone obtained at Station 23 (Fig. 64.2, profile 31) resembles closely the Mio-Pliocene Skonun sandstone (Sutherland Brown, 1968) exposed on Graham Island.

A profile across the outer shelf and slope in the northwestern portion of Queen Charlotte Sound (Fig. 64.3, profile 3) shows unconformities and erosion by canyon-forming processes, but no faults are identified. The inferred position of a southeastern extension of Sandspit Fault (Chase *et al.*, 1975) should coincide with the shelf-slope break in this region. The lack of faults

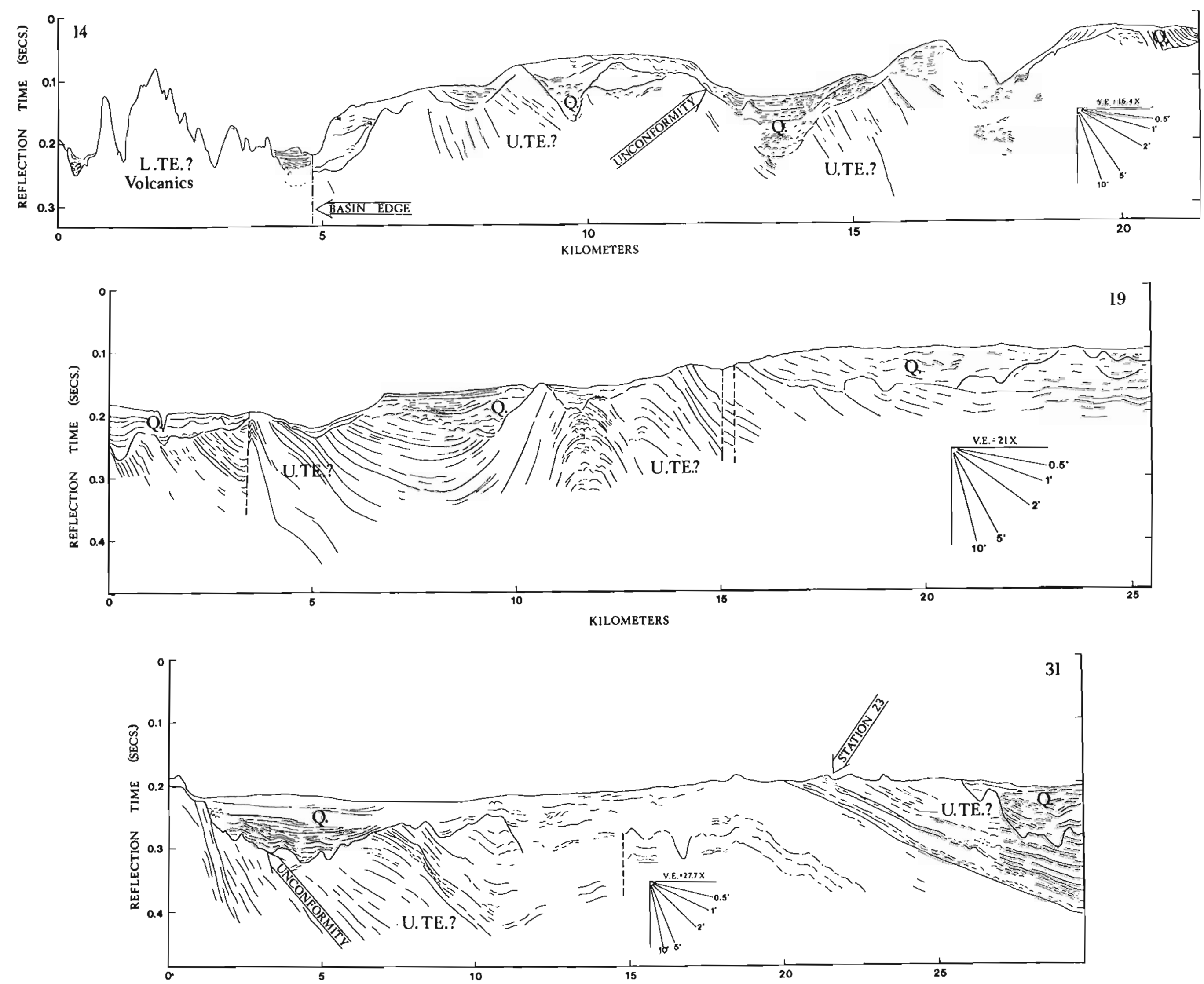

Figure 64.2. Line drawing interpretations of continuous seismic profiles numbered 14, 19, and 31 located on Figure 64.1. The Queen Charlotte Islands are beyond the left end of the profiles.

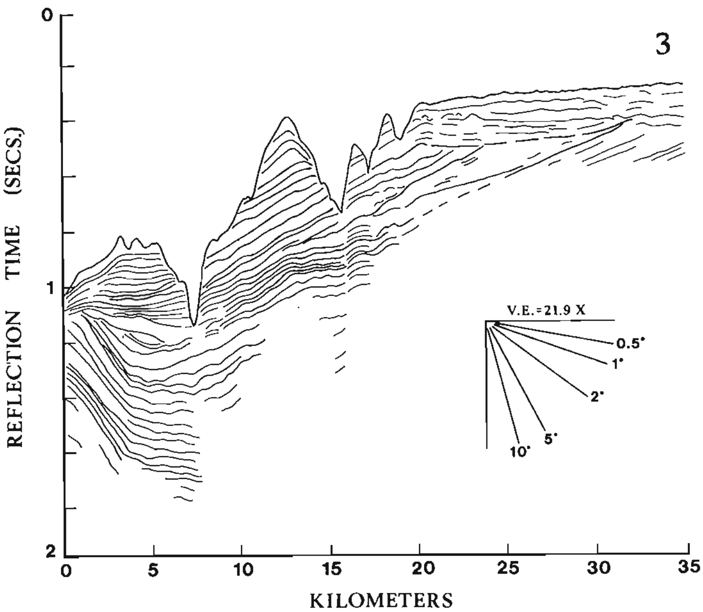

Figure 64.3. Line drawing interpretation of continuous seismic profile number 3 over the continental shelf-slope break in Queen Charlotte Sound, southeast of the Queen Charlotte Islands. An extension of the Sandspit Fault into this area is not evident from the profile.

may suggest either that this sector of Sandspit Fault has not moved in the Quaternary at least, or that the fault location is not in the area of the profile.

Summary

Evidence that the Sandspit Fault has been recently active, or even that it extends southeastward under the waters of Hecate Strait, was not unequivocally obtained during this investigation. A zone of gentle folds with axes parallel to the fault direction affects the assumed Tertiary but not the Quaternary sediments at the western edge of Queen Charlotte Basin east of Moresby Island. The possibility still exists that these folds, which are rarely faulted themselves, are draped over a fault-zone in the underlying basement which was only moderately active during or after deposition of the overlying sediments.

Acknowledgments

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COAL MINING AND SURFACE WATER QUALITY: CROWSNEST PASS, ALBERTA AND BRITISH COLUMBIA – PRELIMINARY DATA

Project 720083

John E. Harrison
Terrain Sciences Division

Introduction

As part of a preliminary evaluation of environmental problems related to coal mine development in the mountains of western Canada, 35 water samples from areas in or adjacent to operational and abandoned coal mines were analyzed. The area sampled was restricted to mining operations in the Crowsnest Pass region of Alberta and British Columbia. Fourteen samples of water flowing from abandoned underground mines contained seven samples in which the water was red as a result of precipitated ferric iron. Twelve small ponds developed in abandoned surface mining pits were sampled. Of these, the water in four appeared to be dirty at the time of sampling. In addition nine streams, which originated within strip mining areas, were sampled; of these, three were clear whereas the remaining six were slightly to very dirty. The fine grained matter responsible for the dirty brown to black colour of this water, in all cases, could be traced back to debris introduced into the stream by landslides triggered by mining operations. The locations from which samples were taken are shown in Figure 65.1.

Data gathered in the field and laboratory are summarized in Table 65.1. Field descriptions, pH determinations, and dissolved oxygen measurements were made on unfiltered samples. The remaining analyses were conducted by the Calgary laboratory of the Inland Waters Branch (Department of Environment) on filtered samples. Only a portion of the data from laboratory tests is presented here.

Flow From Abandoned Underground Works

Of the over 20 sites where water was seen flowing from abandoned underground works, 10 were selected for sampling, and 14 samples were collected for analysis.

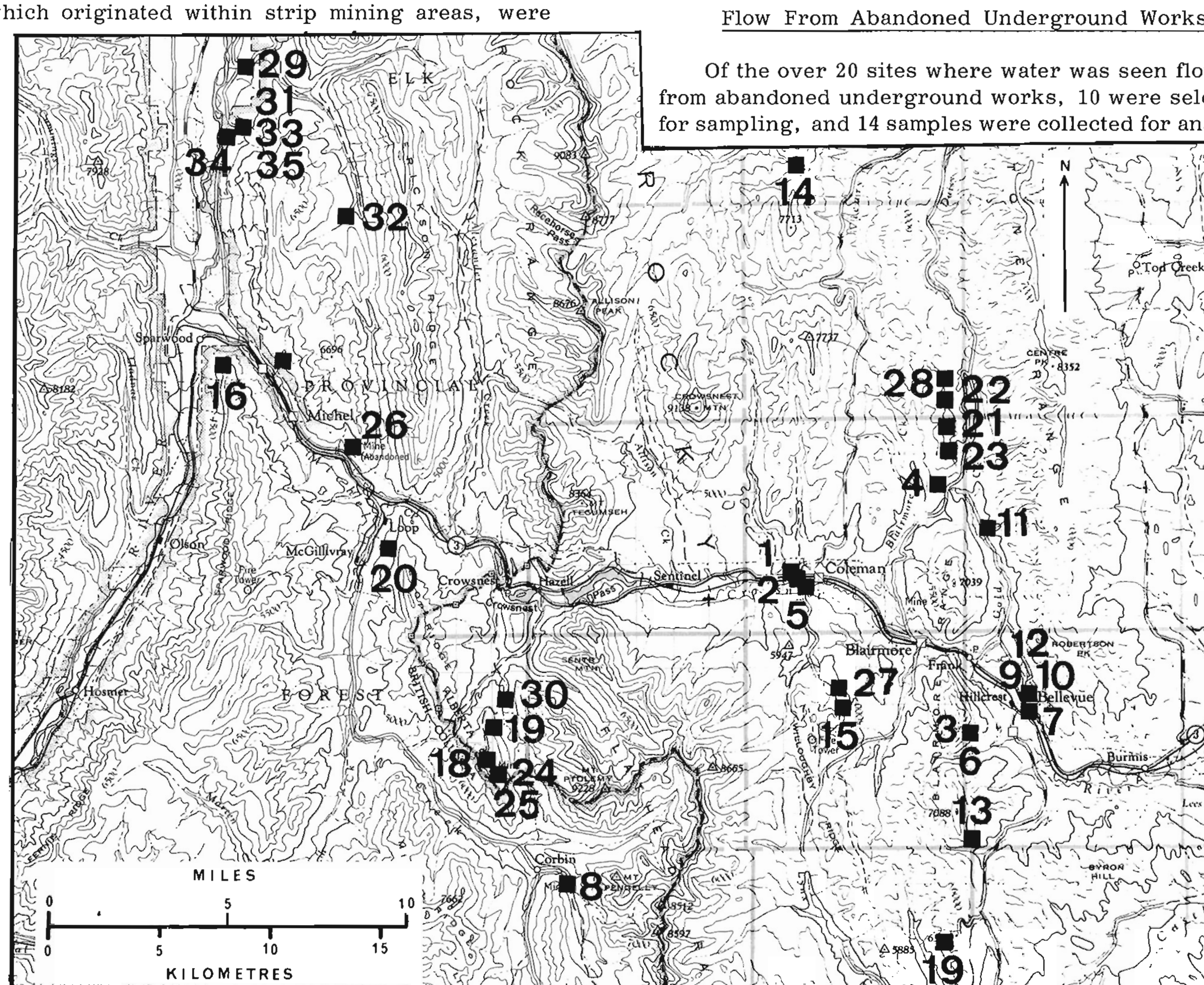

Figure 65.1. Location map showing sample sites in the Crowsnest Pass region of Alberta and British Columbia

From: Report of Activities, Part A;
Geol. Surv. Can., Paper 77-1A (1977)

Table 65.1

Field and Laboratory Data (*see* Fig. 65.1 for sample locations)

Sample No.	Location	Colour of suspended sediment (Field)	pH (Field)	Dissolved O_2 (mg/litre) (Field)	Colour (True Colour Units) (Laboratory)	pH (Laboratory)	Turbidity (Jackson Turbidity Units)	Residue Nonfilterable 550°C (mg/litre)	Residue Fixed Nonfilterable 550°C (mg/litre)	Fe (mg/litre)	Mn (mg/litre)	K (mg/litre)	Specific Conductance (μmho/cm)	Total Alkalinity ($CaCO_3$) (mg/litre)	Total Hardness ($CaCO_3$) (mg/litre)	Ca (mg/litre)	SO_4 (mg/litre)	Si (mg/litre)	Na (mg/litre)	C (Inorganic) (mg/litre)	C (Organic) (mg/litre)	
1	Entrance McGillivray	Green →red	7.8	1.5	5	7.5	88.0	48.8	31.2	56.0	1.550	5.0	1548	138	707	176.0	813.0	7.8	132.0	33	5	RED MINE EFFLUENT
2	McGillivray lower pond	red	7.6	5.7	5	7.7	110.0	-	-	48.0	1.500	5.1	1725	143	714	182.0	821.0	7.9	135.0	32	3	
3	Hillcrest	red	7.6	-	10	7.8	23.0	4.8	0.4	4.6	0.245	2.7	1106	366	425	94.0	237.0	5.2	80.5	102	7	
4	Bois Joli	red	8.2	-	50	8.3	92.0	59.2	42.8	24.0	2.250	1.5	887	162	449	89.5	328.0	3.5	22.8	33	7	
5	Coleman seep	red	-	-	20	7.5	115.0	50.0	26.0	31.0	0.610	7.6	3253	389	658	124.0	1194.0	12.3	450.0	87	11	
6	Hillcrest	red	-	1.4	10	7.8	23.0	48.4	41.2	2.9	0.240	3.2	1203	352	348	68.1	214.0	6.4	108.0	102	6	
7	Bellevue	red	7.7	7.4	20	7.9	43.0	27.2	16.8	6.0	0.310	2.2	1160	280	508	107.0	431.0	4.9	93.2	72	5	
8	Corbim	clear	-	-	0	7.6	0.6	-	-	0.20	0.040	1.8	562	288	302	29.0	33.0	6.3	11.5	32	10	CLEAR MINE EFFLUENT
9	Mohawk east	clear	8.2	-	5	8.2	0.9	-	-	0.05	0.120	1.6	1018	152	484	102.0	427.0	4.4	55.2	32	6	
10	Mohawk west	clear	8.1	-	20	8.1	1.5	-	-	0.35	0.120	1.1	746	330	139	33.8	33.7	6.0	105.0	67	5	
11	Little under-ground	clear	8.6	-	0	7.6	0.5	-	-	0.10	0.115	1.3	783	157	423	92.7	279.0	6.0	10.0	27	3	
12	Mohawk east	clear	-	8.6	5	8.0	1.5	-	-	<0.10	0.120	1.7	1010	164	495	108.0	427.0	5.7	59.5	22	7	
13	Byron Creek	clear	8.2	-	10	8.2	0.5	-	-	0.15	0.365	3.7	1164	376	453	70.1	216.0	4.2	70.3	106	18	
14	Racehorse	clear	-	-	5	6.1	2.9	-	-	<0.1	3.000	1.7	1591	7.1	1054	21.0	1010.0	7.1	7.8	2	2	
15	Lower Lake York Creek north	dirty	-	7.6	20	8.1	25.0	37.5	26.5	0.80	0.045	0.5	154	69.8	77.4	23.0	9.2	4.6	1.7	16	8	DIRTY PONDS IN STRIP PITS
16	Upper Portal Sparwood ridge	dirty	7.2	-	10	8.3	4.0	37.5	<0.1	<0.1	<0.005	3.4	841	119	463	103.0	329.0	0.1	1.5	24	7	
17	Tent Mtn. four pit	dirty	8.6	-	5	-	8.9	34.8	29.6	0.05	0.015	2.3	398	-	-	-	44.1	3.2	0.5	23	9	
18	Tent Mtn. main pit	dirty	9.5(?)	-	5	8.2	26.0	16.8	10.4	0.70	0.140	1.5	505	130	249	48.4	109.0	3.9	0.5	25	10	

Table 65.1 (cont'd.)

19	Adanac strips pond	clear	6.2	-	10	5.0	4.5	5.66	<0.1	0.14	0.100	3.1	78	0.59	27.1	4.4	26.1	1.6	0.5	<2	2	CLEAR PONDS IN STRIP PITS
20	McGillivray strip pond	clear	8.5	-	5	7.9	3.9	3.6	<0.1	0.15	0.040	5.2	1357	139	882	200.0	681.0	1.3	2.7	39	5	
21	Grassy Mtn. Lake No. 2	clear	8.1	-	0	7.9	1.8	-	-	0.05	0.005	1.3	511	146	262	61.4	111.0	1.6	1.0	25	5	
22	Grassy Mtn. Lake No. 4	clear	8.1	-	0	8.2	1.2	-	-	<0.05	<0.005	1.8	552	189	299	51.9	110.0	1.4	3.4	37	7	
23	Grassy Mtn. Lake No. 1	clear	7.9	-	5	8.2	1.3	-	-	<0.004	0.060	1.5	751	125	423	88.6	284.0	2.8	1.2	21	4	
24	Tent Mtn. 2 Pit, east lake	clear	8.1	-	5	8.1	1.1	-	-	<0.10	0.015	1.8	483	150	247	61.0	92.5	1.6	0.5	28	6	
25	Tent Mtn. 2 Pit, west lake	clear	8.0	-	5	7.8	1.8	-	-	<0.10	0.005	1.2	225	70	104	21.8	34.7	1.3	0.5	16	2	
26	Erickson strip pond	clear	-	-	5	8.0	1.3	-	-	0.008	0.005	3.0	615	227	365	72.6	128.0	3.8	1.8	18	12	
27	York Creek north	clear	8.0	-	10	8.1	2.6	-	-	<0.004	0.020	0.6	168	76.4	85.5	28.0	9.7	4.4	1.4	17	3	CLEAR STREAMS FROM STRIP AREAS
28	Grassy Mtn. seep north of central pit	clear	8.0	-	5	8.1	1.3	-	-	0.70	0.025	0.8	433	161	231	54.7	65.7	1.3	0.1	31	2	
29	Grave Creek at CN and RD bridge	clear	8.6	-	5	8.3	1.5	-	-	0.025	0.005	0.6	271	119	146	37.0	22.2	3.9	1.2	12	3	
30	Tent Mtn. stream on northeast side	dirty	7.8	-	20	7.8	19.0	46.4	22.4	0.30	0.015	0.6	122	34	55.8	13.0	24.0	3.8	0.8	9	14	DIRTY STREAMS FROM STRIP AREAS
31	Harmer Knob slide area	very dirty	-	-	40	8.4	150	1551	1370	1.40	0.016	1.5	453	196	237	60.8	43.4	6.7	7.2	27	74	
32	Baldy Creek	dirty	8.6	-	20	8.3	15.0	13.0	<0.1	0.25	0.040	1.2	397	102	181	47.7	13.2	6.0	2.2	13	8	
33	Harmer Knob east side	dirty	8.2	-	20	8.3	33.0	112.0	43.0	0.15	0.055	1.1	391	154	196	47.2	46.7	6.6	4.1	21	47	
34	Harmer Knob slide area	dirty	-	-	-	-	-	408.0	370.0	-	-	-	-	-	-	-	-	-	-	-	-	
35	Harmer Knob slide area	dirty	8.6	-	50	8.1	150.0	838.0	712.0	2.55	0.390	1.3	520	146	177	47.9	44.1	6.9	7.8	33	48	

The most obvious example of mine drainage occurs in the town of Coleman where approximately 5 cfs of greenish water flows from the entrance of the mine and down a baffled sluice. This water flows from the sluice into a settling pond, from the pond through a culvert under the main highway into a second pond, and finally into a small stream. In the course of this journey the water changes from green to bright red to almost clear. Although this red water looks much like the red acidic mine drainage so common in the coal mining areas of the eastern United States, field tests showed it to be slightly basic. Sample 1 was obtained at the mine entrance where the water was still green. Oxygen content determinations were conducted in a field laboratory on samples which were collected and immediately treated to stabilize the oxygen content. Oxygen content of the green water at the mine entrance was 1.5 mg/litre. A second sample (2) from the lower settling pond gave a result of 5.7 mg/litre. A supplementary O_2 determination at the point where the effluent joined the stream provided a reading of 7.6 mg/litre. This last value is comparable to values determined from a small pond and is only 1 to 1.5 mg/litre below levels determined from a nearby small stream. Analysis of the red precipitate from the water shows it to be composed primarily of ferric hydroxide. The following generalized chemical equation is believed to represent the reaction:

$$Fe^{++}(OH)_{0-1}(SO_4)_{1-2} \xrightarrow[O_2]{\text{oxidation}}$$

$$Fe^{+++} \xrightarrow[H_2O]{\text{hydrolysis}} Fe(OH)^3$$

The iron sulphate in the equation is stable and soluble in the absence of oxygen and at a pH of 7 to 8.

Two samples were obtained from red mine waters ponded by a cement wall and apparently originating from a vertical shaft connecting the Hillcrest mine works to the surface. No oxygen determinations were made on sample 3. On a return visit to the site (sample 6) the oxygen content of the water in the pond was 1.4 mg/litre. For comparison, the stream into which this water discharged had an oxygen content of 9.1 mg/litre. Samples 4 and 5 were obtained from very small streams discharging from old works. Sample 7 was taken from a stream discharging 1 to 2 cfs from the entrance of Bellevue mine. Cold, moist air could be felt flowing from this mine entrance indicating a circulation of fresh air through at least a portion of the main tunnel. Not surprisingly, oxygen contents in the water at the mine entrance were almost normal (7.4 mg/litre). The red colour is probably attributable to $Fe(OH)_3$ in suspension.

The presence of the red colour in the mine effluent is directly related to iron content. From Table 65.1 it can be seen that the iron content of the clear water discharged from underground mines (samples 8-14) is one to two orders of magnitude lower than that of the preceding seven samples. One of the critical and unanswered questions is where is the iron coming from? A second question is how is the iron getting into solution in the ferrous state? The only suggestion which can be offered by the author is that the iron is from abandoned machinery and rails and is being dissolved with the aid of anerobic bacteria.

Ponds in Strip Pits

Ponds in strip pits were sampled because it was reasoned that the water in these ponds enters them either through the adjacent coal-bearing rock or by seeping through adjacent spoil piles. In addition, the water has a relatively long period of contact with the rock and debris exposed by mining and therefore has an opportunity to react with this material. In the eastern United States and other areas where high sulphur (pyrite-rich) coal is mined, water ponded by the coal-bearing rock and adjacent debris quickly becomes acid due to the oxidation of pyrite and the formation of sulphuric acid. As can be seen from the data for samples 15 to 26, only one pH reading was below 7. This anomalous value (6.2) probably can be attributed to the fact that the sample was collected from a small pond fed directly by a snowbank. There was little opportunity for the water to come in contact with the surrounding rock. This hypothesis is strengthened by the low total hardness value for the sample.

A number of ponds sampled appeared to contain dirty water at the time of examination. The suspended sediments which gave this dirty appearance probably were washed into the pond by heavy rains eroding the nearby slopes. There does not appear to be any significant difference between the chemistry of clean and dirty ponds, nor does pond water appear to be very different from that of clear mine effluent except with respect to the total amount of dissolved matter.

Streams Draining Surface Mined Areas

In terms of the chemistry of the water there is little difference between streams and the above two categories except for a small decrease in overall dissolved ions. The most important environmental concern with respect to these streams is not the chemistry of the water but rather suspended solids. The single most serious threat to water quality in the vicinity of surface coal mining regions is the introduction of fine suspended material into the streams. The low density of coaly material allows larger sized particles to remain in suspension for a given level of turbulence. These larger particles intercept more light, and thus a smaller amount of fine coaly sediment is required to "dirty" the stream. The black colour of the sediment also contributes to a reduction in the transmission of light through the water in these streams. Further research needs to be done on the transport of fine coal in streams, its effect on organisms in the stream, and more appropriate methods of gauging acceptable levels of stream sediment load.

Conclusions

The data in Table 65.1 indicate that the impact of mining coal on the surrounding water is to increase total dissolved solids, total alkalinity, total hardness, iron, and sulphate contents. In addition, under certain circumstances chemical oxygen demand may be high, and field observations indicate suspended solids represent a serious threat to water quality.

66. PERMAFROST GROWTH AND SUBPERMAFROST PORE WATER EXPULSION, TUKTOYAKTUK PENINSULA, DISTRICT OF MACKENZIE

Project 680047

J. Ross Mackay[1]
Terrain Sciences Division

The mean annual ground temperature in the Tuktoyaktuk Peninsula area, Northwest Territories ranges from about -5°C to -10°C. When a lake becomes drained, therefore, permafrost growth will be initiated on the exposed lake floor. During the past few hundred years many lakes have become drained, usually by erosion of ice wedges at the lake outlet. Holes have been drilled through relatively thin permafrost (< 40 m) in five recently drained lakes and artesian flow was encountered from subpermafrost water in four of the five drained lakes. The fifth lake, which drained in 1972, appears to have a positive subpermafrost pore water pressure, but as yet no artesian flow.

Two sources for the subpermafrost artesian flow can be suggested: (a) a higher artesian head and (b) pore water expulsion beneath aggrading permafrost. An artesian head from a higher recharge area can be disregarded for several reasons. First, as hundreds of sublake taliks (unfrozen zones) perforate permafrost in Tuktoyaktuk Peninsula, artesian pressure beneath a frozen lake bottom would be most unlikely, because the large unfrozen zones beneath the numerous lakes would dissipate artesian pressures a short distance from the source. Second, artesian flow has been observed at sea level (e.g. Pelly Island, N.W.T.) where there is no recharge area.

Numerous laboratory and field studies have shown that pore water may be expelled in advance of a penetrating frost line (Mackay, 1976). The purpose of this paper is to discuss two field experiments designed to study subpermafrost pore water pressures beneath aggrading permafrost in drained lakes.

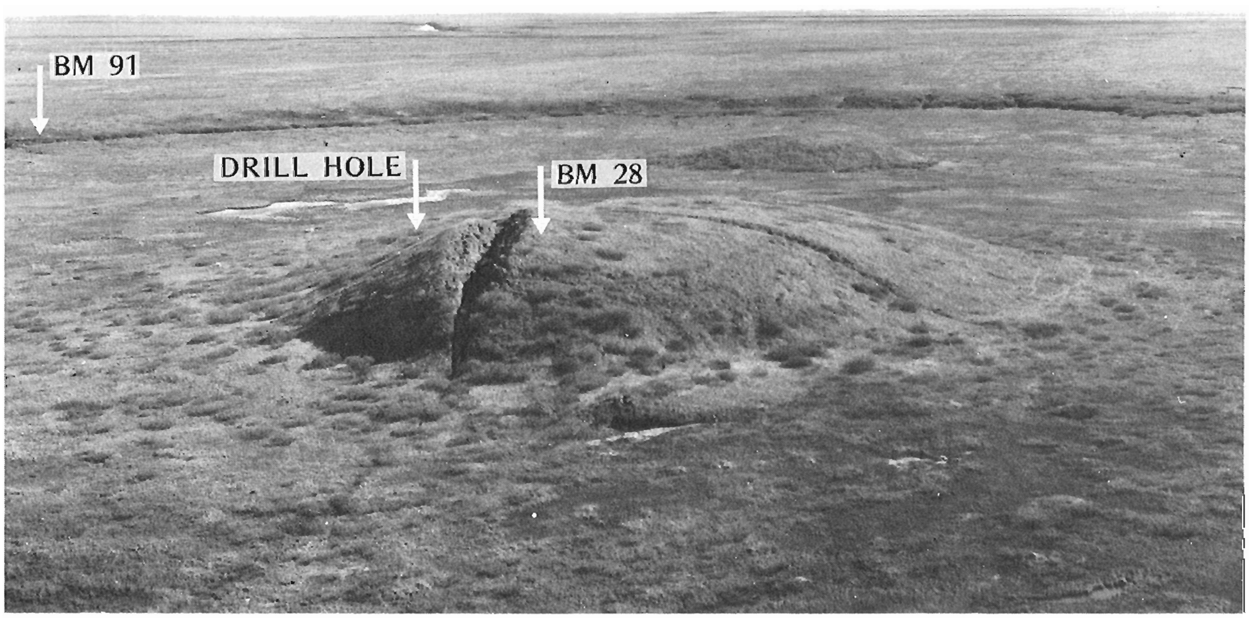

Figure 66.1. A 12-m-high pingo at Site 1. Bench mark (BM) 91 is above the old lake shore; BM 28 is at the top of the pingo; and the locations of the other bench marks are shown in Figure 66.2. The drill hole is on the lower slope of the pingo.

[1]Department of Geography, University of British Columbia, Vancouver, B.C. V6T 1W5

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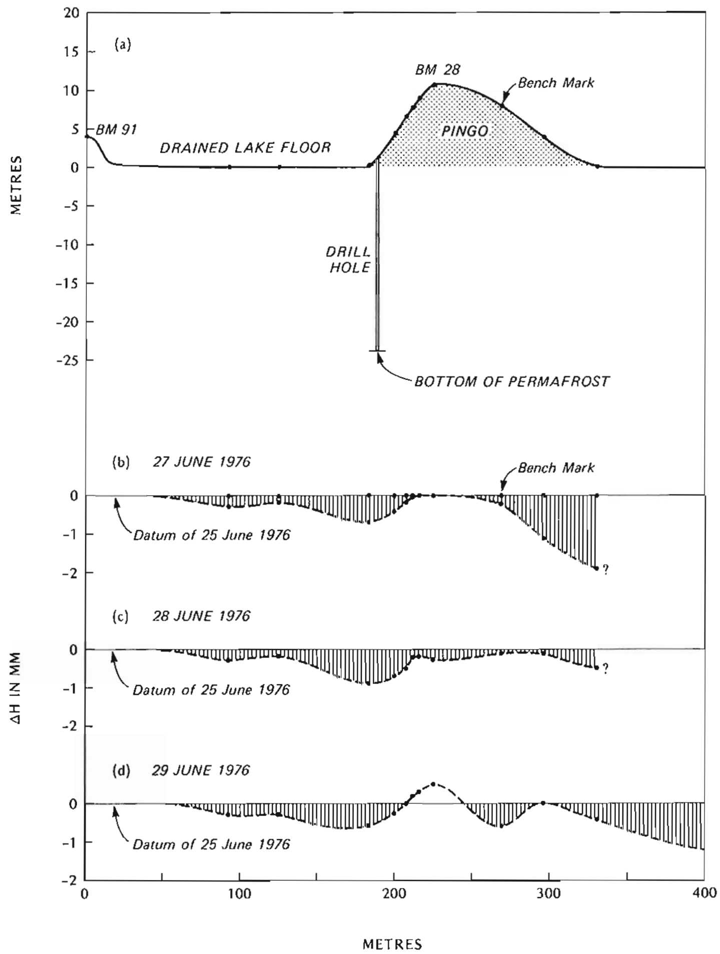

Figure 66.2. (a) Cross-profile of the pingo and drained lake floor; (b), (c), and (d) show altitude changes of the bench marks subsequent to drilling.

Field Experiments

Field experiments were carried out at two drained lake sites where permafrost was known to be aggrading. The experiments involved:

a) precise levelling of bench marks installed into permafrost on the drained lake beds and the pingos which had grown in the lake beds;

b) drilling of holes through the pingos to release artesian flow from the subpermafrost pore water; and

c) resurveying the bench marks after cessation of artesian flow.

Site 1

At Site 1, which is at the east end of Tuktoyaktuk Peninsula (NTS 107 D/15 West; about 69°56'50"N, 129°47'00"W), a pingo has been under precise survey since 1969 (Mackay, 1973, pingo 15). Bench marks have been installed into permafrost on higher land above the former lake floor, on the lake floor, and on the pingo (Figs. 66.1 and 66.2). During the 1969-76 period, the pingo pulsated with periods of uplift and subsidence. In 1973-74, subsidence was observed to occur with spring flow (Mackay, 1975) thus showing that the subpermafrost water was under artesian pressure. The field experiment was designed to see if the pingo and adjacent lake bed could be made to pulsate under artifically induced artesian flow.

On 25 June 1976, the bench marks were surveyed (Wild NA2 with optical micrometer reading to 0.1 mm; invar stave with supporting struts). The next day a water-jet hole was drilled (Judge *et al.*, 1976) on the lower slope of the pingo. Permafrost was penetrated at a depth of 23 m, and an artesian spring (Fig. 66.3) flowed for three days (Fig. 66.4). The flow ceased from infreezing of the drillhole, not from a drop of pressure below ground level. About 2.5 m^3 of sand was brought to the surface and deposited around the spring orifice. Water temperatures decreased daily (Fig. 66.4), but the total dissolved solids increased.

The bench marks were resurveyed on 27, 28, and 29 June 1976. The altitude changes from the predrilling datum of 25 June 1970 are plotted in Figures 66.2b, 66.2c, and 66.2d. Subsidence of the pingo and adjacent lake floor showed up clearly by 27 June and local recovery had commenced by 29 June. Obviously, the subpermafrost pore water pressure must have approached the lithostatic pressure otherwise recovery could not have taken place. The field evidence demonstrates that the subpermafrost pore water pressure was sufficient to uplift the 12-m-high pingo and some of the adjacent lake bed with about 23 m of permafrost.

In addition to the above, an oceanographic pressure transducer was emplaced just below the lower permafrost surface along with three electrodes spaced 25 cm apart, the objectives being to measure subpermafrost pore water pressures and to see if freezing potentials were present. The transducer registered a pressure greater than hydrostatic, when last checked, and there was also a suggestion that a freezing potential could be measured. Work of a similar nature, but broader scope, is planned for 1977.

Site 2

Site 2 is about 16 km east of Tuktoyaktuk, N.W.T. One large pingo (Mackay, 1973, pingo 14) has been under survey from 1971-76. In the summer of 1976, a hole was drilled in the pingo, as at Site 1, but with much more spectacular results. A geyser with an initial fountain of 2.6 m flowed from a 7.5-cm-diameter hole for three days. The water was sediment free, showing the presence of a subpingo reservoir. As at Site 1, flow ceased because of infreezing. The entire pingo lost its 1975-76 growth, the top losing 5 cm, and the drained lake bottom subsided 2 to 3 mm up to a distance of 500 m from the drillhole. A resurvey, a month later, showed that recovery of both the lake floor and pingo were already in progress. Additional

Figure 66. 3.

Spring flow from the drillhole at Site 1. Note the large amount of sand brought to the surface. The drained lake floor is in the background.

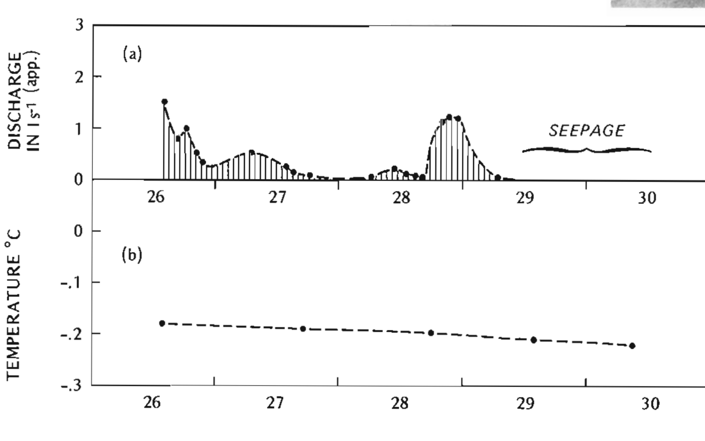

Figure 66. 4. Site 1. (a) approximate spring discharge; (b) water temperatures of the spring.

corroboration comes from previous surveys from 1971-76 and drilling in 1973. The annual growth in 1971-72 and 1973-76 was about 5 cm per year, but after a hole was drilled in April 1973 the pingo showed no growth for 1972-73. As the pingo is about 12 m high and permafrost about 35 m thick on the lake flats, the subpermafrost pore water pressure has been sufficient to lift both pingo and lake bottom, as at Site 1.

Conclusion

Two similar field experiments have shown that artesian flow of subpermafrost pore water from beneath pingos can induce subsidence of both the pingos and the adjacent lake floors. Since there is recovery from subsidence, the subpermafrost pore water pressure must locally exceed the lithostatic pressure. The initial recovery is probably due to the inflow of subpermafrost pore water from around the subsiding area, rather than from a renewal of pressure by freezing and by pore water expulsion, because the time period involved is too brief for sufficient freezing.

It is abundantly clear that free water exists beneath the two pingos; that both the 12-m-high pingos have been uplifted; and that permafrost of 23 m and 35 m thickness adjacent to the pingos has been uplifted. It follows that the piezometric surface, in each site, is far above the top of the respective pingos and far above any possible surface recharge areas.

The field experiments show that the expelled pore water pressure can approach the overburden pressure, as predicted by freezing theory, provided the pore water cannot readily escape. The experimental results are of theoretical interest in the interpretation of ice distribution in permafrost profiles and of practical concern where permafrost growth may generate high pore water pressures.

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Project 680047

J. Ross Mackay[1]
Terrain Sciences Division

The frost table at the bottom of the active layer in a permafrost environment usually is located either by temperature measurement or by probing to a resistant layer which is assumed to be the frost table. The purpose of this note is to discuss some comparative tests on locating the frost table by means of temperature measurement and probing.

Field Procedures

The frost table (i.e. 0°C isotherm) was located with a soil temperature probe (Yellow Springs Instrument Company No. 419 soil probe). The steel probe is pointed, 0.65 cm in diameter, and has a thermistor head at the tip with a stem effect of about 1.5 cm. The soil temperature probe was calibrated before and during use, and temperatures were taken with a portable bridge (Biddle No. 72-430). The soil temperature probe was pushed down in 5 cm increments, allowing ample time for temperature equilibrium, until "firm resistance" just sufficient to bend the hollow steel rock was felt. Steel rods of larger and smaller diameter than the soil temperature probe then were pushed down adjacent to the soil temperature probe to assess the effect of probe diameter. If the soil was clayey, a Geonor shear vane borer (Geonor H-60) was used to measure the undrained shear strength. The general procedure was to repeat the above steps at 10 or 20 cm horizontal intervals across an earth or mud hummock and then to dig out a representative example. About one hundred hummocks in the Tuktoyaktuk Peninsula, Northwest Territories and adjacent areas have been field checked in the summer period for several years.

Field Results

In icy organic soils, ice-bonded coarse grained soils, and fine grained ice-lensed soils, "firm resistance" to probing was typically encountered at the frost table (i.e. 0°C isotherm); but in fine grained soils free of ice lenses, probes of varying diameter usually could be pushed far below the frost table. The general results are summarized below:

Ice lenses and icy peat: Ice lenses and icy peat feel "hard" when probed, and probes "stop" at the frost table. However, small ice lenses, such as those 1 mm or less in thickness, offer little impediment to a probe unless the lenses are closely spaced.

Sand: If sand is ice-bonded, the sand is hard, and the frost table is readily located by probing. This may not be so if the sands are poorly bonded with a low ice content.

Silty clays and clays: Serious errors can arise in probing for the frost table in fine grained soils. Inasmuch as many soils in permafrost are fine grained, the ambiguity should be widespread. To illustrate, on 13 June 1976 the frost table in the centre of a mud hummock, 75 km north of Inuvik, N.W.T., was at a depth of 7 cm, as measured with the soil temperature probe (Fig. 67.1). Nevertheless, the soil temperature probe was pushed without difficulty, to a depth of 30 cm where the temperature was -0.6°C, a depth of 23 cm below the frost table. There was no abrupt resistance change at the 0°C isotherm but a gradual increase in penetration resistance to the probe. A larger diameter steel probe (0.95 cm diameter as compared to the 0.65 cm diameter of the soil temperature probe) also was pushed down to 30 cm. In general, more pressure was required to force a large probe down than a small one, but as thinner probes were used bending became a problem.

Discussion

It is well known that fine grained soils with a large specific surface area may have a substantial amount of unfrozen pore water at and slightly below 0°C. The amount of unfrozen pore water in a soil probably can be estimated from an equation given by Anderson and Tice (1972):

$$\ln w_u = 0.2618 + 0.5519 \ln S - 1.4495\, S^{-0.264} \ln \theta \quad (1)$$

where w_u is the unfrozen water content ($g\ H_2O/g$ soil)

θ is the temperature in °C below zero, and

S is the specific surface area (m^2/g).

The specific surface area of the silty clay of the mud hummock in Figure 67.1 is about 85 m^2/g (analysis by Department of Soil Science, University of British Columbia). Substitution of the specific surface area ($S = 85\ m^2/g$) into equation 1 gives an unfrozen water content of about 40% at -0.1°C, 20% at -0.5°C, and 15% at -0.9°C. In view of the appreciable amount of unfrozen water at slightly negative temperatures, it is not surprising that the soil probe could be pushed well below the frost table shown in Figure 67.1 to where the temperature was -0.9°C. The obvious implication is that if no temperature had been measured, the depth to the frost table would have been overestimated by an amount which would have varied with the observer and the probe.

[1]Department of Geography, University of British Columbia, Vancouver, B.C. V6T 1W5

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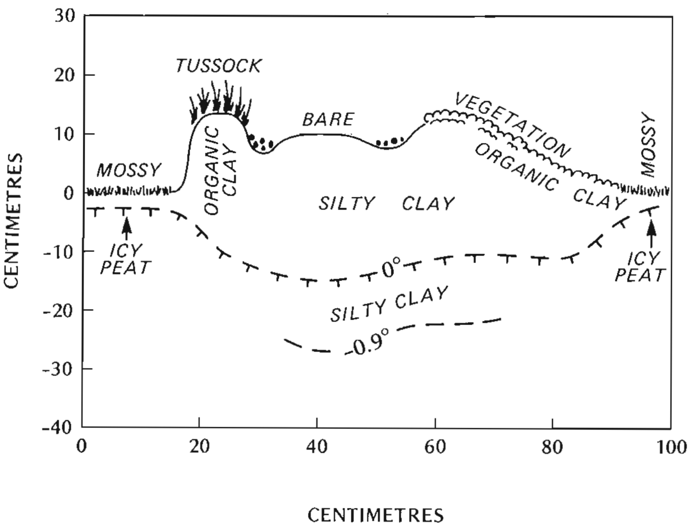

Figure 67.1. Cross-section of a mud hummock, 75 km north of Inuvik, Northwest Territories. The 0°C isotherm was located by measurement with a soil temperature probe and the -0.9°C isotherm by pushing the probe below the frost table until firm resistance was felt.

Conclusion

The frost table (bottom of the active layer) can be located by probing in icy and ice-bonded soils. In fine grained soils with a high specific surface area and an appreciable amount of unfrozen pore water content at and just below 0°C, the penetration resistance to a probe is gradational at 0°C. Consequently probes commonly can be pushed several decimetres below the frost table. Large diameter probes (e.g. 1.25 cm) may give a better "feel" for the frost table than small diameter probes (e.g. 0.50 cm). In early summer when the active layer is thin, a large error can be made in mapping the frost table if a probe is pushed to resistance in soils with a high unfrozen water content. In late summer when the active layer is near the maximum, frictional resistance on the probe in fine grained soils may make probing rather insensitive to the frost table. The penetration depth of a probe below the 0°C isotherm appears to be a measure of the unfrozen pore water content.

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SUMMER SOIL TEMPERATURE AS A FACTOR IN REVEGETATION OF COAL MINE WASTE

Project 720083

John E. Harrison
Terrain Sciences Division

Introduction

During field studies in the Crowsnest Pass area of Alberta and British Columbia (Fig. 68.1) to investigate some geological engineering problems associated with reclamation of coal mine waste dumps, a number of observations indicated that high summer soil temperatures are an important factor in revegetation. Temperature appears to control both the natural pattern of vegetation and the success or failure of natural and man-induced revegetation of disturbed areas. Three main observations prompted interest in soil temperature: a) the presence of a grass-shrub community on south and southwest aspects whereas other aspects of the Interior Douglas fir and Engleman spruce-Alpine fir zones consist of conifer forest; b) the discovery of a number of recently planted tree saplings killed by stem girdling; and c) the lack of natural revegetation on long-abandoned coal waste piles (those facing south and southwest did not even support the sparse vegetation present on some flat and northern exposures).

As revegetation is the ultimate objective of most reclamation efforts, it was considered appropriate to devote a portion of the overall project to gathering data on surface temperatures. Although the information obtained is crude in comparison to energy balance studies, it has application to a number of revegetation problems. This note reports on: a) the composition of the coal waste investigated; b) the variation of surface temperature with time, aspect, slope elevation, and depth; c) moisture conditions; d) thermal properties; and e) the implication of the data with respect to revegetation. It is hoped that the presentation of this data will stimulate further research particularly with respect to species tolerance, amelioration procedures, and the sequencing of plant introduction.

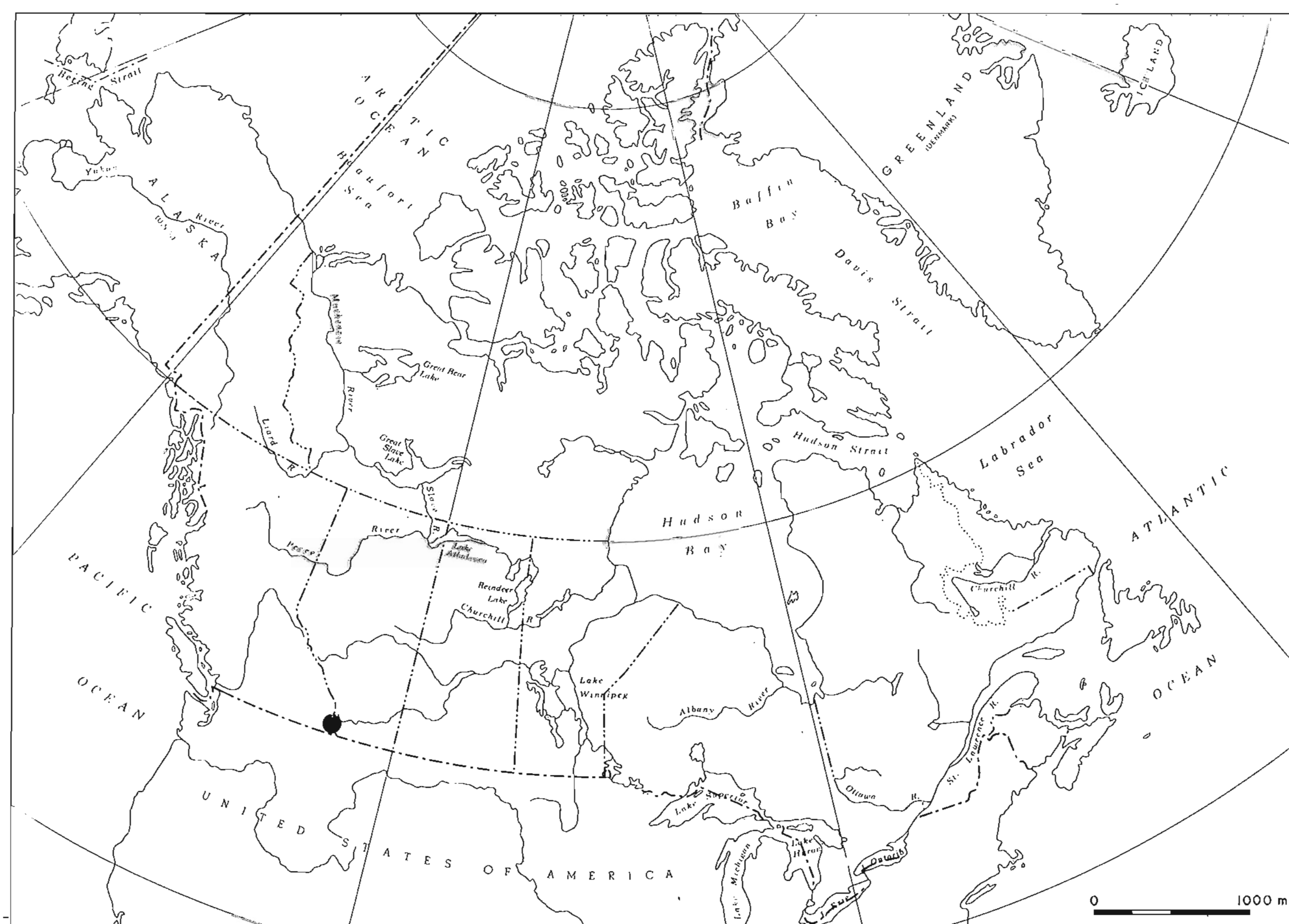

Figure 68. 1. Location map showing study area.

From: Report of Activities, Part A;
Geol. Surv. Can., Paper 77-1A (1977)

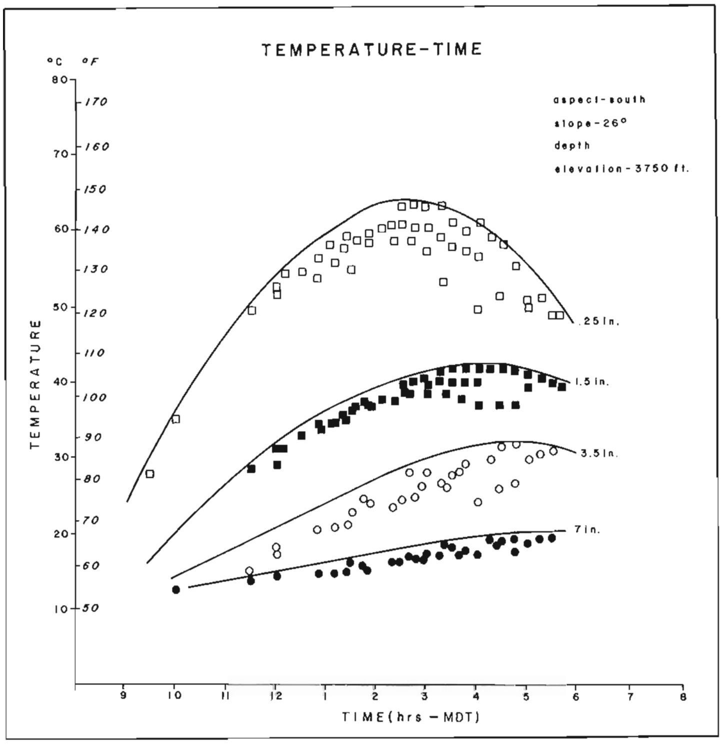

Figure 68. 2. Temperature vs. time plot at four depths for the same slope.

The "soil"

Preliminary measurements on the wide range of materials distributed by surface coal mining indicate that the highest surface temperatures occur in the fine grained, highly carbonaceous waste produced by the cleaning plant. On the assumption that any recommendations derived from investigations of such a "worst case" could be more than adequate for less severe conditions, this material was chosen for extensive testing. Produced as waste from the froth flotation separation of finely ground raw coal, these tailings are piped in a water slurry to large lagoons, where the water is lost by evaporation or seepage and the waste accumulates inside the retaining embankments. Analysis of 12 samples from 2 lagoons yielded an average median grain size of 0. 327 mm (medium sand) with a negligible amount of material exceeding 2 mm (very coarse sand) and 5 to 15% less than 0. 0625 mm (silt). Burning 11 samples revealed an average ash content of 40% by weight with a range from 16 to 73%. Microscopic inspection of both slide-mounted grains and polished sections yielded an average of 56% (recalculated to weight per cent) coal. Soil pH averaged close to 7 with low nutrient levels despite application of fertilizer. In places bulk density of the lagoon material yielded an average value of 48 pounds per cubic foot (0. 77 g/cm^3) with a range of 43 to 57 pounds per cubic foot (0. 69 to 0. 91 g/cm^3) for 12 samples. The average specific gravity of the material is 1. 5.

Temperature Measurements

Four separate experiments were conducted in which temperature changes with time were measured for variations in slope, aspect, elevation, and depth.

Each experiment consisted of temperature readings every 15 minutes for at least 8 hours (0800 to 1700 h) during one or more clear, cloudless, windless days. Although a stretch of 62 days without precipitation during the field season provided many clear days, late afternoon winds generated by differential heating of mountain and valley affected some data gathered after 1500 h.

Temperature was measured using "postage stamp" sized thermistors. These probes consist of a 19 mm square Teflon tape sandwiched over a 2. 5 mm-thick sensor. Ten probes could be read in sequence through a switch box by a portable, direct read, temperature indicator.

Slope and Depth

The experiments relating slope and depth were combined. South-facing slopes of 5, 15, 26, and 35 degrees were measured at soil depths of 0. 25 (60 mm), 1. 5 (3. 8 cm), 3. 5 (9 cm), and 7. 0 (18 cm) inches. No significant difference was discovered between measurements at 7. 0 inches (18 cm) and the various slopes. Data for 3. 5 inches (9 cm) showed slightly warmer values for slopes of 26 degrees than for the other values. At the other two depths, differences in slope for a given depth revealed the 26 degree slope to be the warmest. This was to be expected as the experiments

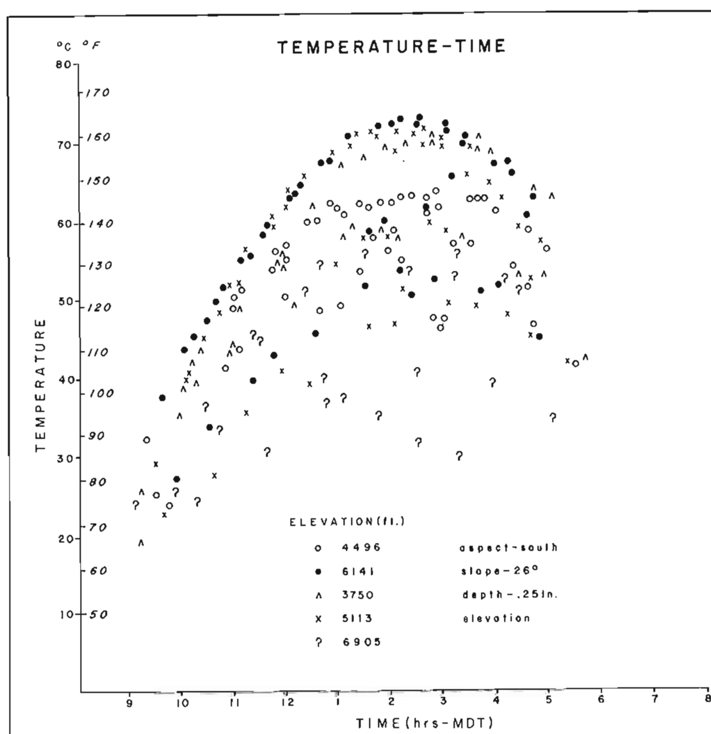

Figure 68. 3. Temperature vs. time plot at five elevations for the same slope.

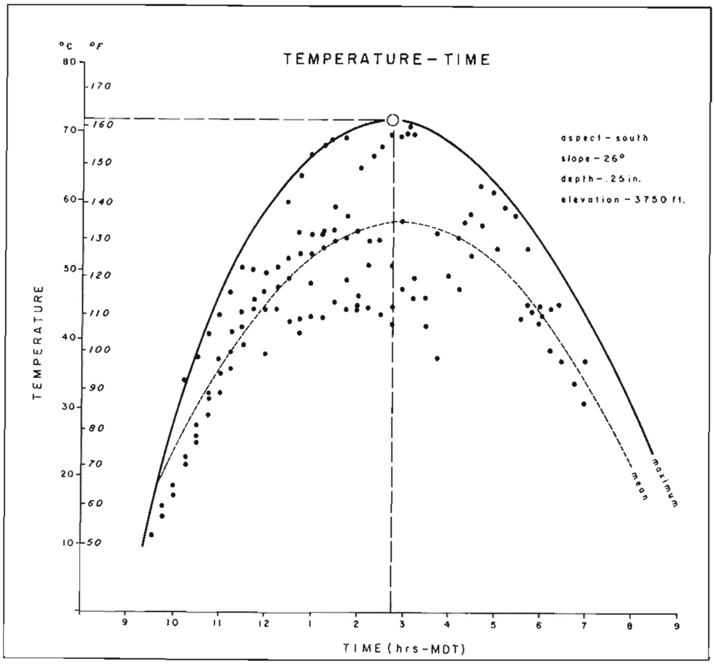

Figure 68. 4. Temperature vs. time plot for south aspect showing maximum temperature.

were run on and just after the summer solstice when the sun is 64 degrees above the horizon at solar noon at that latitude. Therefore the sun's rays at noon are exactly perpendicular to a 26 degree south-facing slope. The individual readings for a south-facing 26 degree slope at an elevation of 3750 feet (1135 m) are shown in Figure 68. 2. Clearly there is a rapid decrease in temperature with depth and a time lag with depth in the attainment of maximum temperature. The solid lines above the data sets are drawn to enclose the upper limit of the data. Since it is the highest temperatures which are of interest, this line is a better generalization of the data than an average value. The assumption is made that without wind disturbance, haze, and scattered cloud, the data points would fall on or slightly below this line.

Elevation

Time-temperature data for elevations 3750, 4496, 5113, 6141, and 6905 feet (1135, 1370, 1560, 1870, and 2105 m) above sea level revealed highest temperatures for 5113 and 6141 foot (1560 and 1870 m) elevations and coolest temperatures at 6905 feet (2105 m). The range of temperatures recorded at the four lower elevations, however, is nearly equal to the range of average values between stations. The only conclusion that can be drawn from these data is that local wind conditions influence surface temperature to a greater extent than elevation. The time-temperature-elevation data are illustrated in Figure 68. 3.

Aspect

Variations of temperature with aspect were measured at the 3750 foot (1135 m) elevation for a slope of 26 degrees and a depth of 0. 25 inches (60 mm); Figure 68. 4 shows the data for the south aspect. The maximum temperature for this aspect occurred at 1445 h mountain daylight time (MDT) and reached 72°C. This value is shown by an open circle. Rather than reproduce eight plots similar to Figure 68. 4, one for each aspect, only the maximum temperature for each data set is plotted in Figure 68. 5. The highest temperatures occur on south aspects at 1445 h MDT, and the lowest maximum temperatures occur on north slopes at approximately the same time. There is a 25°C difference between the temperature on these two slopes. It is interesting to note that southeast and southwest slopes attain about the same temperature, yet field observations show southeast slopes are covered with more trees than similar southwest slopes. Presumably some other stress, perhaps wind, acts in conjunction with temperature to limit tree growth on southwest slopes.

Moisture

Soil moisture, infiltration rate, and capillary rise were measured to determine the source and distribution of water in the test soil. Moisture near the surface will reduce surface temperature by evaporation whereas moisture at a depth of 2 to 36 inches (5 to 90 cm) can be utilized by plants. The amount of water in a soil sample is normally expressed as a per cent of the dry weight of solids. For low density materials this percentage may give a false concept of the actual water available. For example, a test soil with an in place bulk density of 47 lb./cu. ft. (0. 75 g/cm^3) and 10% water content contains 6 gal/cu. ft. (0. 07 ml/cm^3).

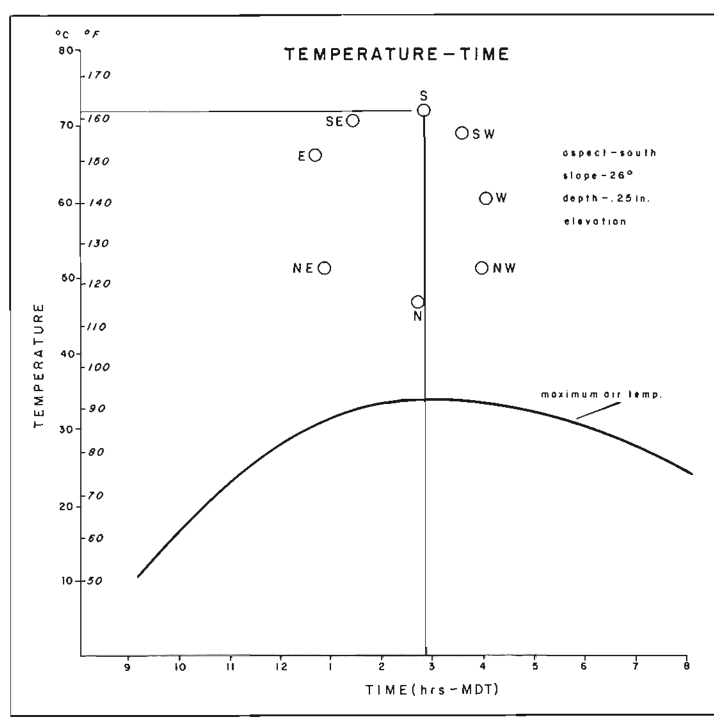

Figure 68. 5. Maximum temperature vs. time plot for eight aspects.

This same amount of water in an average soil (85 lb./ cu. ft., 1.36 g/cm^3) would represent a water content of only 6%.

After two weeks without rain 143 samples were collected at 2 inch (5 cm) intervals from the surface to a depth of 36 inches (91 cm). The average water content for all samples was 10% with values ranging from 0 to 22%. In general the top 1 to 2 inches (2.5 to 5.0 cm) were completely dry with less than 5% moisture down to 6 inches (15 cm). Below this level average moisture values quickly rose to 13% with finer lenses containing up to 22%. Field capacity tests indicated this soil capable of holding up to 26% moisture in the top 6 inches (15 cm). Infiltration rates varied from 2 to 8 in/h (5 to 20 cm/h) depending upon the near-surface grain size. Laboratory studies of capillary rise for material recompacted to field density yielded an average height rise (3 samples) of 38 inches (96 cm). The water table in the waste areas tested was at least 10 feet (3 m) below the surface.

The conclusions drawn from these data plus field observations include:

1. Surface water infiltrates into the soil rapidly.
2. The soil is capable of holding most of the available summer rain in the upper 36 inches (90 cm).
3. Capillary rise from the water table does not supply any water to the upper 36 inches (90 cm).
4. Except in drier coarse grained material, plant roots penetrating below 4 inches (10 cm) probably have at least 10% moisture available.
5. The surface is dry.

In terms of temperature the main conclusion is that once the top 1 to 2 inches (2.5 to 5 cm) dry out no water is available at the soil surface to ameliorate the temperature. Even though sapling roots may be supplied with sufficient water and leaf temperatures are maintained below lethal levels, the stem is damaged at the air soil interface by high soil temperatures.

Thermal Properties

A sample of the lagoon tailings was compacted to field density 47 lb./cu. ft. (0.75 g/cm^3). The thermal conductivity was measured for a water content of 14% and was 0.194 Cal m^{-1}h^{-1} ^{o}K^{-1} yielding a thermal diffusivity of 1.11 x 10^{-3}m^2/h. This is almost identical to the value of thermal diffusivity given for asbestos sheeting. This value indicates that while the surface may become hot, little of the energy penetrates to raise the temperature at depth. The black colour of the material assures that little of the incoming visible radiation will be reflected, thus the absorbed energy raises a very thin layer of the surface to unusually high temperatures.

Revegetation

Revegetation of areas subject to high surface temperatures can only be carried out if some method is found either to protect the plants or to reduce the temperature. The temperature can be reduced by: (a) keeping the surface moist, (b) changing the surface colour with mulch, or (c) establishing a rapidly growing shrub or grass species which will grow to cover the surface before midsummer. Trees and shrubs can be protected by: a) placing slit paper or aluminum plates around their base, b) planting them in areas where natural obstacles provide shade to the stem during midday, or c) providing a nursery crop of grasses or shrubs to shade the surrounding ground. Other less practical methods include increasing soil density by compaction or addition of noncarbonaceous material and providing artificial shade such as snow fence or jute netting.

Field observations indicate that the cheapest and most practical method of dealing with the problem is either to leave the waste with a hummocky surface (up to 1 m) or to create microrelief on originally flat areas. Planting of grasses on this surface in early spring allows establishment before the heat of midsummer. After several seasons, when the grasses are well established, hardy trees and shrubs may be introduced on the north sides of hummocks and in hollows.

Projet 760003

F.J. Morin
Division de la science des terrains

Ce projet de cartographie fut entrepris dans la région de Rivière-du-Loup/Cacouna en suivant le modèle de cartographie géotechnique utilisé systématiquement depuis quelques années par le ministère des Richesses naturelles du Québec dans la vallée du St-Laurent et ailleurs au Québec (Maranda 1975). La cartographie de la région de Rivière-du-Loup s'inscrivait dans le cadre des études éventuelles du Service de Géotechnique du M.R.N.; pour cette raison, ce service consentit à effectuer certains des forages et des essais en place dans la région de l'étude. De plus, une aide très appréciable, sous forme de conseils, fut rendue par les agents de ce service lors de la cueillette d'informations existantes. Sans cette aide, l'efficacité de cette cueillette aurait été très diminuée.

Les limites de la région sous étude s'inscrivent sur 3 cartes au 1: 50 000, soit les cartes 21 N/13, 21 N/14 et 22 C/3 (approximativement entre les latitudes 47°45' et 48°02' et les longitudes 69°15' et 69°38'). Ceci comprend le territoire situé entre Rivière-du-Loup et Isle-Verte. Lee (1972), Dionne (1972) et Dion (1975), parmi les plus récents, ont étudié différents aspects des dépôts meubles de la région. Les dépôts paléozoïques font l'objet d'une étude particulière par Vallières (1974, 1975).

Les réalisations pour la période de ce rapport d'activité se résument à:

- La cueillette d'information existante sur des fiches UGAIS (Bélanger, 1975) pour fins d'entreposage et de traitements subséquents par ordinateur.
- L'étude exploratoire du Quaternaire de la région de Rivière-du-Loup/Cacouna, se basant sur les études déjà disponibles, en vue de se familiariser avec les dépôts de la région et ses problèmes d'interprétations.
- L'accumulation de mesures permettant de dresser une carte des pentes.
- La localisation de sites devant servir à des études ultérieures: forages, sondages séismiques, sondages au pénétromètre statique, sondages scissométriques.
- La détermination de l'épaisseur des dépôts meubles résultant d'une campagne préliminaire de sondages séismiques.
- La localisation des sites d'érosion active et de glissements de terrain récents et anciens.

Les techniciens du ministère des Richesses naturelles du Québec effectuèrent une série de sondages au pénétromètre et de forages jusqu'au roc, avec échantillonnage.

L'étude exploratoire sur le Quaternaire de la région révèle plusieurs problèmes d'interprétations. Ceux qui semblent les plus importants à résoudre, dans le cadre du projet de cartographie géotechnique, se rattachent à la moraine de St-Antonin et à la direction de l'écoulement glaciaire. Ces problèmes traités par Lee (1962) doivent être revus à la lumière des idées nouvelles qui circulent depuis, sur le Quaternaire de l'Est du Québec et de l'étude d'une grande quantité de gravières exploitées depuis. Une étude détaillée de ces gravières s'avère d'autant plus pressante qu'elles sont présentement très exploitées. Cette exploitation s'accélère d'année en année; même quelques beaux sites présents au début de la saison ont disparu quelques mois plus tard.

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NEW OCCURRENCES OF MINERALS IN PARTS OF ONTARIO

Project 640048

Ann P. Sabina
Central Laboratories and Administrative Services Division

Occurrences of several rare and uncommon mineral species were found in the Bancroft-Parry Sound and southern Ontario regions during field investigations in the summers of 1975 and 1976. Some of the species are of rare occurrence, having been reported from only two or three localities in the world. Others occur less rarely but were previously not known to occur at a Canadian locality. New occurrences of minerals that are uncommon in the regions investigated were also encountered. Specimens were collected from current and former mining operations and from numerous road-cuts. The minerals were identified by X-ray powder diffraction (by A. C. Roberts) supplemented in some cases by microprobe analysis (by A. G. Plant). These minerals and their localities are described briefly:

Bastnaesite (La, Ce)F CO_3

This mineral was reported by Satterly (1957) to occur as a rare accessory mineral in the granitic rocks of the Bancroft area; he noted one occurrence: on the property of Consolidated Uranium Corporation, Limited, Chandos Township. Other occurrences were reported by Traill (1970): Blue Rock Cerium Mines Limited property, Monmouth Township, and Bicroft Uranium Mines Limited, Centre Lake and Croft properties. The following are new occurrences:

1. Rare Earth Mining Company Limited, No. 1 shaft, Lot 20, Concession VIII, Monmouth Township (31 D/16) where it occurs as dull dark green to black and brownish grey resinous finely granular massive patches in grey chloritized granite.

2. Road-cut on Highway 503 at a point 0.35 miles from its junction with Highway 121 in Tory Hill (31 D/16). Bastnaesite occurs as orange-red to brownish orange resinous small irregular crusts on crystals of green pyroxene and taffy-coloured plagioclase in white to salmon-pink calcite veins cutting granitic rocks. Associated minerals include apatite, thorianite, thorite, talc, tremolite, titanite and barite.

3. Desmont Mining Corporation Limited property near Wilberforce (31 E/1). Pinkish brown, brown and black resinous to waxy, small granular masses of bastnaesite occur in cream-white to pink massive calcite in association with stillwellite, monazite, uranothorite and a number of other minerals (*see* Stillwellite).

4. Watson feldspar mine, north half of lot 22, concession VI, Monteagle Township (31 F/4W). Dull black laths of bastnaesite were found with betafite, titanite, pyrite, hornblende, pyroxene, pyrite and jarosite in feldspar.

5. Cairns feldspar mine, lot 21, concession VII, Monteagle Township (31 F/4W). Dull black laths of bastnaesite occur in feldspar with cyrtolite, titanite, hornblende, magnetitie, pyrite and jarosite.

6. Nu-World Uranium Mines Limited property in Glamorgan Township near Gooderham (31 D/16). Smooth dull black small masses of bastnaesite occur in granite.

Boehmite Al O (OH)

A widely distributed mineral in bauxite deposits in several parts of the world, boehmite has not previously been reported from a Canadian locality. It was found as a result of this investigation at the Princess sodalite mine near Bancroft (31 F/4W). It occurs sparingly as cream-white pearly, platy and fluffy granular aggregates partly filling small vugs (2 to 5 mm in diameter) in massive natrolite, as pearly white powdery to finely granular aggregates on platy to columnar natrolite which commonly lines vugs, and as silky white wispy fibres on massive natrolite. In some vugs, nordstrandite and another mineral (possibly gibbsite) are commonly associated with boehmite. Other minerals occurring in natrolite are listed under Dawsonite.

Brugnatellite $Mg_6FeCO_3(OH)_{13} \cdot 4H_2O$

This rare carbonate has previously been reported to occur in Colorado and at a few localities in Italy. It was found at two localities near Bancroft:

1. York River tactite zone, a cliff-side exposure on the east side of the York River, 0.2 miles north of Highway 500 at a point 6.9 miles east of Brancroft (31 F/4E). Brugnatellite occurs as cream-white silky to waxy, scaly to compact granular nodules streaked with charcoal grey brucite. The nodules occur in dolomitic marble overlying alkaline gneisses. Other minerals occurring as crystals, nodules or grains in the marble, as reported by Hogarth *et al.* and as a result of this investigation include: light to dark brown and charcoal grey brucite; white, yellow, green and brown serpentine; pink, orange and brown grossular garnet; yellow-green and brown vesuvianite, yellow to orange clinohumite; colourless and pink olivine; green diopside; white wollastonite; light green and colourless tremolite; dark green spinel; colourless cancrinite; white scapolite; pink zircon; blue calcite; colourless to greyish yellow monticellite; amber aragonite; white hydromagnesite; and chlorite, plagioclase, pyrrhotite, graphite and magnetite. Sodalite is also present (L. Moyd, pers. comm.).

From: Report of Activities, Part A;
Geol. Surv. Can., Paper 77-1A (1977).

2. Road-cuts on Highway 500 on the east side of the York River bridge at a point 6.9 miles east of the junction of highways 500 and 62 in Bancroft (31 F/4E). Nodules of brugnatellite similar to those occurring at the York River tactite zone occur in crystalline limestone exposed by these road-cuts. The nodules are associated with small crystals or grains of black brucite, yellow to orange clinohumite, dark green spinel, green serpentine, smoky violet olivine, amber mica, pyrrhotite and graphite.

Coffinite $U(SiO_4)_{1-x}(OH)_{4x}$

Widely associated with uranium and vanadium mineralization in western and southern United States, coffinite has been reported from two occurrences in Canada: the Blind River area in Ontario (Traill, 1970) and the Beaverlodge area in Saskatchewan (Traill, 1974). It was found during this investigation at the Besner feldspar mine located near Britt in the Parry Sound district (41 H/15E). It occurs as dull black masses in association with allanite, garnet, apatite, magnetite and chlorite in microcline pegmatite.

Dawsonite $NaAl(CO_3)(OH)_2$

First described in 1874 from a Montreal locality, reported occurrences in Canada of this mineral have been confined to the Monteregian intrusive complex of the Montreal region, although it has been reported from several foreign localities. This investigation yielded an occurrence of dawsonite, at the Princess sodalite mine near Bancroft (31 F/4W). It occurs as a colourless to white silky fibres (averaging 1 mm long) forming cross-fibre veinlets and radiating, stellate, parallel and felted aggregates, and as colourless transparent striated masses with a bright vitreous lustre (Fig. 70.1). These masses and aggregates measure up to one centimetre long and are embedded in mottled white to pink and orange red massive natrolite that occurs abundantly with sodalite. Associated with natrolite are plagioclase, potash feldspar, biotite, calcite and less commonly, analcime, apatite, zircon, chlorite, pyrite, magnetite and hematite. Small vugs measuring 2 to 5 mm in diameter in the natrolite are commonly lined with acicular aggregates of natrolite and, rarely with nordstrandite crystals and with fluffy granular aggregates of boehmite. The specimens were collected from the new open cut in the east side of the original quarry through the courtesy of Mr. Paul Rasmussen, owner and operator.

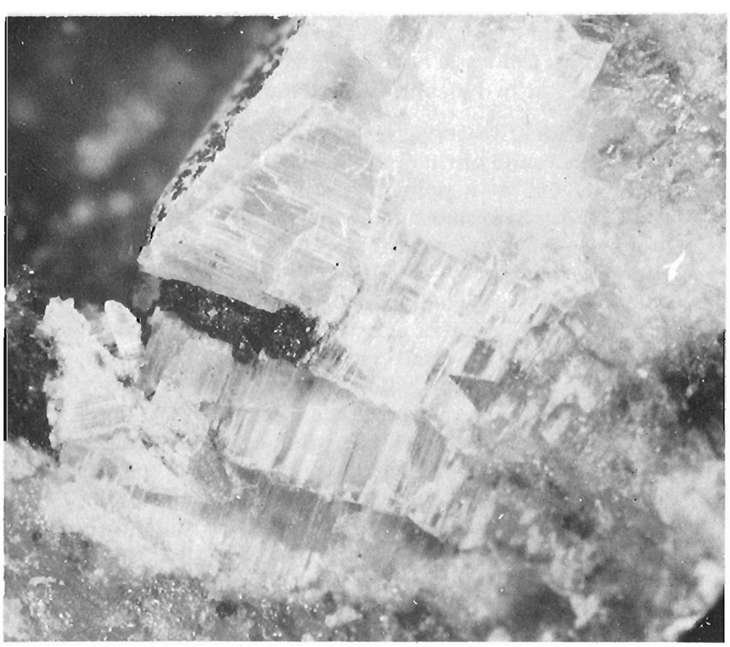

Figure 70.1. Striated massive dawsonite in natrolite, Princess sodalite mine, Bancroft. (X25, GSC 203093-Q).

Diaspore $AlO(OH)$

Diaspore was identified in a specimen collected from a small pit on the north side of the Hartley Bay Road at a point 3.1 miles west of its junciton with Highway 69 near Bigwood (41 I/2) in the French River district, about 60 miles north of Parry Sound. The diaspore occurs as a partial replacement of corundum prisms embedded in white biotite syenite. The prisms are transparent and bluish to purplish grey in colour. Pink zircon, colourless mica and graphite also occur in the syenite. In Canada diaspore has previously been reported from the Lorrain quartzites of Ontario and from the pyrophyllite deposit in the Avalon Peninsula, Newfoundland.

Fluoborite $Mg_3(BO_3)(F, OH)_3$

This rare borate was found at four localities in the Bancroft-Haliburton region, the first occurrences reported in Canada. In these occurrences, it is characteristically associated with a mineral of the humite group in crystalline limestone. It occurs as crystals resembling apatite and as granular masses.

1. Road-cut on the south side of the Crystal Lake Road at a point 6.7 miles east of its junction with Highway 121 which in turn is 4.4 miles south of Kinmount (31 D/9W). Dolomitic crystalline limestone containing fluoborite is in contact with red granite pegmatite. Fluoborite occurs as colourless to white, vitreous to chalky hexagonal prisms measuring up to 5 mm long; they occur individually and as parallel, divergent or stellate groups embedded in the limestone (Fig. 70.2). It is also present as colourless transparent to white granular patches and as silky white friable small masses in the limestone. Associated minerals include: orange norbergite, colourless to light green tremolite, grey and green serpentine, violet and amber fluorite, amber mica, rutile, graphite and pyrrhotite.

2. Cardiff Uranium Mines Limited property near Wilberforce (31 E/1). Light peach-coloured transparent individual crystals of fluoborite occur in crystalline limestone containing orange norbergite, amber mica,

Figure 70.2. Fluoborite crystal (actual size is 4 mm long) in crystalline limestone, Crystal Lake Road occurrence. This crystal grades from colourless, transparent at one end to chalk-white at the other. (GSC 203093-N).

blue apatite, colourless to grey tremolite, green serpentine, white and light blue talc, peach-coloured tourmaline, colourless to white fluorite, brownish pink titanite, light green diopside, white to amber sepiolite, graphite and pyrite. Specimens were collected from the mine dump adjacent to the shaft in the South zone.

3. Road-cuts on Highway 35 at a point 7.7 miles south of its junction with Highway 121 at Minden (31 D/15). (A cottage road leads west from Highway 35 at this point.) Light peach-coloured transparent granular patches of fluoborite occur sparingly with grains of orange chondrodite, amber mica, green tremolite, blue apatite, green serpentine, colourless fluorite, pink zircon, pyrite and magnetite in dolomitic crystalline limestone.

4. Road-cut on Highway 648 at a point 0.2 mile south of the junction of the Dark Lake Road north of Wilberforce (31 E/1). White powdery patches of fluoborite occur sparingly with orange chondrodite, colourless to light yellow, green and grey tremolite, green serpentine, amber mica, blue apatite, pyrite and graphite in crystalline limestone.

Gunningite $(Zn, Mn)SO_4 . H_2O$

This rare secondary zinc mineral originally described in 1960 from the Keno Hill (Yukon) silver-lead-zinc deposit, it was found at the Canada Crushed Stone quarry, Steetley Industries Limited near Dundas (30 M/5W). It occurs as greyish white efflorescences on sphalerite-pyrite-galena assemblages in a sulphide zone in limestone exposed on the west wall of the south end of the quarry along the ramp leading from Moxley Road. Goethite, rozenite and sulphur are associated with it.

Hydromagnesite $Mg_5(CO_3)_4(OH)_2 . 4H_2O$

This mineral occurs with brugnatellite forming cream-white waxy nodules in calcite at the York River tactite zone near Bancroft (31 F/4E). These nodules are associated with nodules and grains of brucite and serpentine; a list of other minerals occurring at this locality is given under Brugnatellite.

Hydrozincite $Zn_5(CO_3)_2(OH)_6$

A secondary mineral formed by the alteration of sphalerite, hydrozincite was found at the Albermarle zinc mine in the Bruce Peninsula north of Wiarton (41 A/4). It occurs as white to yellowish white powdery encrustations on yellow, amber to dark brown sphalerite, and as white waxy patches on the dolomite host rock.

Monazite $(Ce, La, Th)PO_4$

A widely disseminated accessory mineral in granitic and pegmatitic rocks, monazite was found sparsely distributed in calcite at the property of Desmont Mining Corporation Limited near Wilberforce (31 E/1). It occurs as yellow, amber and orange transparent to translucent, vitreous to resinous finely granular masses embedded in cream-white to salmon-pink massive calcite. It is associated with stillwellite, bastnaesite and number of other minerals listed under Stillwellite.

Monticellite $CaMgSiO_4$

This mineral was found in whtie sugary calcite at the York River tactite zone near Bancroft (31 F/4E). It occurs as colourless and greenish to yellowish grey transparent grains associated with grains of olive green transparent diopside, green to brown serpentine, bluish green mica and light blue calcite. A list of minerals occurring at this locality is given under Brugnatellite.

Nordstrandite $Al(OH)_3$

This rare mineral has previously been reported from localities in West Sarawak, Borneo (Wall *et al.*), Guam (Hathaway and Schlanger) and from New South Wales, Australia (Goldbery and Loughnan). It was found during this investigation at the Princess sodalite mine near Bancroft (31 F/4W) where it occurs as colourless transparent pointed blade-like crystals protruding from walls of vugs (measuring up to 5 mm in diameter) in massive natrolite (Fig. 70.3); as nests of colourless crystals embedded in natrolite as white fibrous patches on natrolite; and as colourless to white striated masses (about 3 mm long) in natrolite. The individual crystals measure up to 1 mm long. In vugs the crystals occur alone or with colourless acicular or pearly white tabular to columnar aggregates of natrolite.

Oligoclase and Quartz

Oligoclase exhibiting very coarse twinning striations, and quartz with a pseudo-cleavage occur in pegmatite at the McKenzie Lake (Gunter) feldspar quarry on the McKenzie Lake South Road at a point 4.6 miles from its junction with Highway 127 which in turn is about 10½ miles north of Maynooth (31 E/8E). The oligoclase is greenish white, bluish and greyish green and charcoal grey, transparent glassy to translucent. The quartz varies from colourless glassy to white, and reddish due to iron-oxide staining. Pink potash feldspar is also present in the pegmatite. Accessory minerals include biotite, pyroxene, pyrite, scapolite, titanite and calcite.

Sepiolite $Mg_4Si_6O_{15}(OH)_2 \cdot 6H_2O$

Sepiolite (meerschaum), an alteration product of serpentine and magnesite, occurs at several localities outside of Canada, and was found during this investigation at the Cardiff Uranium Mines Limited property near Wilberforce (31 E/1). It occurs as white to light amber silky matted fibres forming patchy encrustations on crystalline limestone containing amber mica, tremolite graphite and pyrrhotite. The specimens were collected from the mine dumps adjacent to the shaft in the South zone.

Sinhalite $MgAlBO_4$

This mineral was originally described in 1952 (Claringbull and Hey) as a result of an investigation of several light yellow to brown gemstones labelled as olivine in the collections of the British Museum and the Geological Survey Museum in London. Some of the gemstones originated in Ceylon, others were of unknown origin. Sinhalite has since been reported from New York, Russia, Burma and Tanzania. This investigation produced another occurrence: in a road-cut on the north side of the South Baptiste Lake Road at a point 7.5 miles west of its junction with Highway 62, which in turn is 4.2 miles north of Bancroft (31 E/1). At this occurrence, sinhalite forms colourless transparent sugary crusts on dark green transparent massive spinel. These minerals occur in crystalline limestone along with a number of other minerals listed under Warwickite.

Stillwellite $(Ce, La, Ca)BSiO_5$

Stillwellite , was first described in 1955 (McAndrew and Scott) from Queensland, Australia, and has since been reported from the Langesundfjord district in Norway. A new occurrence was found near Wilberforce, on the property of Desmont Mining Corporation Limited (31 E/1). Stillwellite occurs at this locality as maroon-red opaque, waxy tabular crystals measuring up to 5 mm by 4 mm, and as grey, pink maroon-red and reddish brown, waxy to resinous, smooth trextured massive patches in cream-white to salmon pink massive calcite (Fig. 70.4). Emerald green diopside and amber mica are common in the calcite. Associated minerals include monazite, bastnaesite, thorianite, thorite, uranothorite, titanite, tourmaline, serpentine, scapolite, apatite, quartz, plagioclase, potash feldspar, actinolite, marcasite, pyrite, molybdenite, magnetite and goethite. The calcite occurs in veins and lenses in sugary diopside-calcite rock enclosed in marble.

Sulphur S

Sulphur, as an oxidation product of pyrite, was found at two localities: at the Canada Crushed Stone quarry, Steetley Industries Limited near Dundas (30 M/5W), and at the Desmont Mining Corporation Limited property near Wilberforce (31 E/1). At the Dundas locality, it occurs as a pale yellow to rusty yellow powder on pyrite-galena-sphalerite assemblages in a sulphide zone in limestone (*see* Gunningite). At the second occurrence, sulphur is admixed with pyrite to form black loosely granular masses in small pockets in calcite-diopside rock.

Szaibelyite (Camsellite) $MgBO_2(OH)$

This rare borate was found in crystalline limestone exposed by a road-cut on the north side of the South Baptiste Lake Road 7.5 miles west of its junction with Highway 62 north of Bancroft (31 E/1). It has previously been reported from two Canadian localities: Douglas Lake, British Columbia, and Bryson, Quebec. At the Bancroft occurrence, it occurs sparsely as buff-coloured silky to earthy, finely granular to woody crusts, and as nodules associated with amber mica and several other minerals which are listed under Warwickite.

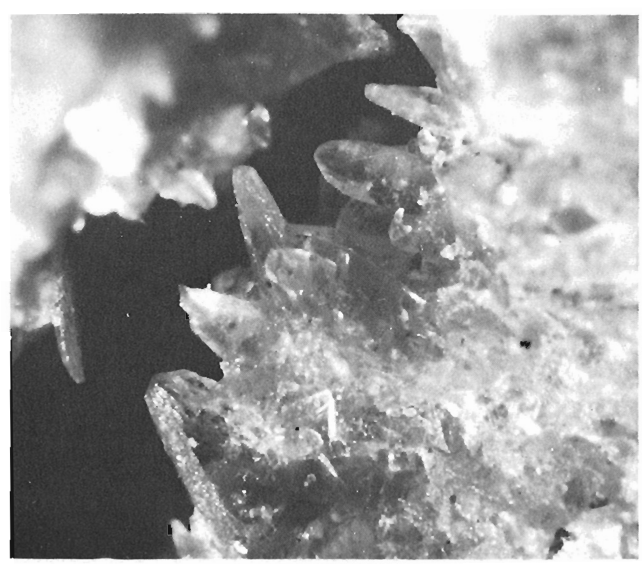

Figure 70.3. Nordstrandite crystals protruding from walls of vug in natrolite, Princess sodalite mine. The crystals average 0.5 mm long. (GSC 2023093-G).

Figure 70.4. Stillwellite, partly capped with monazite, in calcite, Desmont Mining Corporation, Limited property, Wilberforce. (X10, GSC 203092-Z).

Warwickite $(Mg, Ti, Fe^{3}, Al)_2 (BO_3)O$

Until recently this borate was known to occur only at its type locality in Warwick, New York. Recent discoveries were made in Korea, Japan, Russia and as a result of this investigation, in the Bancroft area where it occurs in crystalline limestone exposed by a road-cut on the South Baptiste Lake Road (*see* Sinhalite for exact locality). It is disseminated as small finely granular masses measuring 2 to 5 mm in diameter, and as slender prisms averaging 2 mm long. It is black with a dull, submetallic or pearly lustre, a grey or red tint on the surfaces, and reddish black streak. Minerals occurring with warwickite are: amber mica, dolomite, grey to light green and light brown tremolite, amber and orange to dark brown tourmaline, buff-coloured szaibelyite, green serpentine, light blue apatite, dark green spinel, colourless sinhalite, grey to brown anatase, light green and yellow fluorite, yellow to orange chondrodite, yellow scapolite, ilmenite, marcasite, pyrrhotite, pyrite, graphite and goethite.

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IGNEOUS AND METAMORPHIC ROCKS BETWEEN ROCKY BAY AND RAGGED HARBOUR, NORTHEASTERN NEWFOUNDLAND

Project 730044

K. L. Currie and G. E. Pajari, Jr. [1]
Regional and Economic Geology Division

Introduction

Williams (1964) observed that the Appalachian orogen forms a symmetrical two-sided system, bounded on both sides by crystalline basement. The western side of the orogen in Newfoundland, extensively studied and modelled since that pioneering paper, represents one of the most convincing demonstrations of plate tectonic theory (Dewey and Bird, 1971; Williams *et al.*, 1974). By contrast, the eastern margin remains imperfectly understood. Kennedy and McGonigal (1972), Kennedy and Blackwood (1974) and Blackwood and Kennedy (1975) proposed that the rocks bounding the eastern side of the central mobile belt could be divided into six major categories. They consider the mid-Ordovician (Caradoc) "Davidsville Group" to lie unconformably upon the "Gander Group" which in turn lies unconformably upon gneisses, presumed to be of Precambrian age. They divide the plutonic rocks into three categories, the oldest of which, the megacrystic granites, are supposed to intrude only basement gneisses, while "leucocratic granites" intrude both basement and "Gander Group", and tonalites and quartz diorites intrude the "Davidsville Group". This model provides a convenient framework for much of the geology of northeastern Newfoundland, but its conclusions have been attacked on both terminological and substantive grounds (Brückner, 1972; Jenness, 1972). According to maps and descriptions by Kennedy and co-workers, all six of the major rock categories occur within a small area of unusually good outcrop between Rocky Bay and Ragged Harbour. We have mapped this difficult and physically punishing area in some detail, in order to define relations between various units.

Description of units

We found it appropriate to divide the rocks into seven major units, four of supracrustal origin and three plutonic complexes. Since we have nowhere recognized unconformities between supracrustal units, or unambiguous basement, we attach the non-stratigraphic and non-genetic descriptor "sequence" to mappable assemblages of supracrustal rocks.

The Flinn's Tickle sequence (unit 1), well exposed within the village of Musgrave Harbour and for 5 km to the south, consists of a lower unit of finely banded mesocratic to melanocratic tonalite gneiss with numerous quartzite interbeds, and an upper unit rich in amphibolitic and quartzitic gneisses. As noted by Kennedy and Blackwood (1974), the northern part of the gneisses contains many disoriented boudins or xenoliths, some of which exhibit internal folding or deformation not seen in the matrix. A similar phenomenon occurs in Gander sequence rocks south of Aspen Cove. In our opinion, the xenoliths demonstrate a high degree of ductility or mobilization of the host at some point in its history. The main foliation and minor folds within the Flinn's Tickle sequence dip and plunge gently away from the Ragged Harbour complex (units 2-15). We interpret this to mean that the latest movement of the latter domed the Flinn's Tickle sequence, presumably already complexly deformed and metamorphosed.

The river sequence (unit 2) outcrops superbly in chutes along Ragged Harbour River and very poorly elsewhere. The rocks consist mainly of cleaved quartz-biotite-plagioclase rocks of slaty or phyllitic aspect, with lesser amounts of greenstone and some conformable layers of mylonite of granitic composition. The rocks exhibit a deceptively simple fabric with only one prominent cleavage in most specimens. Complex, polyphase small folding can be detected in a few specimens and is suspected in others. The strong cleavage dips gently away from the southern margin of the Ragged Harbour complex, so that the River sequence occupies the same structural position as the Flinn's Tickle sequence. The compositions of the two units are not grossly different, despite the striking difference in appearance, and they may be correlative.

We define the Gander sequence (units 3, 4) to be generally semipelitic rocks containing orange to pink psammitic beds from 0.5 to 10 cm thick. This criterion is mappable and persistent, whereas other criteria for distinguishing Gander from Davidsville rocks break down. In this region the Davidsville generally shows relict sedimentary features and simple structure, whereas Gander rocks commonly lack sedimentary features and exhibit complex structure. These criteria collapse when the Davidsville is traced into areas of high metamorphic grade, where sedimentary features disappear and the structure becomes complex. Our criterion produces a mappable, consistent stratigraphy, according to which the Gander-Davidsville contact is conformable where exposed. Thin pelitic units (unit 3) within the Gander sequence, commonly contain very high andalusite contents. Southeast of Aspen Cove, Gander sequence rocks contain diffuse granitoid masses, ranging from boudin fillings to reticulate networks. From their character and development, these masses are interpreted as anatectic melt.

Mafic volcanic rocks commonly form the base of the Davidsville sequence (units 6-11), although fissile slate or schist may be present. The volcanic rocks (unit 6) consist mainly of coarse agglomerate basalt

[1] University of New Brunswick, Fredericton, New New Brunswick E3B 5A3.

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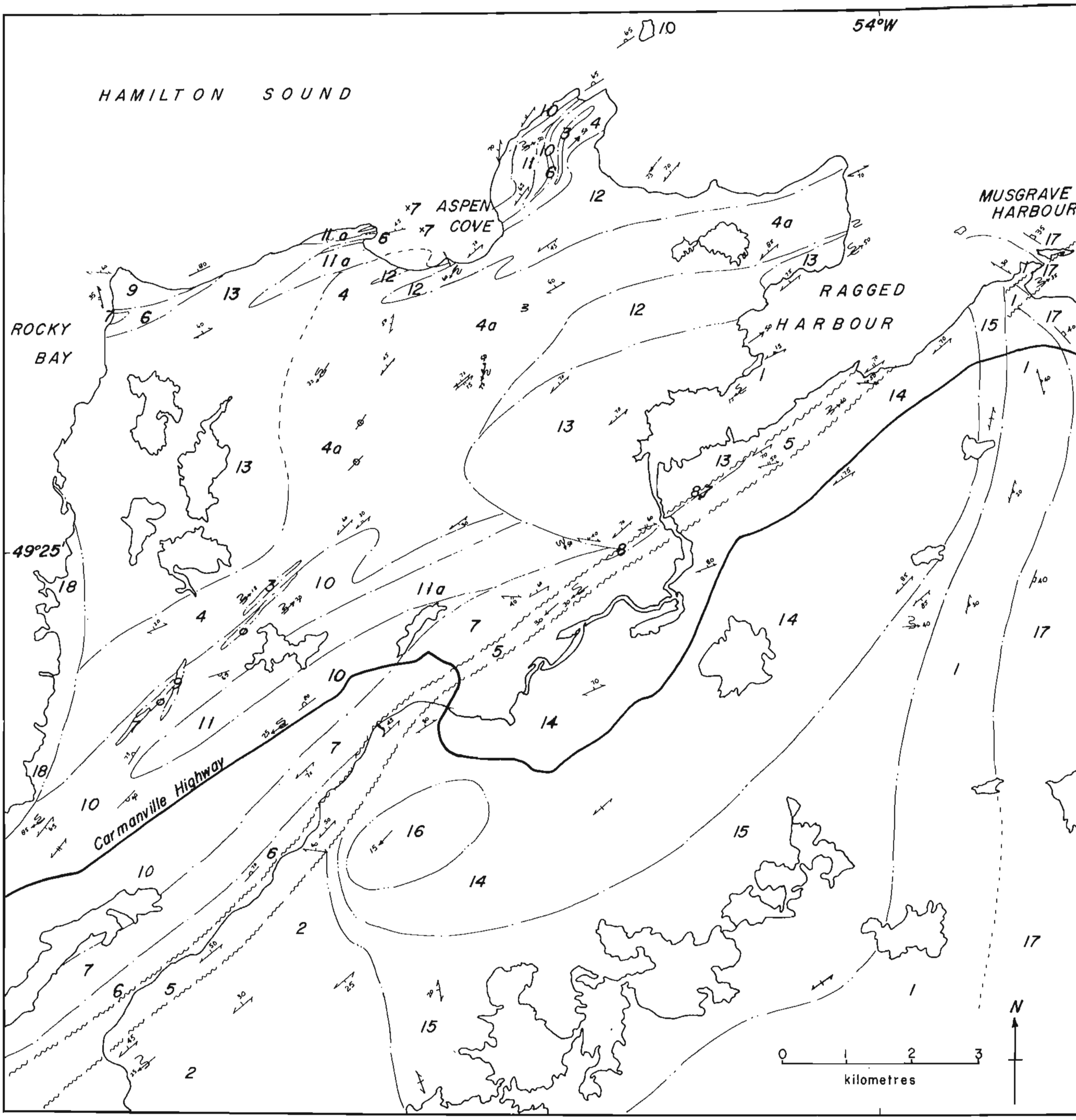

Figure 71. 1. Geological sketch of the area between Rocky Bay and Ragged Harbour, northeastern Newfoundland.

containing vesicular fragments with dense rims. South of the Carmanville highway, these agglomerates locally contain rounded fragments of gneissic tonalitic rocks together with blocks of ultramafic rocks. In this region there appears to be a gradation from mafic volcanics containing many gabbroic sheets and dykes to ultramafic and gabbroic rocks of the Gander River Ultramafic Belt (units 7, 7). North of the highway ultramafic rocks occur in similar stratigraphic position only as isolated "pips" of brecciated material, and the gabbro sills and dykes occur mainly within Gander sequence rocks. The upper parts of the Davidsville sequence consist of a monotonous repetition of black pelitic beds and grey greywacke beds (units 10, 11). Minor pebble conglomerates occur as well as distinctive rusty black pelites. A preliminary sedimentological analysis by R.K. Pickerill, University of New Brunswick, suggests that this part of the sequence comprises an extensively

MAP LEGEND

18 ROCKY BAY PLUTON: Coarse grained, homogeneous grey tonalite with poikilitic biotite clots (relations to 17 unknown).

17 DEADMANS BAY PLUTON: Coarse grained, grey biotite granodiorite with 25-40 per cent pink alkali feldspar megacrysts.

RAGGED HARBOUR COMPLEX (units 12-16)

16 White muscovite leucogranite, commonly aplitic and garnetiferous with coarse grained to graphic patches and schliers.

15 Coarse grained, heterogeneous granitoid gneisses and pegmatite.

14 Trachytoid, porphyritic biotite granodiorite (southern facies).

13 Heterogeneous fine- to medium-grained, granular biotite-muscovite granodiorite, with pegmatitic patches (northern facies).

12 Sheeted complex of biotite-muscovite granodiorite, commonly garnet-bearing, and rheomorphosed rocks of units 3, 4, 10 and 11.

DAVIDSVILLE SEQUENCE (units 6-11)

11 Black slate, pelitic schist, and (11a) highly metamorphosed equivalents.

10 Interbedded slate, greywacke and sandstone.

9 Sub-angular blocks of units 3-11 in black, pyritic, pelitic schist. Minor calcareous breccia and calc-silicates. (Not a stratigraphic unit).

GANDER RIVER ULTRAMAFIC BELT (units 7-8)

8 Dykes and sills of fine grained blue-black gabbro.

7 Massive to schistose pyroxenite, feldspathic pyroxenite, serpentinite and talc-chlorite-amphibole schist.

6 Mafic volcanic rocks; massive to pillowed greenstone, agglomerate, minor tuff and black pelitic schist.

BURNT COVE SEPTUM

5 Interbedded semipelitic and psammitic schists intensely deformed.

GANDER SEQUENCE (units 3-4)

4 Semipelitic schists with interbeds of pink quartzite; 4a, rocks with granitoid portions (anatectic melt).

3 Intercalated pelitic and semipelitic schists.

RIVER SEQUENCE

2 Semipelitic schist, greenstone, granitic mylonite. Very strong cataclastic foliation gives slaty or phyllitic appearance.

FLINN'S TICKLE SEQUENCE

1 Tonalitic, quartzitic and amphibolitic gneisses, locally containing intensely deformed boudins.

Primary fabric; tops unknown, tops known (from sedimentary or volcanic structures), primary igneous fabric.

Schistosity or cleavage, first, second, third.

Gneissosity

Rodding, pencil structure, mullions, and mineral lineation.

Minor folds with plunge and axial plane, sense of tectonic transport known, unknown.

Fault or shear zone, sense of movement unknown.

Orientation of granitoid "sweats" and partial melts.

Geological contact, defined, approximate, deduced from air photos, assumed.

bioturbated, shallow-water unit. The top of the Davidsville sequence was not observed within the mapped area.

The Davidsville sequence is transected by a distinctive rock unit consisting of large subangular to subrounded blocks of volcanic, ultramafic, calcareous, and Gander sequence rocks set in a fissile, contorted, black, pyritic pelite matrix (unit 9; mélange of Kennedy and McGonigal, 1972). Mapping shows that this unit cuts downward across the stratigraphy of the Davidsville rocks, terminating in the upper part of the Gander sequence.

The Ragged Harbour plutonic complex (units 12-16) is divided into two dissimilar parts by the Burnt Cove septum (unit 5), a narrow belt of highly deformed semipelitic and psammitic schists containing gabbroic dykes and pips of ultramafic rocks. To the northwest of this zone the pluton consists of heterogeneous, fine- to medium-grained biotite-muscovite granodiorite with weak foliation (unit 13). Various phases of this rock intrude and grade into one another. Unfoliated pegmatite masses, up to 50 m in largest dimension, comprise about 15-20 per cent of the outcrop. Deformed and partially assimilated mafic dykes containing xenoliths of the host granite occur on the north shore of Ragged Harbour, demonstrating at least one period of remobilization of the complex. The northern part of the Ragged Harbour complex has invaded Gander and Davidsville sequence rocks in a complex series of approximately conformable sheets ranging in thickness from a few millimetres up to 300 m (unit 12). The edges of many of these sheets are more leucocratic than the bulk composition, approaching garnet-muscovite granite.

The southern part of the Ragged Harbour complex forms a strikingly homogeneous mass of pale grey, trachytoid, porphyritic biotite granodiorite (unit 14), containing 20-40 per cent of feldspar phenocrysts as large as 3 by 6 mm. The foliation of this mass remains consistent and unaffected by structures around it. The southwest extremity of the mass contains a roughly circular stock, about 1 km in diameter, of garnetiferous muscovite leucogranite (unit 16). We believe this body to be a phase of the Ragged Harbour Complex, presumably a late differentiate. Similar compositions occur ubiquitously as aplitic dykes up to a few centimetres in width. The wide distribution and negligible volume of these dykes preclude correlation with any single pluton. We deduce that they form the final product of differentiation in all the granitic (sensu lato) rocks of the area, and hence are without value for correlation.

The southeastern margin of the Ragged Harbour complex exhibits a sheeted relationship with country rocks (unit 15). The granitoid component is more coarse grained, leucocratic and heterogeneous than the porphyritic granodiorite of the main mass.

We observed only the western margin of the Deadmans Bay pluton. The contact with the Flinn's Tickle sequence is approximately conformable, but xenoliths of the latter are found within the pluton. The plutonic rocks vary slightly from place to place, but consist in general of massive, coarse biotite granodiorite containing 25-40 per cent of pink alkali feldspar megacrysts 2-6 cm in length. Megacrysts may exhibit plagioclase rims, and/or zones of inclusions. Alignment of euhedral megacrysts commonly defines an excellent foliation in the rocks, but no tectonic foliation is present.

The relation of the Deadmans Bay pluton to other rock units is unknown, but one inclusion within it probably derives from the Ragged Harbour complex.

The Rocky Bay pluton underlies most of Rocky Bay, but outcrop is generally poor. The rock consists of coarse grained, generally massive, hornblende-biotite tonalite with characteristic poikilitic biotite. The pluton has hornfelsed Davidsville sequence rocks at the southeastern extremity of Rocky Bay, healing and obliterating tectonic fabric. Both the Rocky Bay and Deadmans Bay plutons display the ubiquitous aplitic dykes previously noted.

Structure

With the exception of the Rocky Bay and Deadmans Bay plutons, all rock units exhibit one or more foliations, and display, locally at least, two or more phases of folding. The most prominent foliation commonly trends northeast, a fabric termed by us the "main foliation". Kennedy and McGonigal (1972) referred to this fabric as S_2. Observations of dip and facing show the main fabric folded on a megascopic scale, forming an antiform-synform pair in the northwestern part of the map-area. At least two cleavage directions postdate the main fabric, and locally obliterate it. The late cleavages produce kinking in low-grade metamorphic rocks and micro to meso-scale folding in higher grade rocks. In general all main foliation and later structures plunge northeast in the northeastern part of the area, and southwest in the southwestern part, defining a dome in the Ragged Harbour region.

In Gander sequence rocks of relatively low metamorphic grade we detect vestiges of folding prior to the main deformation. Refolded fold noses in thin psammite beds have been boudined and strung out along the main foliation. Such relicts do not occur in Davidsville sequence rocks, possibly indicating a structural discontinuity between Gander and Davidsville sequences.

In high-grade metamorphic rocks containing anatectic melt, small-scale structures show extremely diverse styles and orientations over small areas. Variously deformed resistant beds "go adrift" forming disoriented boudins. Such fragments occur commonly in the sheeted complexes surrounding the Ragged Harbour complex. Presumably they indicate a very high degree of ductility in the matrix.

North- or northwest-trending faults with small lefthand displacements occur in several places west of Ragged Harbour. Northeast- or east-trending faults of small displacement complicate the northeast boundary of the Ragged Harbour complex. An undetermined number of faults parallel to strike occur within the Gander River ultramafic belt. However the only fault of regional significance is the Burnt Cove fault zone, which forms a narrow northeast-trending belt of highly deformed rocks, bounded on both sides by prominent mylonite zones. This fault truncates the Gander and

Davidsville sequences, small slivers of which are smeared along the fault. We found no direct evidence of the displacement, but the map pattern suggests southeast side up.

Metamorphism

All supracrustal and most plutonic rocks observed by us have been metamorphosed. The lowest grade rocks, in the Davidsville sequence on the Carmanville highway, form spotted slates containing andalusite, garnet and biotite. In higher grade rocks cordierite appears and garnet is more abundant. Sillimanite occurs within rocks of the sheeted complexes, and kyanite occurs in disoriented blocks of the Flinn's Tickle sequence.

A progressive Abakuma-type metamorphism with a culmination extending southeast from Aspen Cove produced anatexis. Stages of arrested melting can be observed in spectacular detail, from initial formation of melt pods at boudin terminations, through progressive disorientation and loss of fabric in the host, to formation of sheeted complexes. Psammitic and semipelitic compositions melted most readily, leaving pelitic layers preserved within anatectic melt. The metamorphic culmination follows no known pluton, although it bears a relation to the Ragged Harbour complex. We believe this relation to be one of related causes, rather than cause and effect.

Earlier periods of metamorphism can be detected. Two ages of andalusite are obvious in many rocks of the Davidsville sequence. Two or three ages of porphyroblasts can be defined by relations to fabric in some rocks of the Gander sequence. However, profound reconstitution of the rocks during latest metamorphism makes it difficult to reconstruct the sequence of earlier events.

Conclusions

We emphasize the complexity of the geological history, and the inapplicability of some classic concepts. At least three periods of granite emplacement occurred within the Ragged Harbour complex, and both evidence for regeneration and the ubiquity of late aplite dykes suggest that the plutonic rocks are polygenetic and polychronous. The concept of "age of intrusion" must be defined with extreme care if it is to have any meaning. The polymetamorphism and polydeformation of all supracrustal sequences, together with the local occurrence of plastic or fluid deformation, make the unravelling of stratigraphic and structural successions a matter of great difficulty. Despite these problems, we have drawn three major conclusions from this preliminary field study.

(1) All supracrustal rocks in this area have been significantly deformed and metamorphosed at least twice. Although the distinction between Davidsville and Gander sequences is valid (Jenness, 1963), the interpretation given to it by Kennedy and McGonigal (1972) fails on this point, and must therefore be considered generally suspect.

(2) Granitoid rocks in the mapped area fall into the general categories of quartz diorite (tonalite), biotite (-muscovite) granodiorite, and megacrystic granodiorite. "Garnetiferous leucogranites" (Kennedy and McGonigal, 1972; Strong *et al.*, 1974) form late phases of negligible volume in all these types, and have no value in indicating relative time of intrusion.

(3) The Gander and Davidsville sequences are truncated by a major southwest-trending fault. Their relation to the Flinn's Tickle sequence and the Deadmans Bay pluton thus remains unresolved.

Acknowledgments

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PRELIMINARY PALYNOLOGICAL INVESTIGATION OF THE WINDSOR GROUP (MISSISSIPPIAN) OF NOVA SCOTIA

Project 740110

J. Utting[1]
Regional and Economic Geology Division

Work has begun on the palynology of the Windsor Group (Mississippian) of Nova Scotia with a view to establishing a detailed zonation based on palynomorph assemblages, and to correlate these zones with those based on brachiopods, corals, conodonts and foraminifera. The study was carried out under a contract from the Geological Survey of Canada as support for H. H. J. Geldsetzer (Regional and Economic Geology Division, Geological Survey of Canada, Ottawa), who is studying the Windsor rocks of Eastern Canada.

For the preliminary reconnaissance, prepared palynological slides, along with certain unpublished information, were provided by M.S. Barss of the Atlantic Geoscience Centre. The samples included representatives from subzones B?, C, D and E (Fig. 72.1) and also material of uncertain stratigraphic position which was collected near the Windsor/Canso boundary. Unfortunately no sample rich in palynomorphs has yet been found in subzone A, the lithology of which is generally a limestone. Most of the material, mainly from localities situated in Cape Breton Island, Nova Scotia (Fig. 72.2), has been collected over the years by officers of the Geological Survey of Canada. The identification of the various macrofossil subzones in the Cape Breton area was carried out by Stacy (1953) and in the Antigonish area by Sage (1954); their work was based largely on the subdivisions recognized earlier by Bell (1929) utilizing brachiopods and corals in the Windsor type area. Previous work concerning the palynology of the Windsor Group has been carried out by Barss (Barss, 1967; Hacquebard, 1972; Howie and Barss, 1975) who tentatively proposed the existence of two zones based on palynomorphs and who identified and illustrated miospores of subzone C from a locality at Frenchvale Brook, Northwest Arm of Sydney Harbour, Nova Scotia.

The results of the initial investigation of eleven samples suggest that there are general vertical variations in the miospore assemblages throughout the Windsor, and that these variations are both of a qualitative and quantitative nature; acritarchs occur only rarely (e.g. *Cymatiosphaera* sp.). Most of the samples are dominated by the miospore *Rugospora minuta* Neves and Ioannides, although this species is less common in the samples from the Windsor/Canso boundary. The genus *Punctatisporites* is also abundant, but becomes less common in the younger part of the Upper Windsor (subzones D and E). In the single sample investigated from the Lower? Windsor (subzone B?), there is an apparent lack of diversification of the palynomorphs and *Rugospora minuta* and *Punctatisporites* spp. are supplemented mainly by *Crassispora, Cyclogranisporites* and less commonly *Stenozonotriletes*. Strongly ornamented forms such as *Convolutispora* and *Verrucosisporites* are rare. Many species recorded by other workers from the Horton Group continue into the Windsor, but some species, which have not been reported from the Horton (Hacquebard, 1957; Playford, 1963; Barss, 1967; Varma, 1969), appear in the subzone B? sample, e.g. *Rugospora minuta* and *Anaplanisporites* sp. It should be stressed that these observations concerning the Lower Windsor are based on a single, but extremely fossiliferous, sample from Cape Dauphin (loc. 3, Fig. 72.2), but that the subzone B? identification of the sample is not certain. The sample from the B? subzone comes from an argillaceous unit below the B_2 Limestone of Stacy (1953). The validity of the B subzone identification of Stacy (1953) and Sage (1954) for this limestone was questioned by Mamet (1970), who on evidence from foraminifera favours a Late Windsor age (zone 16 sup.), although he did record outcrops of subzone B in the area surrounding Cape Dauphin. In addition the anomalous occurrence of *Nodosinella* recorded by Stacy (1953) in a B subzone

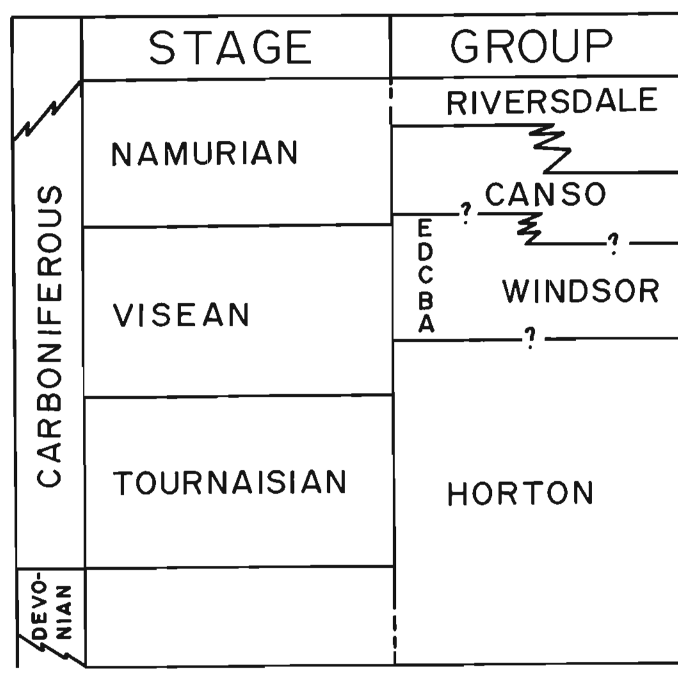

Figure 72.1. Stratigraphic subdivisions and ages of part of the Upper Paleozoic in Eastern Canada (modified from Table 1 of Howie and Barss, 1975).

[1]Institut national de la recherche scientifique, INRS-Pétrole, Université du Québec, Sainte-Foy, Québec. G1V 4C7.

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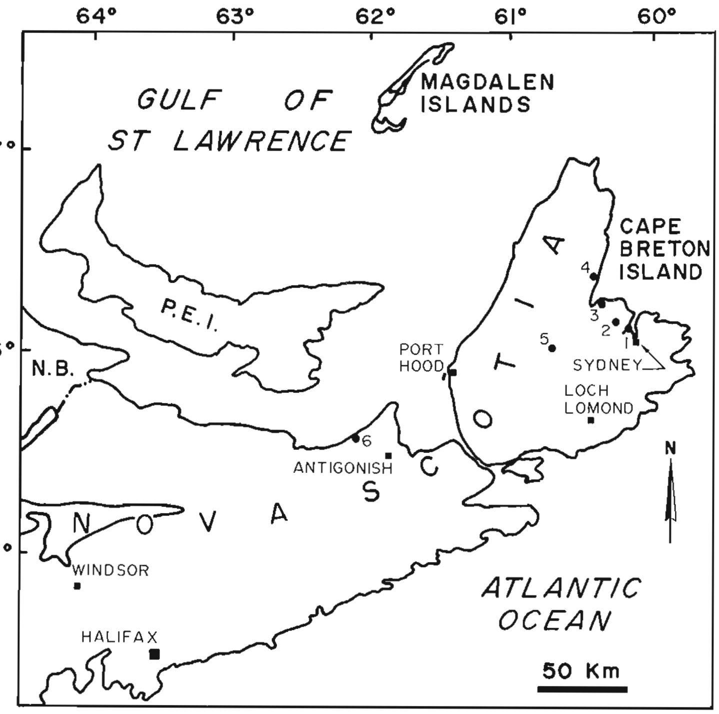

Figure 72.2. Location of samples investigated.

Key to sample localities, giving general geographical location and probable subzone of beds sampled (all sample numbers are those of the Geological Survey of Canada).

1 = GSC 1659, 1 sample, Frenchvale Brook, Balls Creek, Northwest Arm of Sydney Harbour (subzone C).
2 = D1013, 1 sample, Saunders Cove, Groves Point, St. Andrews Channel (subzone C?).
3 = D1014, 3 samples, Cape Dauphin (subzones B?, D and E).
4 = D1011, 1 sample, Skir Dhu (subzone C? or E?).
5 = GSC667, 786 and 787, 3 samples, Baddeck Bridge and Baddeck (Windsor/Canso boundary).
6 = D1010, 2 samples, Knoydart Point (subzone uncertain; samples underlie limestone attributed by Mamet (1970) to zone 16, latest Viséan).

was commented on by Bell (1958). Thus if the revised age for the 'B_2' Limestone proposed by Mamet is accepted and there is no significant sedimentary or tectonic break between the Limestone and the underlying argillaceous unit then the latter would be younger than subzone B.

In subzone C the assemblage is generally similar to that described above in that there is a lack of strongly ornamented forms; *Punctatisporites, Cyclogranisporites, Rugospora* and *Crassispora* are all common as, in one sample, is *Auroraspora.* In this subzone occur the species *Schopfites claviger* Sullivan and *Densosporites variomarginatus* Playford, but they are rare, and the latter has yet to be found in any other subzone.

Most significant changes occur in the sample from the upper part of subzone D. In this sample *Rugospora minuta* remains the dominant form, although *Crassispora trychera* Neves and Ioannides also is abundant; the remainder of the assemblage is more diverse than any of the older samples studied and strongly ornamented genera such as *Knoxisporites, Raistrickia, Pustulatisporites,* are well represented. Certain of the forms found, e.g. *Raistrickia ponderosa* Playford and *Pustulatisporites gibberosus* (Hacquebard) Playford occur in the Horton (Hacquebard, 1957; Playford, 1963) although they were not seen in the material from subzones B?, C and the lower part of D.

The samples from subzone E were relatively poor in palynomorphs, although common are *Rugospora minuta, Crassispora trychera* and *Schopfites claviger.* The material from the Windsor/Canso boundary shows many similarities to that from the upper part of subzone D with respect to diversity of the assemblages and present in significant numbers are *Knoxisporites, Spelaeotriletes, Schopfipollenites* and *Vallatisporites.* *Rugospora minuta* is present in less significant numbers than in the older samples. These Windsor/Canso assemblages show many features in common with those described from the lower part of the Canso Group by Neves and Belt (1970).

Thus from the material studied and in spite of the uncertainty concerning the subzone identifications of some of the samples, it would appear that there are significant differences in the palynomorph assemblages within the Windsor Group and this suggests that it should be possible to recognize assemblage zones, although some of the variations outlined above will almost certainly require modification as additional samples are investigated. Yet to be established is the nature of the lateral variations in the assemblages due to the lateral lithofacies changes, the latter being most significant in the Windsor. Also to be determined are the effects of vertical facies changes; that these had some measure of control is suggested by the fact that some forms occurring in the upper part of subzone D have been recorded from the Horton Group, but have yet to be seen in subzones B?, C and the lower part of D. Also further work is required on material from the Lower Windsor (subzones A and B).

It is proposed to continue the project by investigating in detail vertical sections where a number of the macrofossil subzones are thought to be present. For example in the Lake Enon area, near Loch Lomond, Cape Breton Island, boreholes have been drilled through the Windsor by Kaiser Celestite Mining Ltd. and detailed sampling of the core has already been carried out in co-operation with H.H.J. Geldsetzer. Also it is planned to investigate samples from the Windsor section of Port Hood Island in order to correlate the palynomorph assemblages not only with the macrofossil subzones but with those established by Mamet (1970) using foraminifera and by von Bitter (in press) and by Globensky (1967) using conodonts. Finally correlation of the Windsor palynomorph assemblages with those of similar age elsewhere in the world may resolve some of the problems concerning the precise age range of the Windsor Group and may indicate also the presence, or absence, of significant breaks in sedimentation.

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GEOLOGICAL SETTING OF ASBESTOS-BEARING ULTRAMAFIC ROCKS ALONG THE BAIE VERTE LINEAMENT, NEWFOUNDLAND

E. M. R. Research Agreement 1135-D13-4-130/76

Harold Williams[1], James P. Hibbard[1], and J. T. Bursnall[2]
Regional and Economic Geology Division

Mafic-ultramafic complexes, probably all of ophiolitic parentage and of Early Paleozoic age, occur as discontinuous plutons along a linear zone in Newfoundland known as the Baie Verte Lineament (Kidd, 1974). Some of the ultramafic rocks contain asbestos and a large commercial deposit is being mined at Advocate near Baie Verte. The mafic-ultramafic plutonic rocks can be traced within Burlington Peninsula southward about 100 km from Baie Verte to Sandy Lake (Neale and Nash, 1963) and beyond to Glover Island in Grand Lake. From there the Baie Verte Lineament is probably coincident with the Cape Ray Suture (Brown, 1973) on the south coast of Newfoundland and it presumably continues much farther southwestward through the Quebec Serpentinite Belt to Brompton Lake. The Baie Verte-Brompton Line (St. Julien *et al.*, 1976), is an important structural junction in the Canadian Appalachians and it is the worlds richest asbestos belt (Lamarche, 1975).

In Newfoundland, the geology of the Baie Verte Lineament is complex, nomenclature is confusing, and opinion is divided on the significance of local geological relationships. The present study was undertaken with the hope of compiling updated maps of the lineament, to trace its extent in Newfoundland and improve comparisons with Quebec, and to rationalize the regional geology of Burlington Peninsula while reconciling relationships there with well established geological relationships in western and central Newfoundland. One of us (H. W.) conducted a reconnaissance survey of all of the easily accessible geology of Burlington Peninsula in an attempt to trace the Baie Verte Lineament across insular Newfoundland; another (J. P. H.) is concerned with the delineation and correlation of volcanic rocks along the eastern margin of the lineament; the other (J. T. B.) conducted detailed studies west of Baie Verte and including Advocate Mine.

All mafic-ultramafic complexes along the Baie Verte Lineament are interpreted as ophiolite suites in various states of preservation and structural complexity. Some are only mildly deformed and metamorphosed, and these exhibit the essential components of ophiolite, e. g. the body at Point Rousse (Fig. 73. 1). Others are dismembered, imbricated, and in places intensely foliated with separate structural slices containing only one or two ophiolite components, e. g. faulted slices of gabbro, ultramafic rocks, sheeted dykes and pillow lavas at various places along the lineament.

[1]Department of Geology, Memorial University of Newfoundland, St. John's, Newfoundland A1C 5S7.

[2]Department of Geology, Syracuse University, Syracuse, New York, 13210.

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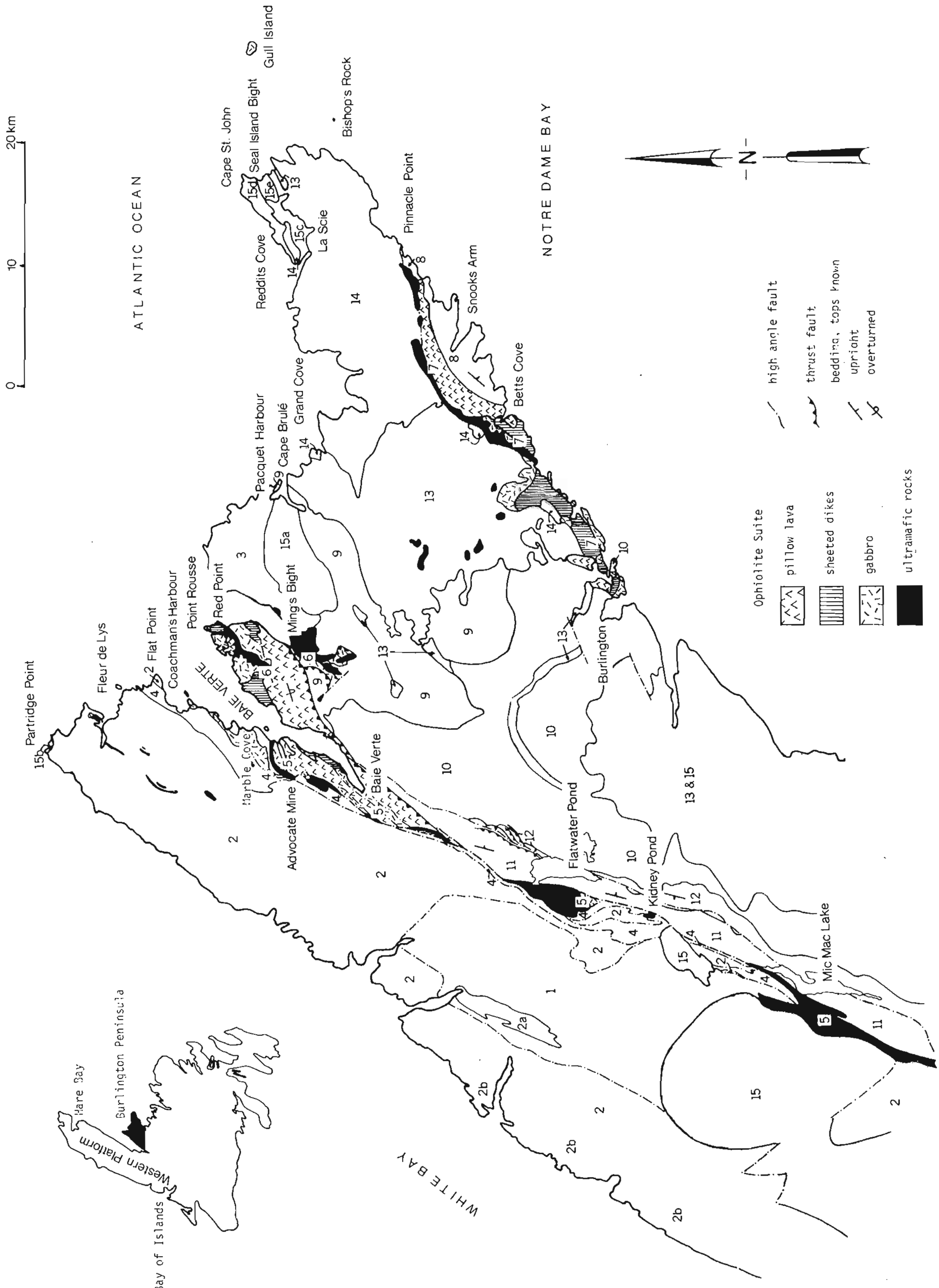
The ophiolite complexes along the lineament are bounded to the west by polyphase deformed and metamorphosed rocks of the Fleur de Lys Supergroup (Church, 1969). Some of the mafic plutonic rocks northwest of Advocate Mine are integral parts of the polyphase deformed and metamorphosed terrane. The age of the Fleur de Lys Supergroup, the timing of its deformations, and its relationships to nearby ophiolite suites and other rock groups are all points of major contention in the geology of the Burlington Peninsula (e. g. Neale and Kennedy, 1967; Church, 1969; Dewey and Bird, 1971; Church and Stevens, 1971; Kennedy and Phillips, 1971; Bursnall and De Wit, 1975; Kennedy, 1975; Williams, 1975; and DeGrace *et al.*, 1976).

The mafic-ultramafic complexes are bounded to the east by a variety of rocks. In the north at Ming's Bight, ultramafic rocks are thrust eastward against psammitic schists of the Mings Bight Group (Baird, 1951) and metamorphosed mafic volcanics of the Pacquet Harbour Group (Church, 1969). The Mings Bight has been correlated traditionally with the Fleur de Lys to the west of the lineament, and as no major stratigraphic or structural breaks have been recognized between the Mings Bight and nearby volcanic groups (DeGrace *et al.*, 1976), i. e. Pacquet Harbour Group, Cape St. John Group (Baird, 1951), Grand Cove Group (Church, 1969), Cape Brulé Porphyry (Baird, 1951), all have become known as the "Eastern" Fleur de Lys (e. g. Church, 1969). None of these rocks east of the lineament resembles the Fleur de Lys, except the Mings Bight Group, and this extension of the Fleur de Lys and implied correlation of rocks and structures across the lineament is erroneous, at least in part, for some of the eastern volcanic rocks, e. g. Cape St. John Group, are younger than the ophiolite suites of the Burlington Peninsula (Neale *et al.*, 1975; DeGrace *et al.*, 1976).

South of Baie Verte in the vicinity of Flatwater Pond and Mic Mac Lake, the mafic-ultramafic complexes are bordered to the east by a well-established east-facing sequence of mainly mafic volcanic and volcaniclastic rocks with a coarse conglomerate unit at its base (Kidd, 1974). These rocks were traditionally grouped with the ophiolites and their mafic volcanic cover rocks and assigned to the Baie Verte Group (e. g. Watson, 1947; Neale and Nash, 1963). The results of the present study suggest that they postdate the earliest deformation of the ophiolite suites. They are referred therefore, informally, to the Flatwater group. The Flatwater group is overlain by silicic volcanic rocks of the Siluro-Devonian Mic Mac Lake Group (Kidd, 1974) with only slight angular discordance. Nearby, the Mic Mac Lake Group overlies the Burlington Granodiorite (Neale and Kennedy, 1967) with pronounced nonconformity. At the south end of the Burlington Peninsula, the Mic Mac


ATLANTIC OCEAN
NOTRE DAME BAY
WHITE BAY
0
10
20 km
N
Cape St. John
Seal Island Bight
Gull Island
Bishop's Rock
Reddits Cove
La Scie
Pinnacle Point
Snooks Arm
Betts Cove
Grand Cove
Cape Brulé
Pacquet Harbour
Ming's Bight
Red Point
Point Rousse
Coachman's Harbour
Flat Point
Fleur de Lys
Partridge Point
BAIE VERTE
Marble Cove
Advocate Mine
Baie Verte
Flatwater Pond
Kidney Pond
Burlington
Mic Mac Lake
13 & 15
Ophiolite Suite
pillow lava
sheeted dikes
gabbro
ultramafic rocks
high angle fault
thrust fault
bedding, tops known
upright
overturned
Hare Bay
Burlington Peninsula
Western Platform
Bay of Islands

DEVONIAN AND OLDER

INTRUSIVE ROCKS

15 — mainly massive granitic intrusions; 15a, Dunamagon Granite; 15b, Partridge Point Granite; 15c, La Scie Granite; 15d, Reddits Cove Gabbro; 15e, Seal Island Bight Syenite

MIC MAC LAKE GROUP

12 — mainly silicic volcanic rocks

FLATWATER GROUP

11 — megaconglomerate and mafic volcanic rocks

CAPE ST. JOHN GROUP

14 — mainly silicic volcanics

CAPE BRULÉ PORPHYRY

13 — quartz-feldspar porphyry

LOWER ORDOVICIAN AND OLDER

BURLINGTON GRANODIORITE

10 — medium-grained massive pink to grey granodiorite

PACQUET HARBOUR GROUP

9 — mainly mafic volcanic rocks

SNOOKS ARM GROUP (7 & 8)

8 — mafic volcanic and volcanic clastic rocks

BETTS COVE COMPLEX

7 — ophiolite

FLEUR DE LYS SUPERGROUP

BIRCHY COMPLEX

4 — greenschist and ophiolite mélange

ADVOCATE COMPLEX

5 — ophiolite and mafic volcanic rocks

POINT ROUSSE COMPLEX

6 — ophiolite and volcaniclastic rocks

2 — psammitic to semi-pelitic schists; 2a, metaconglomerate; 2b, carbonate breccia

MINGS BIGHT GROUP

3 — psammitic schists

PRECAMBRIAN

1 — Grenville basement gneisses

Figure 73.1. Geological map of Burlington Peninsula, Newfoundland.

Lake Group is faulted against the mafic-ultramafic complexes of the Baie Verte Lineament and the Fleur de Lys Supergroup, and the Flatwater group is absent.

Gabbroic and ultramafic parts of the ophiolite suites along the Baie Verte Lineament were mapped separately by most previous workers (e.g. Neale and Nash, 1963), while dykes and volcanic rocks of the same ophiolite suites were grouped with overlying volcanics and volcaniclastics and assigned to the Baie Verte Group. Several violations of the stratigraphic code by reassignment of some Baie Verte Group rocks to the Pacquet Harbour Group (Church, 1969) and others to the Advocate sequence (Kennedy, 1975), without redefinition of its remaining parts, have destroyed the usefulness of this well known and widely used stratigraphic name. The term Baie Verte Group is therefore abandoned and we refer the ophiolite suites and their cover rocks informally to separate complexes. The well preserved ophiolite and its volcanic cover between Baie Verte and Ming's Bight are referred to the Point Rousse complex. Similar rocks on the west side of Baie Verte, including a polyphase deformed and mylonitic gabbroic sequence at Marble Cove and a less deformed and mainly volcanic sequence at Shark Point (Bursnall, 1975), are designated the Advocate complex. Other deformed and metamorphosed gabbroic and ultramafic rocks toward the west, which form an integral part of the Birchy Schist (Fuller, 1941) of the Fleur de Lys Supergroup, are grouped with the greenschists and designated informally as the Birchy complex.

Relationships along the western margin of the Baie Verte Lineament

The Fleur de Lys Supergroup consists mainly of polydeformed psammitic to semipelitic schists that are metamorphosed to greenschist and amphibolite facies. Thin-bedded marbles and metamorphosed lime-breccia units are common in western exposures on the east side of White Bay and the lithologies of this part of the Fleur de Lys can be equated in a general way with those of the Humber Arm Allochthon in western Newfoundland (Bursnall and De Wit, 1975; Williams, 1975). The recognition of reworked basement gneisses deformed with the Fleur de Lys cover, a basal conglomerate to the cover sequence, and metamorphosed mafic sheets throughout thick basal psammitic parts (De Wit, 1972) all support the contention that the Fleur de Lys represents a clastic sequence built upon the ancient margin of a rifted continent (Bird and Dewey, 1970; Williams and Stevens, 1974). Easterly parts of the Fleur de Lys Supergroup, west of the Baie Verte Lineament, consist mainly of greenschists with thin psammitic and black pelitic units, the Birchy Schist (Fuller, 1941). These metamorphic rocks include large bodies of gabbro and altered peridotite, but there is agreement that the Birchy Schist is an integral part of the polyphase deformed and metamorphosed Fleur de Lys Supergroup (Neale and Kennedy, 1967; Church, 1969; Bursnall, 1975). The Birchy Schist is interpreted to underlie Flat Point psammites along its western margin at Coachman's Harbour (Kennedy, 1971), although structures are complex, facing criteria are scarce or suspect, and most contacts are tectonic. Toward the east, the contact of the Birchy Schist and the Advocate Complex is placed at a tectonic junction in Marble Cove (*see* however Bursnall, 1975), and in general the location of this boundary is also detectable.

Chaotic black pelitic zones within the Birchy Complex at Coachman's Harbour and 3 km southward at Slaughter House Cove are interpreted now as metamorphosed mélanges that contain ultramafic and gabbroic blocks, a variety of clastic metasedimentary blocks, and local marble blocks. Most conspicuous are deformed lenses of bright green actinolite-fuchsite schist up to a few metres long. These were originally ultramafic blocks, as indicated by their mineralogy and the fact that larger nearby ultramafic bodies are recrystallized to similar mineral assemblages at their margins. The black pelites and their discrete blocks show all of the deformational structures of the Birchy complex so that mélange formation preceded polyphase deformation and metamorphism characteristic of the Fleur de Lys. The chaotic zones are thought to be typical ophiolitic mélanges (Gansser, 1974) and their presence here, and their similarity to mélanges in western Newfoundland (Williams, 1975), imply a disruption of the Birchy complex before its polyphase deformation and metamorphism. We therefore prefer the term Birchy complex, rather than Birchy Schist, for this largely tectonic assemblage, although one of us (J.T.B.) maintains that a gross lithostratigraphic correlation is possible across some mélange zones.

Recognition of ophiolite mélanges within the Birchy schists that are deformed by the full sequence of Fleur de Lys deformations, bears upon one of the major points of contention in west Newfoundland geology, i.e. the timing of polyphase deformation and metamorphism in the Fleur de Lys terrane in relation to the time of displacement and transport of the west Newfoundland ophiolite complexes from their site of formation to their present position above the west Newfoundland carbonate bank succession. Thus, if the ophiolites were transported across or through a previously undeformed Birchy terrane, there is no need to appeal to a White Bay root zone for the transported Bay of Islands Complex (Dewey and Bird, 1971), nor are models that propose more easterly root zones for west Newfoundland ophiolites hampered with the alleged existence of an ancient Fleur de Lys orogen intervening between the west Newfoundland carbonate terrane and the site of formation of the ophiolite suites (Church and Stevens, 1971; Kennedy, 1975; Williams, 1975).

Rocks similar to the Birchy schists form the structural metamorphic aureoles of the west Newfoundland ophiolites, e.g. the Goose Cove Schist at Hare Bay (Williams and Smyth, 1973), and occur as possible greenschist tectonic aureoles in the Old Man Cove Formation of the Humber Arm Allochthon (Williams, 1975) and the Murrays Cove Formation at Great Coney Arm of White Bay (Lock, 1969). The implication is that all of the higher igneous slices of the west Newfoundland allochthons passed through or across a terrane such as that represented by the Birchy complex at the Baie

Verte Lineament. The formation of mélange and imbrication within the Birchy complex is therefore to be expected.

The protolith of the Birchy schists is mafic volcaniclastic and volcanic rocks that at deposition may have been the distal westerly equivalents of mafic volcanics like those above the Baie Verte ophiolite suites, e.g. volcaniclastic rocks exposed at the top of the Point Rousse complex on the west side of Ming's Bight.

Small ultramafic and gabbroic bodies in a discontinuous zone that extends from Fleur de Lys Harbour southward through the central psammitic part of the Fleur de Lys belt are also thought to be localized along early structural surfaces, rather than to represent small high level intrusions (Kennedy and Phillips, 1971). Their presence implies further early structural telescoping and imbrication across the Fleur de Lys terrane.

In the Quebec Serpentinite Belt (St. Julien and Hubert, 1975), the ophiolitic rocks, e.g. those at Thetford Mines, are bounded on the northwest by the polydeformed and metamorphosed Sutton-Bennett Schists that are a Fleur de Lys equivalent. Serpentinite screens within the schists, e.g. the Pennington Dyke, are sited at structural surfaces and locally these Quebec serpentinites are also asbestos-bearing.

Mafic-Ultramafic Complexes along the Baie Verte Lineament

The distinction between the Advocate and Point Rousse complexes is based mainly upon structural style, degree of alteration, and the geographic separation of the two. Their boundary is located along a major structural discontinuity that extends northeastward through Baie Verte (Fig. 73.1). Grassy Island, Tin Pot Islands and the Sisters, all in Baie Verte, are all composed of ultramafic rocks that are interpreted as basal parts of the Point Rousse complex. Locally at Tin Pot Islands of Coachman's Harbour, the Point Rousse complex occurs within 1 km of the Birchy complex and the Advocate complex is extremely narrow or absent. Along the lineament south of Baie Verte, the ultramafic rocks are assigned to the Advocate complex.

The Advocate complex has a steep northeast-trending foliation in most places and it is cut by numerous steep northeast-trending shear zones that repeat its rock units (Bursnall, 1975; and in prep.). A large ultramafic body at its northwest margin near Baie Verte consists, in places, of intensely foliated and fractured serpentinites that are host to the Advocate asbestos deposit. Asbestos is also present in serpentinized ultramafic rocks at Marble Cove. The Advocate complex gabbros vary from massive to intensely foliated and some are distinctive white, altered rocks with green fuchsitic smears, e.g. those at Marble Cove. Sheeted dykes are best exposed 1 km northeast of Baie Verte, and the clearest pillow lavas occur at Shark Point, 6 km northeast of Baie Verte.

Black and green slates, black pebbly slates and chaotic black slaty mélange are also present in the Advocate complex. Some of these rocks occur at tectonic contacts so that they are structurally co-mingled with the segmented ophiolite. Their chaotic character results, in part, from tectonic processes during imbrication of the ophiolite suite, e.g. mélange on the west side of Schooner Cove to the northeast of Advocate Mine. Others are clearly depositional breccias with volcanic fragments and rafts of gabbro in a black slaty matrix, e.g. those immediately north of Mine Pond at Advocate Mine. Rather similar, though more metamorphosed black schists, also occur at structural boundaries in the Birchy complex at Seal Cove and Slaughter House Cove, i.e. between Coachman's Harbour and Advocate Mine. Collectively the steep tectonic boundaries marked by smeared black, slaty or schistose rocks are interpreted as significant structural dislocations that localized the incompetent argillaceous rocks. A parallel situation is evident in the Humber Arm and Hare Bay allochthons of western Newfoundland where chaotic black shale or mélange occurs between each structural slice. The regional situation therefore suggests that these black argillaceous zones in the Birchy and Advocate complexes mark zones of imbrication and that this imbrication was related to Early or Middle Ordovician movements responsible for the transport and emplacement of Taconic allochthons in western Newfoundland.

The Point Rousse complex is made up of several distinct structural blocks with their tectonic boundaries marked by foliated serpentinite or foliated carbonate-talc-fuchsite alterations of ultramafic rocks (Norman and Strong, 1975). Three separate blocks on Point Rousse Peninsula contain southeast-facing, overturned sections of gabbro, sheeted dykes and pillow lava. The most complete section occurs south of Red Point where a northwest-dipping, southeast-facing sequence of gabbro, sheeted dykes and pillow lava is followed by local chert beds and a thick section of volcaniclastic rocks, all southeast-facing. Locally at Lower Green Cove and Deer Cove on the western side of Point Rousse Peninsula, pillow lava of the Point Rousse complex is intensely deformed and converted to greenschists. North of Deer Cove, southward-thrusting of gabbros above pillow lavas postdates the development of a steep foliation in the volcanic rocks (Kidd, 1974).

Intense zones of deformation within the Point Rousse complex are possibly of the same generation as similar structural zones in the Advocate and Birchy complexes (Taconic). Later southward to eastward thrusting is probably Acadian.

A large segmented body of ultramafic and gabbroic rocks to the south of Ming's Bight is locally thrust eastward across already deformed rocks of the Mings Bight Group (Kidd, 1974), and it is in tectonic contact with the Pacquet Harbour Group. Its position has been interpreted to suggest a synclinal form for the Point Rousse complex (Kidd, 1974), but structures are complex and of several generations so that this may be an oversimplification. The règional pattern of ophiolite complexes on Burlington Peninsula indicates steep-dipping, southeast-facing sequences, e.g. Advocate Mine, Point Rousse Peninsula, Betts Cove, and Gull Island. With this in mind, some of the ultramafic and gabbroic rocks south of Ming's Bight may form the disrupted basal parts of an ophiolite sequence that has the Pacquet Harbour volcanics at its top. The general pattern appears to reflect repetition by early northwest-thrusting (Taconic) with later steepening, disruption and overturning resulting from later Acadian deformation. A similar structural pattern of steeply-dipping to overturned, south-facing ophiolite sequences is also characteristic of continuations of the Baie Verte Lineament in the Quebec Serpentinite Belt (St. Julien and Hubert, 1975).

Relationships along the Eastern Margin of the Baie Verte Lineament

The interpretation of tectonic relationships along the west side of the Baie Verte Lineament is in large part conditioned by interpretations of relationships among rocks along its eastern side. Toward the north, the Point Rousse complex is bordered to the east by psammitic schists of the Mings Bight Group and metamorphosed mafic volcanics of the Pacquet Harbour Group. These are followed farther east by the Cape Brulé quartz-feldspar porphyry and dominantly silicic volcanic rocks of the Cape St. John Group. Farther southeast, the Cape St. John Group is in contact with the well-known ophiolite suite of the Betts Cove Complex and its overlying volcanic and volcaniclastic rocks of the Lower Ordovician Snooks Arm Group (Fig. 73.1).

South of Baie Verte between Flatwater Pond and Mic Mac Lake, ophiolites of the Advocate complex are bordered eastward by a narrow belt of mainly mafic volcanic rocks (Flatwater group), followed eastward by mainly silicic volcanic rocks (Mic Mac Lake Group), and then the Burlington Granodiorite. All of these rocks to the east of the lineament have been mapped in some detail (Kidd, 1974) but there has been little consensus on the significance of the relationships between the various rock groups.

No tectonic or stratigraphic breaks are recognized among the Mings Bight, Pacquet Harbour and Cape St. John groups so that all are viewed as parts of a continuous and conformable succession. Toward the north between Ming's Bight and La Scie, all of these rocks share the same polydeformation and metamorphism (DeGrace *et al.*, 1976), although more southerly parts of the Pacquet Harbour and Cape St. John groups are relatively little deformed and metamorphosed. A strong lithic resemblance between psammitic schists of the Mings Bight Group and the nearby Fleur de Lys has led to lithic correlation between the two and thus an inferred Early Ordovocian or earlier age for all of these rocks to the east of the Baie Verte Lineament (e.g. Church, 1969; Dewey and Bird, 1971). Correlation across the lineament, based mainly upon local structural style and metamorphic grade in the case of the Pacquet Harbour and Cape St. John groups, is now suspect because of reconfirmation of a clearly unconformable relationship at Pinnacle Point between the Cape St. John Group and underlying layered mafic volcaniclastics of the Lower Ordovician Snooks Arm Group (Neale *et al.*, 1975; DeGrace *et al.*, 1976). Results of this present study also reconfirm the Pinnacle Point unconformity and indicate that the Betts Cove ophiolite complex and upper parts of the Snooks Arm Group extend far to the northeast in subsurface beneath the Cape St. John volcanics to reappear at Gull Island and Bishop's Rock, respectively. The Cape St. John Group is therefore younger than the Snooks Arm Group and Betts Cove ophiolite complex and by inference it must be younger than the ophiolite complexes along the Baie Verte Lineament, i.e. the Advocate and Point Rousse complexes. Whether or not the Pacquet Harbour and Mings Bight groups are also younger than nearby ophiolite complexes remains debatable. Conceivably the Pacquet Harbour Group is a Snooks Arm equivalent. The Mings Bight remains enigmatic, but there is now no assurance that it is a Fleur de Lys equivalent. It is also possible that none of the rocks east of the lineament were involved in the main Fleur de Lys deformation and metamorphism.

Farther south along the Baie Verte Lineament between Flatwater Pond and Mic Mac Lake, ultramafic rocks of the Advocate complex are faulted against a narrow belt of marine mafic volcanic and volcaniclastic rocks that are referred to the informal Flatwater group, for its well exposed section along the north shore of Flatwater Pond. Detailed mapping between Flatwater Pond and Mic Mac Lake (Kidd, 1974) has shown that the Flatwater group is an east-facing sequence that can be divided into a number of formations. Polymictic conglomerates and megabreccias are common at the base of the group along its entire length but it is everywhere faulted against ultramafic rocks of the Advocate complex, or against the Fleur de Lys Supergroup to the west. Toward the east, the Flatwater group includes some silicic volcaniclastics near its top and it is overlain at a sharp contact by bright red silicic volcanics and clastic sediments of the subaerial Siluro-Devonian Mic Mac Lake Group. The contact is apparently conformable in outcrop but regionally it is interpreted as a low angle unconformity (Kidd, 1974). Farther east, on the east side of a faulted syncline, west-facing Mic Mac Lake rocks overlie Burlington Granodiorite with marked nonconformity. Both the Mic Mac Lake and Flatwater groups contain pebbles and boulders of Burlington Granodiorite, implying that the Flatwater, like the Mic Mac Lake, postdates emplacement of the Burlington Granodiorite. Since the Burlington Granodiorite cuts the Pacquet Harbour Group pre- or syntectonically (Kidd, 1974; Kennedy, 1975), i.e. a presumed eastern equivalent of the Fleur de Lys, the relationship led most workers to interpret the Flatwater group as younger than the deformed Fleur de Lys on the west side of the lineament (Neale and Kennedy, 1967; Church, 1969; Kidd, 1974).

We also conclude that the Flatwater group is younger than some deformation of the Fleur de Lys and also younger than the earliest deformation of the Baie Verte ophiolite complexes, based mainly upon the nature of conglomerates at the base of the Flatwater group. In the pastureland north of Mic Mac Lake, large single outcrops of internally brecciated gabbros and a variety of heterogeneous foliated gabbros and dykes are actually blocks up to several tens of metres across in a sandy to dark shaly matrix. Similar megaconglomerate, with very little matrix material between its hugh blocks, occurs at the northwest corner of Flatwater Pond, and nearby to the east a gabbro-diabase block 10 metres across is encased in dark slate that also contains a metre-long block of silicic volcaniclastic rock. Another exposure of conglomerate in a rock quarry on the west side of the Baie Verte Highway and just north of Kidney Pond has mainly gabbro and volcanic clasts with some silicic volcanic clasts. On the floor of the quarry a lens of internally foliated, conspicuous green "virginite[1]" is bordered by shale, so that the deformed "virginite" is interpreted as a block in a much less deformed shale matrix. Another exposure of conglomerate of rather different make-up occurs on the east side of the Baie Verte Highway about 8 km north of Flatwater Pond. It has elongate cobbles and boulders of tough highly indurated quartzose greywacke and quartz-pebble conglomerate in a cleaved green chloritized matrix. Nearby the conglomerate has mainly quartz pebbles and local serpentinite and marble clasts. The gabbro, volcanic, "virginite" and serpentinite clasts of the Flatwater conglomerates were derived from deformed ophiolite complexes. The quartzose greywacke blocks resemble greywackes of the Fleur de Lys Supergroup (for comparison, *see* outcrop of Fleur de Lys on north side of Trans-Canada Highway just east of Hampden turnoff), although it is debatable whether or not these blocks were deformed before incorporation into the conglomerate (Kidd, 1974). Certainly they were highly indurated and resemble massive greywacke within the now-deformed Fleur de Lys. A single large schistose semipelitic block in the Flatwater group on the west side of Mic Mac Lake further implies a deformed Fleur de Lys terrane as a provenance for some clasts (Kidd, 1974).

The Flatwater conglomerates appear to be derived mainly from a deformed ophiolite terrane such as that presently exposed to the west, with local influx of material from the Fleur de Lys Supergroup, which at that time was structurally covered, presumably, by an imbricated ophiolite complex. Stratigraphically higher in the Flatwater group, local outsize ultramafic, gabbro, and rare trondjhemite blocks indicate a continued source of ophiolitic material during deposition of its mainly marine mafic volcaniclastic rocks.

The Flatwater group was previously correlated with similar volcanics of the Advocate and Point Rousse complexes and all were assigned to the Baie Verte Group (Neale and Kennedy, 1967; Kidd, 1974). This correlation naturally led to the view that both the Advocate and Point Rousse complexes also postdate deformation in the Fleur de Lys. Since the Birchy complex of the Fleur de Lys Supergroup itself contains ophiolitic rocks, the correlation implies two generations of ophiolite, one before and the other after deformation of the Fleur de Lys (e.g. Dewey and Bird, 1971; Kennedy and Phillips, 1971; Kennedy, 1975). If the Flatwater group is younger than the Advocate and Point Rousse complexes as suggested here, all of the ophiolites of the Burlington Peninsula may be viewed as relating to a single cycle of generation.

[1] a local name proposed by Norman Peters for a Fe-magnesite-quartz-fuchsite alteration of ultramafic rock.

The apparent absence of the Flatwater group near Baie Verte is interpreted as the result of eastward thrusting of the Point Rousse complex above the younger rocks, i.e. eastward thrusting like that responsible for the emplacement of ultramafic rocks above the Mings Bight Group on the east side of Ming's Bight and imbrication of the Mic Mac Lake-Burlington Granodiorite contact at Flatwater Pond (Figs. 73.1 and 73.2). An occurrence of pebbly black slate with local slide-blocks of gabbro within the Advocate complex at Mine Pond (Shark Point sequence of Bursnall, 1975) is possibly equivalent to megabreccias of the Flatwater group. If so, it should also be interpreted like the Flatwater megabreccias, i.e. a chaotic deposit synchronous with marine mafic volcanism but following the early deformation in nearby mafic-ultramafic rocks and the earliest deformation of the Fleur de Lys Supergroup. Furthermore, any possible correlation between the Flatwater group and Advocate complex demands revision of the latter (Bursnall, 1974; and in prep.).

Southeast of the Quebec Serpentinite Belt, chaotic rocks of the St. Daniel Formation (St. Julien and Hubert, 1975) resemble slaty matrix conglomerate at the base of the Flatwater group. The Coleraine breccia, spatially related to both the St. Daniel Formation and the nearby Thetford Mines ophiolite complex, contains schistose psammitic clasts like the Sutton-Bennett schists to the north. The situation in Quebec may therefore be similar to that proposed for Newfoundland, implying that the St. Daniel Formation accumulated after the earliest deformations in the Sutton-Bennett schists, i.e. Fleur de Lys correlatives. Farther southeast in Quebec, volcanic rocks of the Weedon and Ascot formations may be equivalent to the Pacquet Harbour and nearby groups of Burlington Peninsula. In both areas, these volcanic rocks are noted for their base metal mineralization.

Tectonic History

Minor modifications of previous stratigraphic classifications combined with a change of emphasis regarding some relationships, and a few new facts, all allow a tectonic synthesis of Burlington Peninsula geology that more closely fits and supplements regional relationships in western Newfoundland. The Baie Verte Lineament can be interpreted as an imbricated zone of ophiolitic rocks that lay immediately east of the ancient continental margin of eastern North America. The presence of ophiolitic mélange throughout the Birchy complex implies structural imbrication and westward transport of oceanic crust and mantle across this terrane. Because greenschists of the Birchy complex so closely resemble greenschists in metamorphic aureoles beneath transported ophiolites and other igneous complexes in western Newfoundland, it seems reasonable to equate the transport of west Newfoundland igneous rocks and the formation of mélange in the Birchy complex. The aureole greenschists and those of the Birchy complex appear to have had similar protoliths and, most likely, all formed at the same time.

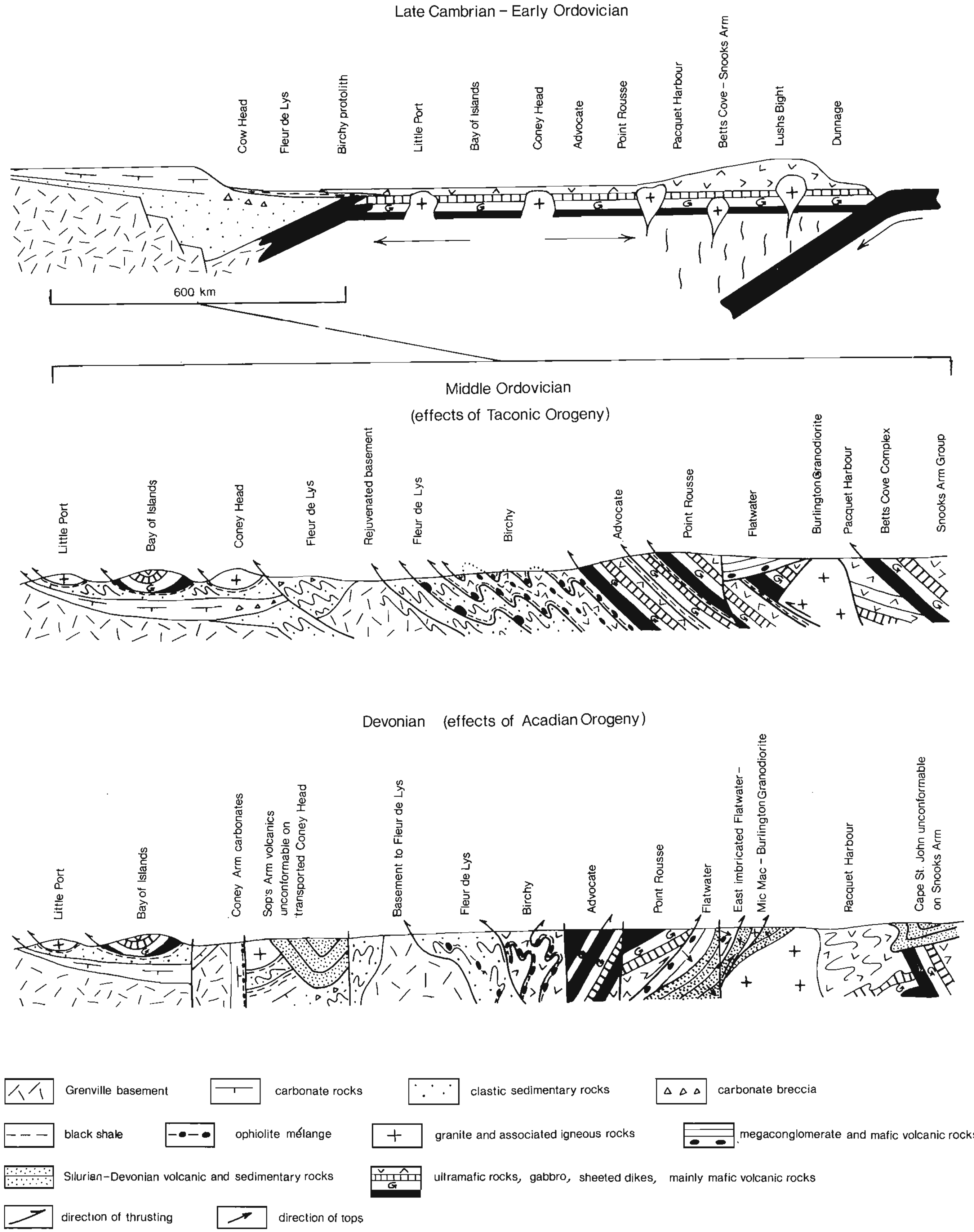


Figure 73. 2. Geological model for the development of the geology of Western Newfoundland.

The Birchy complex was undeformed at the time that its mélange was formed. It follows that the Fleur de Lys was an essentially undeformed continental margin deposit during earliest ophiolite transport. This removes a major obstacle in proposing a root zone for transported ophiolites like the Bay of Islands Complex. Clearly they were derived from the east of the Fleur de Lys terrane, and there was no Fleur de Lys orthotectonic zone (Church, 1969) to impede their transport from central Newfoundland to western Newfoundland. White Bay is ruled out as a root zone, because this analysis requires an undeformed continental shelf-rise prism, which included the Fleur de Lys Supergroup, at the time that the ophiolite obduction began.

The Baie Verte Lineament is therefore the most likely root zone for transported igneous complexes in western Newfoundland, and the variety of transported rocks there reflects a sampling of the once extensive terrane that must have extended east of the lineament for more than one hundred kilometres in Early Paleozoic time. A newly recognized, exposed, sub-Silurian unconformity on the west side of White Bay (*see* also Lock, 1969) and unconformity of the same possible age throughout the Burlington Peninsula reflect Taconic deformation and the Ordovician passage of the transported rocks through local stratigraphic sections. Conformable Ordovician-Silurian sequences in more easterly parts of Notre Dame Bay preclude these areas as a source for the transported rocks of western Newfoundland.

Recognition of the informally named Flatwater group as a volcanic assemblage that is younger and unrelated to the Advocate and Point Rousse complexes of the former Baie Verte Group avoids correlations that demand ophiolite generation after Fleur de Lys deformation. Furthermore, if all of the ophiolites relate to a single cycle of generation, this explains the difficulties in attempts to define a Birchy complex-Advocate complex boundary and to separate the Advocate and Point Rousse complexes. It also allows correlation of all of the ophiolites of Burlington Peninsula with transported ophiolites in western Newfoundland. This simplistic view is also attractive as it avoids inherent complications of multi-stage models.

Correlations of rocks and structures on the east side of the Baie Verte Lineament with those of the Fleur de Lys Supergroup on its west side are all suspect. Even the lithologically similar Mings Bight Group, if of Fleur de Lys age, may once have lain several hundred kilometres farther east, considering the shortening across the west Newfoundland allochthons and the Burlington Peninsula. If far removed, there is no compelling reason to suggest that both were deposited at the same time, and there is less reason to suspect that they were deformed together. The Pacquet Harbour Group resembles the Lower Ordovician Snooks Arm Group and volcanics above the Point Rousse ophiolite suite, rather than anything west of the lineament. Like the Snooks Arm, it probably represents the products of ensimatic arc volcanism. The Cape St. John Group is probably a Mic Mac Lake correlative and both are possibly equivalent to Silurian volcanic rocks on the west side of White Bay. The local intense polydeformation and metamorphism in the Cape St. John may therefore indicate that Acadian deformation was locally as intense and severe as Taconic deformation in this part of Newfoundland.

An evolutionary model that summarizes our present views on the development of the Burlington Peninsula and its relationships to western Newfoundland is depicted in Figure 73.2.

Economic Geology

Asbestos deposits such as those at the Advocate Mine and throughout the Quebec Serpentinite Belt are localized in the intensely deformed ultramafic parts of ophiolite suites. Large transported ophiolites such as the Bay of Islands Complex in western Newfoundland and Mount Albert in Quebec rarely contain significant quantities of fibre, mainly because they were relatively undeformed during transport across the ancient continental margin of eastern North America and, after transport, they lay outside the effective realm of later deformations. In contrast, ultramafic rocks such as those of the Birchy and Advocate complexes were deformed continually during imbrication and destruction of the ancient continental margin with additional deformation during the much later Acadian Orogeny. Consequently, it is in these dismembered smaller ultramafic bodies that the highest concentrations of fibre are likely to occur.

Acknowledgments

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We are all greatly indebted to W.S.F. Kidd for guided excursions along the Baie Verte Lineament, for discussions of Burlington Peninsula geology, for the use of his unpublished geological maps, and for critically reading this paper. We also wish to acknowledge our introduction to the geology of the Burlington Peninsula and numerous discussions with M.J. Kennedy, E.R.W. Neale, J.R. DeGrace, W.R. Church, J.F. Dewey, M.J. De Wit, and R.K. Stevens.

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LANDSAT IMAGERY — A CANADIAN LISTING

V.R. Slaney
Resource Geophysics and Geochemistry Division

A listing of the best quality satellite imagery from the Geological Survey's collection of LANDSAT images has recently been released as Open File 386. This listing, comprising a short introductory text, an index map, and a 60-page catalogue, is available for sales at the various Geological Survey offices.

The Landsat program, was originated and funded by the National Aeronautics and Space Administration (NASA) of the United States of America. It was described by NASA officials as 'a research and devlopment tool to demonstrate that remote sensing from space is a feasible and practical approach to efficient management of the earth's resources'.

From the start, the Canadian Government was interested in the project, and negotiated an agreement with NASA which permitted access to direct transmissions from the satellite while over Canada. The first images from the satellite were received in Canada on July 26th 1972 and have continued to be received to the present day. The Geological Survey of Canada is concerned with the potential applications of remotely-sensed data. To help Survey geologists become familiar with this new class of image and to provide geologists with easy access to good example of it, a collection of Canadian images was established in Ottawa. This reference collection of Multi-Spectral Scanner imagery includes all the material porduced from the whole of Canada with less than 20 per cent cloud cover and of technically good quality. There are more than 15 000 black and white prints representing bands from about 6000 scenes. The scale of the prints is 1: 1 000 000. The collection was begun soon after the first satellite was put into orbit in July, 1972. The latest additions were made in April, 1976. The collection also includes some 500 colour composite prints chosen from the collection of the National Air Photo Library in Ottawa for their quality and for their geological interest. Many of the images are snow scenes and many are affected by atmospheric haze which serves to reduce — sometimes drastically — image contrast. Cloud may also be present anywhere within the image. So far most of the interest shown by geologists has been in locating images with little or no snow, with little or no cloud and with a clear atmosphere.

In 1975 it was decided to establish a sub-collection of very high quality images. About 25 per cent of the full collection (1500 scenes) meets the "high quality" guidelines. The present listing (Open File 386) is the "high quality" imagery.

The criteria used to select imagery for this listing are as follows:

1. Absence or near-absence of perennial snow cover. In areas where there is a permanent snow or ice cover, the period of minimum snow is usually judged to be in August or early September — as close as possible to the first snowfall of the oncoming winter season. There are a few images with snow from a single storm occupying only a few per cent of the image area. Rather more images, usually ones from the Arctic, have snow lying on sheltered slopes and along river valleys. Where either of these conditions exist, mention is made in the 'notes' column of the image concerned.

2. Relatively free of haze. The presence of moisture and particulates in the atmosphere increases the amount of non-image forming light (haze) to reach the satellite-borne scanner. This results in a reduction in density contrasts within the image. Band 4 is always most strongly affected by haze, while band 5 is less so. Bands 6 and 7 are usually not affected except when there is smog or smoke present. The criteria for "haze free" imagery, is that band 4 should be reasonably contrasty when compared with bands 5, 6 and 7 of the same image.

3. Less than 10 per cen cloud within $1\frac{1}{2}$ inches of the edge of the image. The location of cloud is mentioned in the listing. More than 10 per cent cloud is accepted where it lies over a lake or over ocean. Mention is made of this in the accompanying notes. There are a few images with one or more small clouds near the image. Again, mention of this is made in the notes. When calculating the amount of cloud present, one square inch is taken to represent 2 per cent of the image.

4. Good technical quality. This is indicated by an absence or near-absence of dropped lines or other imperfections derived from the original type. The negatives from which the prints are made are also expected to be of good photographic quality i.e. with no blemishes and with a reasonable range of grey tones.

The images are listed using the Fleming system of notation, which is based on the number of the orbit track and position of the image centre alon the track.

More the 90 per cent of Canada is represented. There are still a few areas where coverage is poor, the most notable of which lie in Newfoundland, Labrador and eastern Quebec.

From: Report of Activities, Part A;
Geol. Surv. Can., Paper 77-1A (1977).

At 80°N, this image represents the northernmost limit for the reception of Landsat images. Eureka Sound, crossing the centre of the image, separates Axel Heiberg Island (left of image) from Ellesmere Island (right of image). The image is underlain almost entirely by sedimentary rocks of Carboniferous to Tertiary age. The prominent dark grey ridges crossing Axel Heiberg and Ellesmere islands represent unvegetated shales, mudstones and calcareous siltstones. The flatter light coloured areas with an irregular vegetation cover are underlain by less resistant shales, sandstones and mudstones. Several dome structures are clearly visible in these rocks near the centre of the image. At left, glaciers radiate from the Axel Heiberg snowfields. Ordovician limestones and dolomites form the mountain ridges seen at the right part of the image.

Figure 74.1. Landsat MSS Image of Axel Heiberg and Ellesmere islands Received 28th July 1974 — Image E-1735-19044.

GOLD DISTRIBUTION IN THE KIRKLAND LAKE-LARDER LAKE AREA, WITH EMPHASIS ON KERR ADDISON-TYPE ORE DEPOSITS – A PROGRESS REPORT

E. M. R. Research Agreement 1135-D13-4-62/75

L.A. Tihor[1] and J.H. Crocket[1]
Regional and Economic Geology Division

Introduction

This study over the last year was directed toward two specific goals: (a) to determine background gold values in the major rock-types of the area, and (b) to try to determine the origin of the quartz-carbonate zone (the Larder Lake "Break" of Thomson, 1943) which is closely associated with the gold deposits of the Larder Lake area.

The study is 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).

Background Gold Distribution

Eighty-nine gold analyses have so far been completed by neutron activation analysis, 67 of these from outcrops remote from known economically significant

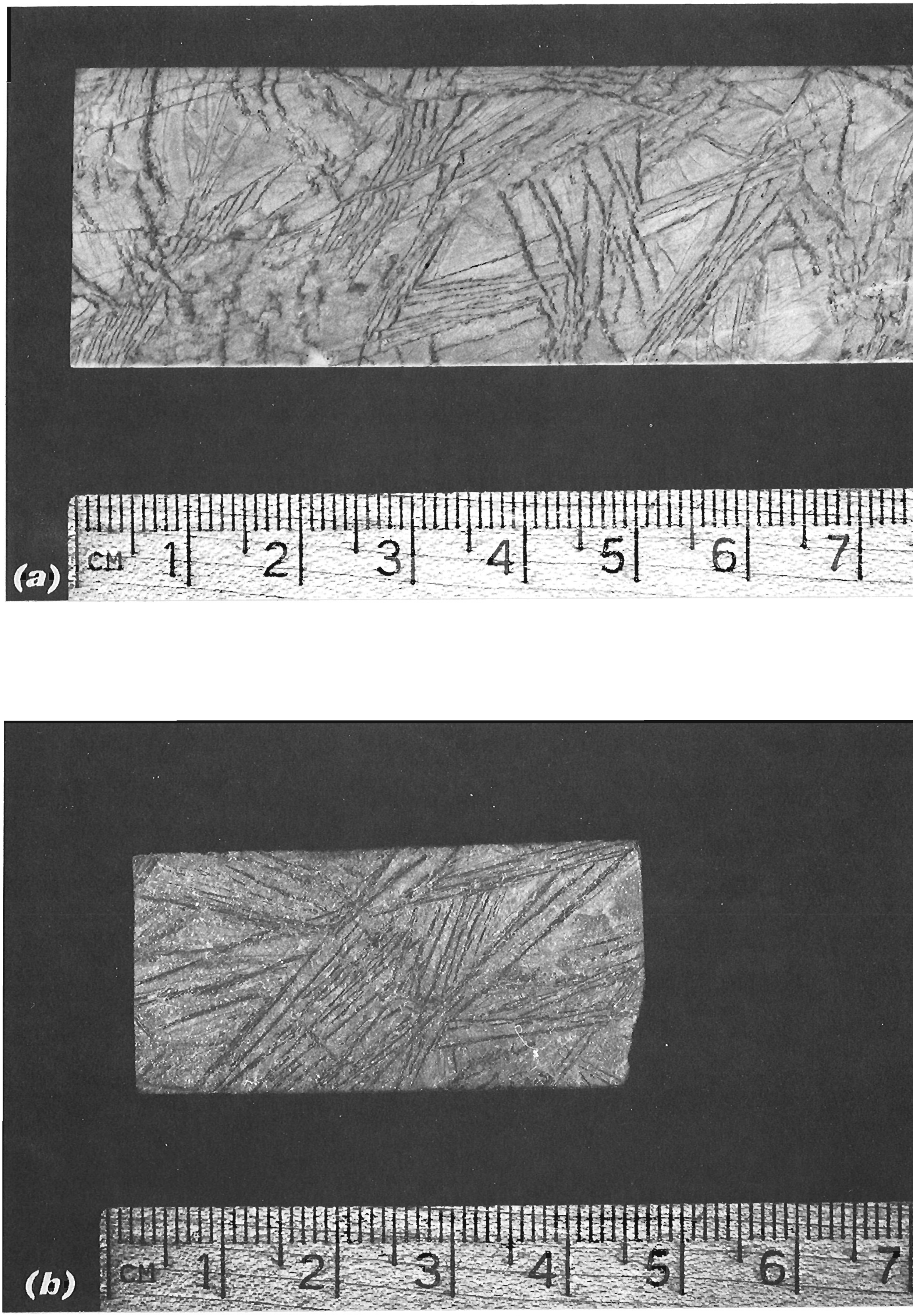
Table 75. 1

Gold in Rocks of the Study Area

	Numbers of samples	Average Au Content (ppb)	Coefficient of Variation (%)
Timiskaming and Younger Rocks			
Otto Stock	2	1. 88	48
Lebel syenite stock	3	1. 45	31
Syenitic dykes not associated with Au ore	2	0. 91	33
Syenitic dyke 3800-foot level – Kerr Addison Mine	1	1751. 06	--
Ultramafic rock not associated with Larder Lake "Break"	1	1. 75	--
Ultramafic volcanics associated with quartz carbonate horizon not near known Au ore	3	0. 68	12
Ultramafic volcanics 3800-foot level Kerr Addison Mine	6	34. 73	192
Oxide facies iron-formation	8	6. 35	118
Graphite schist	1	69. 45	--
Quartz-carbonate rocks not associated with known ore	20	31. 01	280
Quartz-carbonate rocks 3800-foot level Kerr Addison Mine	13	21. 00	62
Intermediate to mafic volcanics associated with quartz-carbonate horizon but not near known Au ore	2	2. 97	0
Intermediate to mafic volcanics, 3800-foot level Kerr Addison Mine	2	16. 86	105
Hypabyssal mafic syenites	3	7. 29	75
Hypabyssal felsic syenites	2	4. 71	114
Sedimentary rocks	6	9. 68	64
Trachytes	7	25. 34	115
Pre-Timiskaming Rocks			
Skead pyroclastics	1	13. 57	--
Gabbro sill	1	1. 05	--
Peridotite at base of gabbro sill	1	0. 93	--
Catharine Basalts	3	8. 51	55
Round Lake Batholith	1	8. 72	--

[1]Department of Geology, McMaster University, Hamilton, Ontario.

From: Report of Activities, Part A;
Geol. Surv. Can., Paper 77-1A (1977)


(a)
CM 1 2 3 4 5 6 7
(b)
CM 1 2 3 4 5 6 7

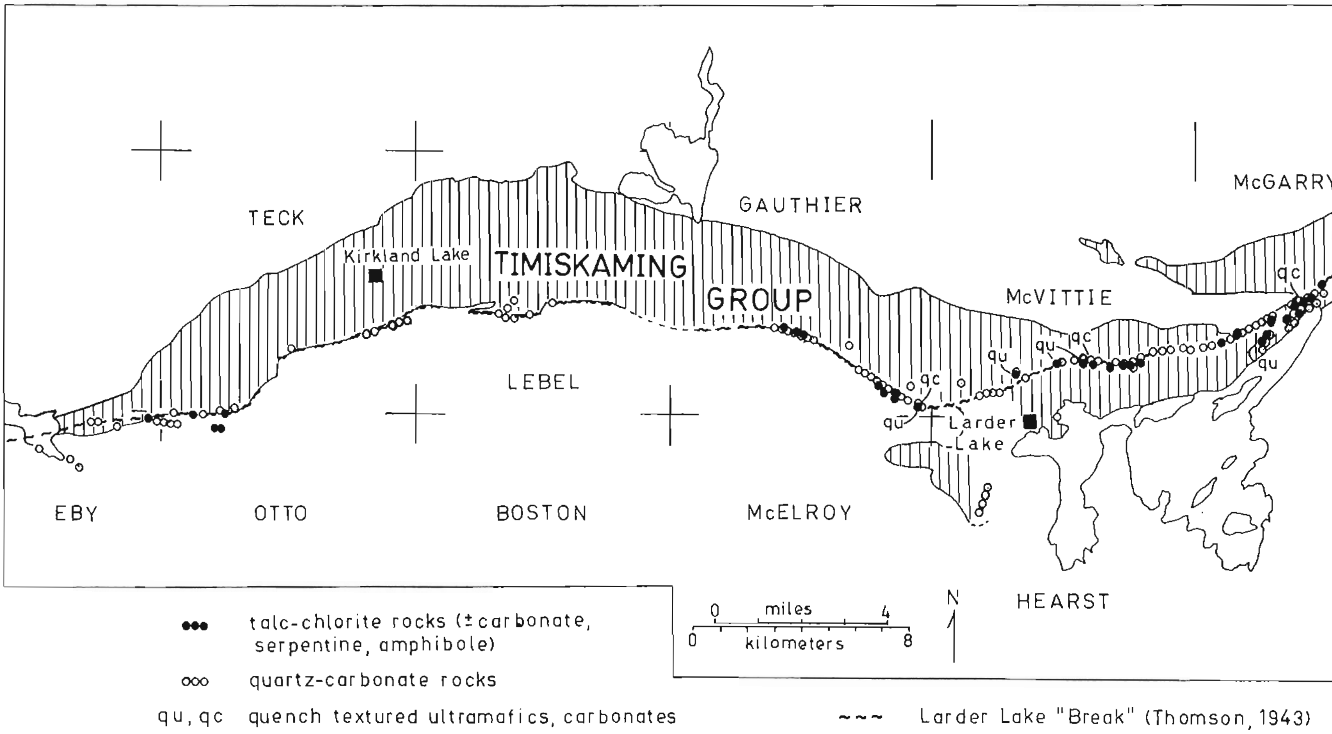


Figure 75. 2. Ultramafic and quartz-carbonate rocks along the Larder Lake "Break" (base map after Hewitt, 1963).

gold occurrences, and the balance across the 3800-foot-level of the Kerr Addison Mine.

Because of the small number of samples of each rock-type analyzed and the great variation in gold content of some of these rocks – notably the quartz-carbonates – some of the average Au concentrations are rough approximations (Table 75. 1).

If 5 parts per billion (ppb) is arbitrarily chosen as the upper limit of "normal" gold content of crustal rocks, the Timiskaming trachytes and the quartz-carbonate rocks are especially enriched in gold. This is not surprising as both rocks are hosts for gold ore deposits. The Upper Canada Mine is an example of the Au-trachyte association (Tully, 1963) and the Kerr Addison Mine demonstrates the Au-quartz-carbonate relationship (Thomson, 1943).

Furthermore, most of the extrusive and sedimentary rocks in this district of both Timiskaming and pre-Timiskaming age appear to be enriched in gold relative to similar rocks elsewhere (Crocket, 1974). However, the paucity in the literature of Au analyses of Archean rocks makes it difficult to know whether this local enrichment is real or a function of the age of these rocks.

Figure 75. 1 (opposite)

(a) Spinifex texture pseudomorphs in green quartz-carbonate rock from diamond drillhole located 330 m east of the Omega Mine, McVittie Township.
(b) Spinifex texture in talc-chlorite rock down strike and 180 m west of the carbonate spinifex.

Only one sample of graphite schist was analyzed. It contained 69. 45 ppb of Au, about 20 times the crustal average (Crocket, 1974).

Eight samples of oxide facies iron-formation were analyzed for gold. These included six from the Adams Mine in Boston Township, one from about 3. 5 km (2 miles) west of the Adams Mine, and one from the vicinity of the Cheminis Mine in McVittie Township. The Au values ranged from 0. 51 to 22. 09 ppb with an average of only 6. 35 ppb. This is unexpectedly low since Ridler (1976) reported carbonate facies of the Boston Iron Formation immediately south of Kirkland Lake grading \$30. 00/ton Au at \$130/ounce (7385 ppb). Also, the Homestake Mine, a large Au producer in South Dakota is located in carbonate facies iron-formation (Norton, 1974). Perhaps gold has a greater affinity for carbonate than for oxide facies iron-formations in this area. Ridler has argued that "Gold shows no favouritism among the major metallic constituents (i. e. Mg/Ca/Fe) of carbonate-rich rocks" (Ridler, 1976). This paper will strongly suggest that Ridler's argument may be correct not only for exhalites but for carbonate rocks in general, regardless of origin.

Origin of the Quartz-Carbonate Rocks

As reported earlier (Tihor and Crocket, 1976), there has been much disagreement concerning the origin of the Larder Lake "Break" quartz-carbonate rocks. They have been considered epigenetic replacement of diverse rocks along a major fault zone (Thomson, 1943), carbonate facies of the Boston Iron Formation

Table 75.2

Mean Concentrations

	Larder Lake Area: Ultramafic Volcanic Rocks	Larder Lake Area: Quartz Carbonate Rocks	Larder Lake Area: Mafic to Intermediate Volcanic Rocks	Analyses from the Literature: Munro Twp., Canada. Komatiitic Peridotites (Pyke *et al.*, 1973; Nesbitt *et al.*, 1976)	Analyses from the Literature: Helen Mine Carbonate Iron Formation (Goodwin, 1961, 1964)
Major Elements (%, reported on volatile-free basis)					
n	7	20	6	3	915
SiO_2	42.2	40.1	53.9	44.08	14.16
Al_2O_3	9.7	8.7	15.4	7.80	2.02
FeO	10.2	9.9	15.8	8.86	62.76
CaO	14.6	16.0	4.7	4.94	3.56
Na_2O	1.2	1.2	-	0.26	-
MgO	21.3	23.3	4.9	32.53	7.68
(MgO')*	31.6	35.0	-	-	-
TiO_2	0.51	0.47	2.2	0.23	0.53
MnO	0.29	0.34	0.37	0.14	3.59
Trace Elements (ppm)					
n	7	20	6	3	97
Cr	1847	1800	36	3333	30
Ni	731	700	66	894	20
Co	90	98	74	92	10
Cu	60	47	71	-	80
Sr	285	340	74	20	-
V	118	90	330	170	40
Y	49	36	66	11	-
Zr	41	36	160	20.2	170
Ratios					
Cr/Ni	2.53	2.57	0.55	3.73	1.50
Co/Ni	0.12	0.14	1.12	0.10	0.50

*MgO' = (MgO + CaO - 4.23), *see* text.
Total Fe as FeO.
Larder Lake Analyses by Emission Spectroscopy.

(Ridler, 1970), and carbonatized ultramafic rocks (Pyke, 1975a, b). The main purpose of the 1976 field season and subsequent laboratory work was to attempt to resolve these viewpoints.

Virtually every known outcrop of quartz-carbonate along the Larder Lake "Break" was examined in detail. A number of previously unreported occurrences were found and studied (northeast Eby Township; south-central Teck Township; northwest Otto Township; on Highway 66, near the Anoki Mine, central Gauthier Township). In addition, almost 3000 m (9000 feet) of diamond-drill core were logged and sampled through the kind co-operation of the mining and exploration companies working in the area.

Briefly, the field observations were:

(a) Virtually all varieties of rock occurring along the Larder Lake "Break" were subject to intense carbonate alteration. Syenitic dykes intruding carbonate zones were usually only weakly to moderately carbonated while quartz-feldspar porphyries and cherts appear to have been unaffected.

(b) The colour of the carbonate rocks tends to reflect the nature of the parent rock. Ultramafic rocks became grey, black or green carbonates; mafic and intermediate volcanics and greywackes were altered to grey carbonates; trachytes, syenites and arkoses changed to pink, buff or rarely, light grey carbonates.

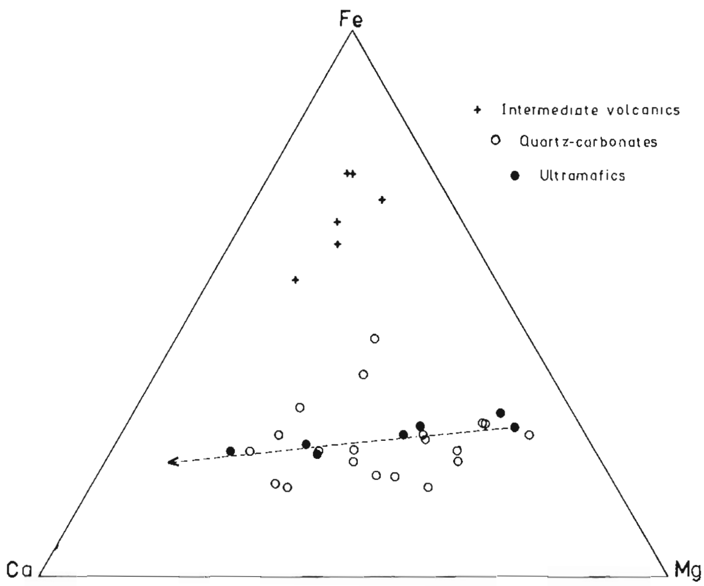


Figure 75. 3. Dashed line represents expected change of Ca/Mg/Fe by weight, with increasing molecular 1-for-1 replacement of Mg by Ca.

(c) The spatial relationships of the various carbonate rocks with each other is consistent and significant. Invariably where the quartz-carbonate zone is well developed, it consists of a "core" of emerald-green, fuchsite-bearing variety with margins of pink, buff, grey or black carbonates grading into the parent rock. This pattern is easily seen in continuous profiles such as in diamond-drill core and underground at the Kerr Addison Mine. On the 3800-foot level of the mine the sequence moving south to north, that is stratigraphically upward, is talc-chlorite schist – grey carbonate-rich talc-chlorite rock – pale-green sericite-carbonate rock – emerald-green quartz-carbonate rock – chlorite quartz-carbonate rock – grey carbonate rock – greywacke.

Similar patterns may be seen along Highway 66, about 800 m (2400 feet) west of Barber Lake, near the McVittie-McGarry townships boundary, and elsewhere, although observation is hampered by poor exposure.

It is important to note that carbonatization tends to be symmetrically gradational into the wall rocks, although the degree of carbonatization is a function of lithology. This behaviour contrasts with that of polymetallic base metal deposits and other exhalites. Proximal exhalites tend to exhibit strongly metasomatized footwall rocks but relatively fresh hanging-wall rocks, e.g., sulphide deposits of the Matagami, Quebec area (Sharpe, 1968).

(d) Although the quartz-carbonate rocks are generally highly sheared and brecciated, some primary textures and structures have survived. Where grey carbonates grade into greywacke, turbidite bedding is sometimes preserved. An example of this relationship is found underground at the Kerr Addison Mine. However, it must be emphasized that where this "replaced bedding" was found in Larder Lake "Break" quartz-carbonates, a gradational relationship could be demonstrated with clastic sedimentary rocks. The sedimentary structure is not a result of primary sedimentation of carbonate rocks. The type of chert-siderite or chert-magnetite repetitive interbedding typical of Archean iron-formations is conspicuously absent from the quartz-carbonate zone herein interpreted to be altered ultramafics.

Elongate polygonal structures reminiscent of polygonal jointing, flow breccias and hyaloclastites, are common in the quartz-carbonate rocks. These are similar in size and form to structures found in associated ultramafic flow rocks and suggest that the carbonate rocks exhibiting these features may be altered ultramafics. Examples are the small rock-cut on the north side of Highway 66 at the Cheminis Mine and on the small point of land at the west end of Bear Lake, McVittie Township.

Spinifex pseudomorphs were encountered in green quartz-carbonate from three locations: 70 m (200 feet) north of Highway 66 just east of the Misema River, Gauthier Township; in a diamond-drill hole 330 m (1000 feet) east of the Omega Mine, McVittie Township; and, on the surface at the Kerr Addison Mine in the hanging wall of the "glory-hole".

The spinifex in the carbonate from the drillhole is well preserved (cf. Pyke, 1975a) and closely resembles spinifex in a talc-chlorite rock which was intersected in another drillhole 180 m (550 feet) down strike (Fig. 75. 1).

Ultramafic Volcanic Rocks

The ultramafic nature of many of the outcrops of volcanic rock associated with the Larder Lake "Break" has previously been largely unsuspected. Brief references have been made to ultramafic rocks in the Kirkland Lake-Larder Lake area and ultramafic volcanic rocks intimately related to the ore zones in the Kerr Addison Mine (Jensen, 1974; Pyke, 1975b). We reported serpentine-bearing ultramafics from south-eastern Gauthier Township, the Omega Mine dump and the Cheminis Mine dump (Tihor and Crocket, 1976). However, while examining quartz-carbonate occurrences for this study it became apparent that there is a close spatial relationship between Larder Lake "Break" carbonates and ultramafic flow rocks in all but the central portions of the study area (Fig. 75. 2). The apparent absence of ultramafics here may be a consequence of poor exposure.

The ultramafic rocks are predominantly talc-chlorite in composition with lesser amounts of carbonate, serpentine and amphibole, in that order. Their mean chemical composition is shown in Table 75. 2.

Seven samples of ultramafics, 20 of quartz-carbonates and 6 associated intermediate to mafic volcanic rocks, were chosen at random along the Larder Lake "Break" and subjected to optical emission spectroscopy at the Geological Survey of Canada. Their mean chemical compositions are compared with that of three komatiitic

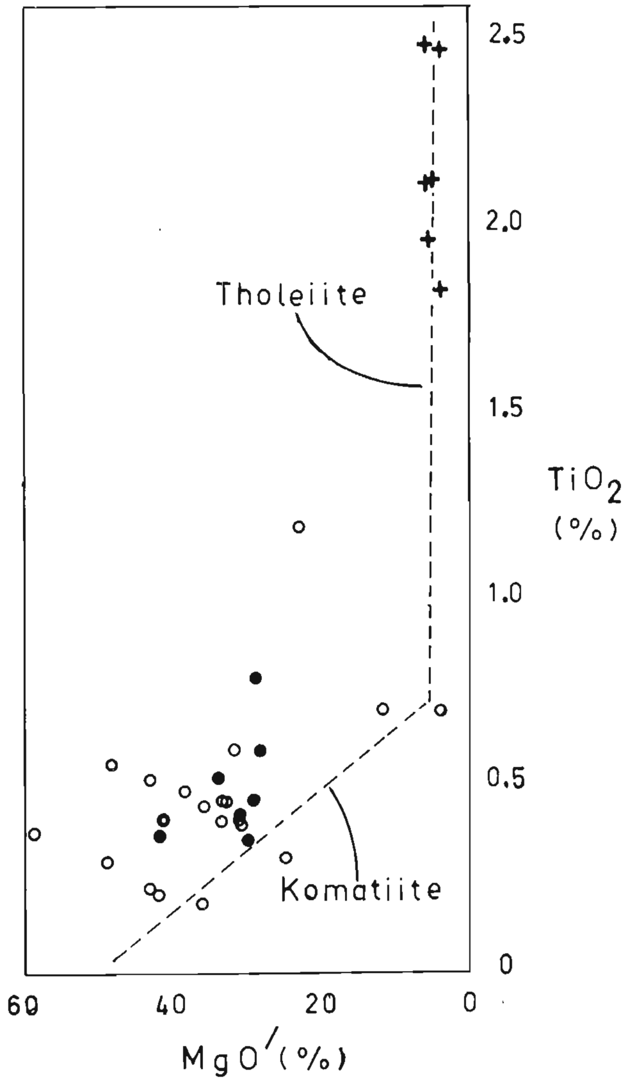


Figure 75.4. Wt. per cent MgO' vs. TiO_2 where MgO' = MgO+CaO-4.23 for LLB ultramafic and quartz-carbonate rocks (*see* text). For intermediate to mafic volcanics MgO'=MgO. Note that all analyses are recalculated to exclude H_2O and CO_2. Dashed lines represent komatiitic and tholeiitic trends (Arndt *et al.*, in prep.). Symbols are as in Figure 75.3.

peridotites from Munro Township, Ontario (Table 75.2). Also, because of the popular idea that the quartz-carbonates are carbonate facies exhalite (Ridler, 1970, 1976), the mean composition of a well-known carbonate facies exhalite, the Helen Mine carbonate iron formation, is included for comparison.

One is immediately struck by the very close similarity of the Larder Lake "Break" ultramafics to the quartz-carbonates with respect to all the major element oxides as well as the trace elements. It is also obvious that these rocks resemble the Helen Mine carbonate rocks only in titania content, all other elemental concentrations in the Table being radically different.

When the Larder Lake "Break" ultramafics and carbonates are compared with komatiitic peridotites from Munro Township, most elements are seen to be similar. However, there is a major difference in CaO and MgO contents. The possibility that this difference is the result of calcium metasomatism was explored.

The Larder Lake rocks were plotted by weight per cent on a Ca/Mg/Fe triangle (Fig. 75.3). The ultramafic rock containing the lowest concentration of CaO (4.23%) was chosen as representative of the ultramafic rocks previous to any possible metasomatism. This is not far from the 4.94% average CaO reported for the Munro rocks (Table 75.2). The dashed line (Fig. 75.3) shows the direction of change of Ca/Mg/Fe with progressive 1-for-1 molecular replacement of Mg by Ca. The line shows excellent agreement with actual compositional variations of the ultramafic rocks. We conclude that the original magnesia content of the ultramafic rocks and hence the quartz-carbonate rocks may be approximated by MgO'=MgO+CaO-4.23. MgO' for these ultramafics and quartz-carbonates (31.6 and 35.0 respectively) compares favourably with MgO for the Munro Township rocks (32.53).

In spite of the close agreement of the means of the compositions of the Larder Lake "Break" ultramafics and quartz-carbonates, individual analyses of quartz-carbonates show much greater compositional variation. This is apparent in Figure 75.3. We suggest that

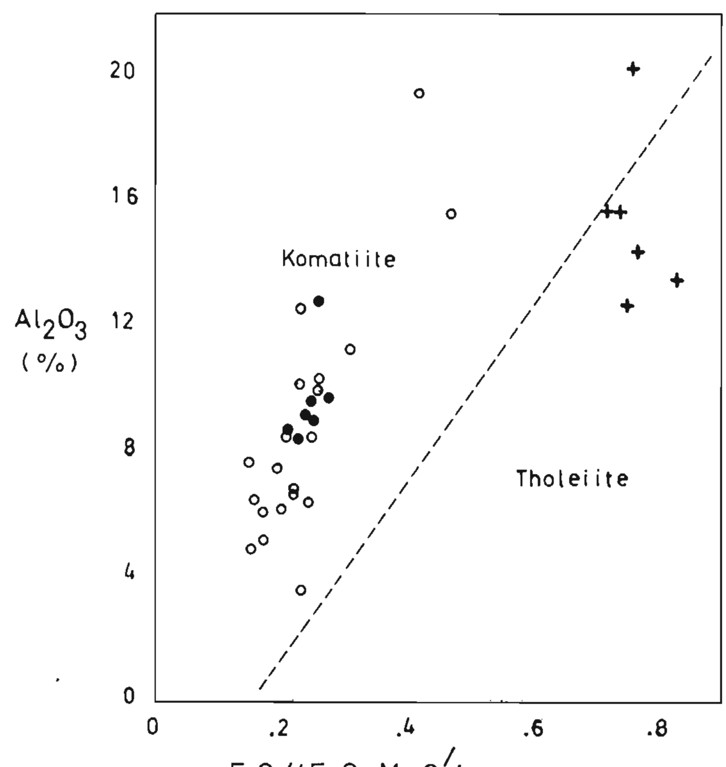


Figure 75.5. Wt. per cent FeO/(FeO+MgO') vs. Al_2O_3. Method of presenting data is as in Figure 75.4. Dashed line separates komatiite and tholeiite fields (Arndt *et al.*, in prep.).

during carbonatization of the ultramafics most elements may have undergone some degree of remobilization.

Arndt *et al.* (in prep.) found that komatiites in Munro Township could be conveniently distinguished from tholeiites by plotting Al_2O_3 against FeO/(FeO+MgO), and TiO_2 against MgO. When the Larder Lake "Break" rocks are plotted on a MgO' vs. TiO_2 diagram (Fig. 75.4) the ultramafic and quartz-carbonate rocks fall along but slightly above the komatiite trend, and the intermediate to mafic volcanic rocks plot exactly along the line representing the tholeiite trend. Since the intermediate to mafic volcanics show no evidence of significant loss of Mg or enrichment of Ca, MgO, is considered equal to MgO.

When the Larder Lake "Break" rocks are plotted on a FeO/(FeO+MgO') vs. Al_2O_3 diagram (Fig. 75.3) ultramafic and carbonate rocks again plot together and well within the komatiite field.

A conclusion that the Larder Lake "Break" ultramafics are komatiitic kindred would be premature. The analytical technique used, optical emission spectroscopy, may produce values with greater than 15 per cent error for elemental concentrations greater than 5 per cent. However, the evidence is highly suggestive that the peridotitic flow rocks of the Larder Lake "Break" are of komatiitic affinity and that most of the quartz-carbonate rocks are altered ultramafics.

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Projet 760008

R.C. Gauthier et V. Cormier[1]
Division de la science des terrains

Introduction

La cartographie des dépôts superficiels de la carte N.T.S. 21P à l'échelle 1: 250 000 a été entreprise au cours de l'été 1976. L'auteur et son assistant ont visité la majorité des routes carossables des cartes topographiques à l'échelle 1: 50 000, 21P/13, P/12, P/5, P/4, P/11 et P/14, couvrant ainsi une superficie de plus de 4470 km^2. Les sites visités étaient localisés et enregistrés au moyen d'une couverture photographique aérienne au 1: 20 000 (1974 et 1975). Les villes et villages sont groupés en bordure des côtes et le long de la rivière Miramichi. Les territoires centraux sont à peu près inhabités et partiellement accessibles; on y pratique la coupe du bois et on y exploite des mines de sulfures. Les coupes naturelles et artificielles dans les dépôts superficiels sont rares, rendant plus difficile l'interprétation des séquences sédimentaires.

Les premiers travaux de synthèse de l'histoire glaciaire locale ont été réalisés par Chalmers (1888 et 1895). Depuis lors, les géologues chargés de la cartographie de la roche en place ont ajouté des bribes d'information (voir la bibliographie). Le ministère des Ressources naturelles du Nouveau-Brunswick est impliqué dans une étude des réserves de dépôts granulaires (matériel d'emprunt de la région de Bathurst et de Belledune). Plusieurs travaux ont été réalisés par ce groupe (voir la bibliographie) et un inventaire détaillé des réserves est actuellement en cours dans la région des cartes P/13 et P/12 ouest (moitié nord). Une évaluation des types et des réserves de tourbes est également en voie de réalisation par le même ministère, dans le secteur de Shippegan (Korpijaakko, 1975). Les travaux de cartographie entrepris par l'auteur principal se continueront l'été prochain et fourniront éventuellement une étude exhaustive des dépôts superficiels de la carte topographique au 1: 250 000, 21P.

Divisions physiographiques

Une distinction géologique majeure permet de diviser en deux zones distinctes la région cartographiée: les hautes terres de l'extrémité ouest, et les basses terres qui recouvrent près de 80% de la région. Les hautes terres, composées de roches cristallines et métamorphiques de l'Ordovicien, du Silurien et du Dévonien, forment un plateau disséqué, incliné vers l'est. L'altitude maximale de cette surface atteint 290 m (950 pieds). Le système fluvial est contrôlé par la fracturation et la position des contacts lithologiques des séquences de roches. La Népisiguit et la Miramichi Nord-Ouest sont les deux principaux exutoires fluviaux de la région. Le Quaternaire des régions des hautes terres se caractérise surtout par un travail érosif des glaciers. Plusieurs générations de mouvements glaciaires ont été préservés sur les surfaces polies et striées des roches. Les dépôts de contact glaciaire et fluvio-glaciaires sont restreints aux régions basses des vallées.

Les régions des basses terres sont composées de séquences rouges, grises et vertes de grès et de shale faiblement consolidés d'âge carbonifère. Ces séquences sédimentaires reposent à plat et aucune déformation majeure ne bouleverse ces structures horizontales. Un dyke de diabase rectiligne d'âge triasique recoupe les séquences sédimentaires plus anciennes. Le relief moyen est peu accentué variant entre 70 et 120 m (250 à 400 pieds). Quelques rivières mineures sont profondément encaissées, jusqu'à des profondeurs de plus de 30 m par rapport à la surface régionale. Ces vallées étroites sont principalement le résultat de l'érosion préglaciaire du réseau fluvial. Dans les basses terres une couverture de till rougeâtre et silteux repose sur une unité sableuse ou graveleuse d'épaisseur variable, composée exclusivement des débris de la roche locale. Cette unité graveleuse qui repose sur la roche de fond représente un dépôt préglaciaire faiblement remanié.

La région côtière du nord de la carte est rectiligne. A l'est de Bathurst, les séquences de grès carbonifères forment une falaise en contact avec la mer avec une dénivellation maximale de 30 m de hauteur. A l'ouest de Bathurst, la roche contrôle la géométrie de la côte et le contact est marqué par un faible escarpement rocheux de un à trois mètres de hauteur. La région du havre de Bathurst est une zone de sédimentation marine active; deux systèmes de flèches littorales progressent vers l'intérieur du havre. L'accumulation de sédiments est plus importante dans cette zone côtière intérieure, causant ainsi des problèmes pour l'entretient des chenaux navigables.

Directions de l'écoulement glaciaire

Dans les régions des hautes terres, la dernière phase glaciaire majeure accuse un mouvement vers l'est des glaces. Une abondante distribution des stries dont le sens est déterminé (plus de 200 sites d'affleurements striés ont été découverts) et de roches moutonnées démontrent que ce mouvement est responsable du modelage de la surface rocheuse de ces régions.

Plusieurs systèmes de stries sont antérieurs à cette pulsion principale, quoique leur chronologie relative n'ait pu être définie l'une par rapport aux autres; ce sont les mouvements nord-nord-est

[1]Etudiant gradué. Université de Bretagne occidentale, Brest, France.

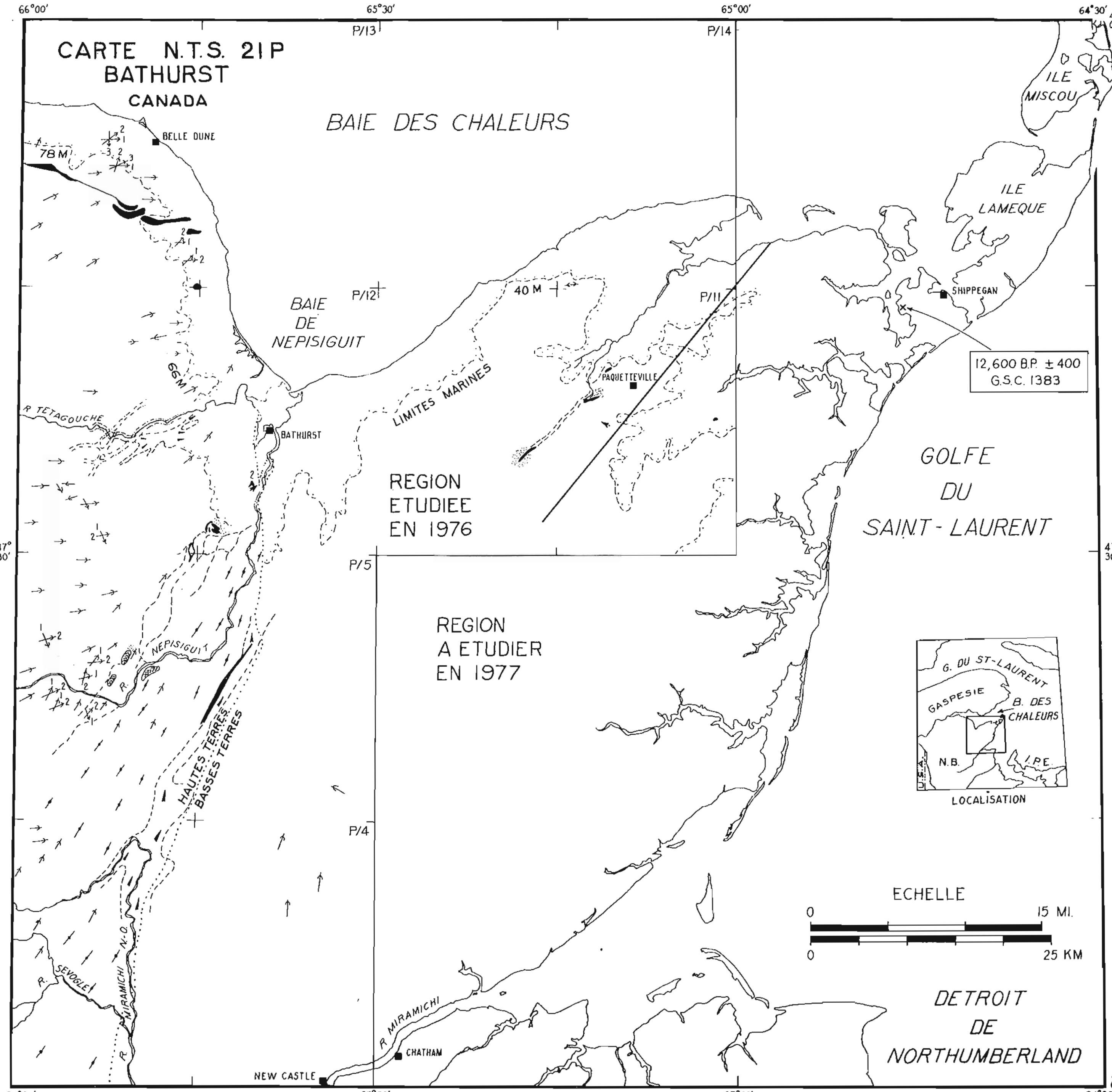

Figure 76. 1. Région étudiée en 1976.

(abondantes observations où le mouvement nord, conservé sur des faces oxydées et protégées, est recoupé par le mouvement est), sud-sud-est (5 sites de stries dont la chronologie relative montre le recoupement du mouvement sud par le mouvement est), et le mouvement ouest (3 sites striés, dont un définit une chronologie relative où le mouvement ouest est antérieur au mouvement est).

Le mouvement glaciaire principal vers l'est est suivi par deux mouvements postérieurs. Dans la région de Belledune, une déflexion du mouvement de 090° laisse une séquence chronologique où ce mouvement principal est suivi d'un mouvement est-nord-est plus jeune. Les surfaces striées à 070° sont superficielles et n'altèrent pas les formes des roches moutonnées orientées vers l'est, forme déjà acquise au cours de la

LEGENDE

Carte d'interprétation des événements glaciaires

Sédiments de contact glaciaire: moraines frontales, latérales, terrasses de kame, dépôts glacio-marins, kames, kettles

Sédiments fluvio-glaciaires

eskers, sens de l'écoulement inféré

plaine d'épandage de contact glaciaire, épandage sous-glaciaire

Chenaux d'eau de fonte glaciaire, chenaux proglaciaires

Sens de l'écoulement des eaux de fonte

Chenaux préglaciaires

Surfaces drumlinisées

Synthèse des stries glaciaires de la localité environnante 1, 2: chronologie relative perçue

Trame du till

Limite de submersion marine 78, 66, 40 m: altitudes maximales perçues

Division physiographique majeure: hautes terres et basses terres

Dyke de diabase: dyke de Caraquet

Limites de la région étudiée, été 1976

Limites des cartes topographiques N. T. S. au 1: 50 000

P/S Numéro de la carte N. T. S. au 1: 50 000

× Localisation d'une datation au radiocarbone dernière phase glaciaire principale. Le système de stries à 090° est aussi souvent préservé.

Un lobe de glace indépendant tardif a laissé dans la vallée de la Népisiguit et au sud de cette dernière de nombreuses traces d'un mouvement glaciaire vers le nord-nord-est. Le développement d'une surface drumlinisée et une distribution abondante de stries attestent de l'action de ce lobe. Ce mouvement est davantage orienté vers le nord-est (direction 030-040°) dans les secteurs sud-ouest de la carte P/5. Cette lente transformation de la direction d'écoulement suggère la présence de la source glaciaire dans la région des hautes terres à des distances relativement proches de la zone d'écoulement décrite sur la carte 21P. Des sites de polis glaciaires montrent des stries liées au mouvement glaciaire majeur vers l'est et qui sont plus vieilles que les stries à 020°. Deux tills superposés, localisés à 6 km au sud de Bathurst dans lesquels on a réalisé des trames du till, ont permi de rattacher les dépôts glaciaires aux séquences d'avancées des glaces vers l'est et le nord; cette séquence est la même que celle définie au moyen des stries. Le till inférieur gris avec une proportion élevée de cailloux s'associe à l'écoulement glaciaire vers le nord-est. Le till supérieur rouge, silteux, déposé dans un environnement marin, comme en témoigne la concentration des particules fines, est relié à une progression des glaces vers le nord. La présence des grès rouges localisés exclusivement au sud et à l'ouest du point d'observation est responsable de la teinte rouge du till supérieur.

A l'intérieur des limites de submersion marine, plusieurs systèmes de stries sont présents; aucun d'eux ne semble avoir une signification glaciaire réelle, la majorité étant sans doute le résultat de l'action des glaces flottantes (Chalmers, 1895) après le retrait des glaciers. Les divers systèmes de stries (sud, est, nord, nord-est) sont souvent superposés au système

Figure 76. 2.

Sédiments de contact glaciaire, 11 km à l'ouest de Belledune. La crête morainique à une hauteur totale de 20 m. (GSC 4-3-76)

Figure 76. 3.

Sédiments fluvio-glaciaires d'une terrasse de kame à 8 km au sud de Belledune. L'épaisseur minimale observée de cette séquence est de 26 m. (GSC 3-33-76) de stries majeur, relié à la dernière phase glaciaire principale, le mouvement est orienté vers l'est.

Les régions des basses terres ne conservent qu'exceptionnellement les traces de stries et de polis glaciaires. Cinq sites seulement ont été découverts, ces surfaces de grès faiblement consolidés se prêtant mal à la conservation des polis glaciaires. Au sud de la région étudiée (fig. 76.1; carte 21P/4E), quatre sites de roches striées, dont le sens de l'écoulement était exprimé, montrent une progression glaciaire vers le nord et vers l'ouest. Ce mouvement nord est cependant d'un âge et d'une signification différents par rapport à celui conservé dans la vallée de la Népisiguit.

Un seul site de roches striées a été trouvé dans les régions nord des basses terres; le sens des stries n'a pu être déterminé. Le mouvement glaciaire est orienté est-ouest; on ignore pour le moment sa signification quant à l'histoire glaciaire de ce secteur, les deux sens d'écoulement semblent plausibles.

Sédiments glaciaires

Dans les régions des hautes terres, deux complexes indépendants de sédiments glaciaires témoignent des dernières pulsions glaciaires.

Région de Belledune

Un complexe de contact glaciaire majeur s'est développé dans la région de Belledune, principalement à l'extérieur des limites de submersion marine. Il est formé de séquences de sédiments de contact glaciaire (fig. 76.2), fluvio-glaciaires (fig. 76.3) et glacio-marins.

Figure 76.4.

Complexe morainique frontal du <<Blue Mountain>> à 14 km au sud de Bathurst, dans la vallée de la Népisiguit. Les séquences graveleuses passent du contact glaciaire au fluvio-glaciaire. L'épaisseur maximale des graviers atteint localement près de 65 m. (GSC 1-22-76)

Figure 76.5.

Moraine latérale de la Népisiquit à 28 km au sud de Bathurst. Une crête double, d'une largeur de plus de 235 m et d'une hauteur supérieure à 17 m est constituée majoritairement de gravier fluvio-glaciaire, localement avec un tri pauvre. (GSC 2-10-76)

L'épaisseur des dépôts est localement supérieure à 40 m. Les formes de dépôts sont des crêtes morainiques, des terrasses de kame et des terrasses marines de sédiments glaciaires remaniés. La formation du complexe morainique est postérieure au dégagement des glaces de la baie des chaleurs. La glace se situait au sud du complexe morainique. Cette position glaciaire marque une phase de stabilisation du dernier glacier dont l'écoulement se faisait vers l'est.

Tableau 76.1

SYNOPSIS

Chronologie des séquences glaciaires

Phases glaciaires	Direction de la progression	Evénements
		Mer à New Richmond
Lobe de la baie des Chaleurs	Est	N'affecte pas la carte 21P
		Complexe de terrasses de kame, 210
?		Mer à Jacquet River
Lobe de la Népisiguit	Nord-Nord-Est	Lobe indépendant provenant du sud-ouest
Blocus de la Miramichi	Nord-Ouest	Provoque l'inversion du drainage de la Miramichi Nord-Ouest
Déflexion	Est-Nord-Est	Complexe glaciaire de Belledune
		Ouverture de la baie des Chaleurs
		Déglaciation de Shippegan
Glaciation principale	Est	Région des hautes terres
	Ouest?	Région des basses terres
		Interstadiaires?
?		
Vieilles glaciations	Sud-Est	
	Ouest	
	Nord	Stries préservées sur des faces polies oxydées

Vallée de la Népisiguit

La vallée de la Népisiguit et la région au sud de cette dernière ont été occupées par un lobe tardiglaciaire indépendant, s'écoulant vers la baie de Népisiguit. Ce lobe représente une réavancée glaciaire mineure prenant source dans les hautes terres du sud-ouest de la région étudiée.

La dispersion des stries et une surface fortement drumlinisée sont associées à ce mouvement tardif vers le nord. Un complexe de sédiments de contact glaciaire massif, (fig. 76.4) le <<Blue Mountain>> (14 km au sud de Bathurst) est relié à une séquence de moraine frontale de ce lobe, mise en place au cours de la régression glaciaire. Localement, l'épaisseur des graviers de contact glaciaire et fluvio-glaciaires est supérieure à 60 m. L'extension maximale du lobe de la Népisiguit n'est pas connue, mais il s'étend au-delà de ce complexe morainique, jusqu'à l'intérieur de la zone de submersion marine, comme en témoigne une séquence de deux tills distincts antérieurement décrite (6 km au sud de Bathurst).

Un complexe morainique latéral (fig. 76.5) d'extension et de dimension limitées, marque la limite maximale est du lobe glaciaire de la Népisiguit. En contact avec cette crête morainique, du côté distal, un réseau important de drainage proglaciaire s'est développé, conduisant les eaux de fonte jusqu'à la baie de Népisiguit.

L'extension ouest du lobe est moins bien définie. De multiples réseaux de chenaux d'écoulement proglaciaires, capturés par le réseau de drainage postglaciaire, attestent de l'amincissement progressif du lobe glaciaire. Grâce à la dispersion des stries, à l'étendue des surfaces drumlinisées et des réseaux de drainage proglaciaire, il est possible d'évaluer à près de 15 km la largeur du lobe glaciaire dans la région de la Népisiguit. Plus au sud, la dimension du lobe est plus importante. Des séquences de sédiments fluvioglaciaires reliées au lobe tardif (eskers, complexes d'eskers, épandages sous-glaciaires) suggèrent une désintégration progressive des glaces au cours de la fonte, après une période de progression active dans la vallée de la Népisiguit.

Région de Paquetteville

La région de la carte P/11 possède des vestiges de sédiments de contact glaciaire, de chenaux proglaciaires et de plaines d'épandage rattachés possiblement à une position frontale de la glace. Les structures des dépôts de contact glaciaire, telles l'orientation des stratifications et la position de chenaux de contact glaciaire supramorainiques, suggèrent que la glace se situait du côté sud-est du complexe morainique. Les quatre sites de roches striées dans la région au sud-ouest du complexe définissent des mouvements vers le nord et vers l'ouest, reliés à ce glacier provenant des régions du sud-ouest. La dispersion des erratiques de diabase et une trame du till supportent également l'idée d'un écoulement glaciaire vers l'ouest. Cette hypothèse sera développée par des travaux ultérieurs au cours de l'été prochain.

Il est à noter que les îles Miscou et Lamèque conservent de nombreuses évidences du passage des glaciers. Une séquence de till, recouverte de dépôts marins, observée en maintes localités sur les îles, permet de croire que ces régions ont été occupées par les glaciers. Il serait prématuré pour le moment de déterminer avec certitude la provenance et l'âge de ces glaces.

Evolution des réseaux fluviaux

La Tétagouche

Le réseau fluvial de la Tétagouche illustre un intéressant mécanisme d'évolution des chenaux préglaciaires, glaciaires et post-glaciaires. Le drainage préglaciaire de la rivière était contrôlé par les réseaux structuraux des failles et des contacts lithologiques de la roche. La rivière, alors divisée en deux embranchements majeurs, s'écoulait à l'intérieur de vallées étroites et profondément creusées dans le roc. Au cours de la fonte des glaces, de nouveaux chenaux se sont développés et la partie inférieure du cours de la vallée préglaciaire a été comblée par les sédiments glaciaires. Le tracé glaciaire de la rivière a formé une large plaine alluviale à l'extérieur des limites de la submersion marine. L'évolution post-glaciaire de la rivière lui a permis de retrouver la vallée préglaciaire du cours inférieur, jadis comblée par la sédimentation glaciaire, mais la rivière n'a jamais pu retrouver la partie du cours supérieur de la vallée préglaciaire. La configuration actuelle de la Tétagouche est une fusion des vallées glaciaires et préglaciaires, laissant libres les vallées préglaciaires des cours supérieurs ainsi que la surface alluvial glaciaire du cours inférieur. Les chenaux préglaciaires du cours supérieur utilisés aujourd'hui n'ont pas été comblés par la sédimentation glaciaire et sont occupés par des réseaux de ruisseaux mineurs.

La Miramichi Nord-Ouest

La Miramichi Nord-Ouest a été influencée par un blocus glaciaire de son cours inférieur à l'époque où le lobe glaciaire de la Népisiguit fermait cette vallée, si bien que l'écoulement à l'époque glaciaire a dû s'effectuer vers le nord, en contact avec le lobe tardiglaciaire de la Népisiguit. La configuration anormale en "U" de la rivière post-glaciaire et les larges plaines alluviales abandonnées témoignent de cet écoulement glaciaire. Le cours inférieur de la Miramichi Nord-Ouest a enregistré une inversion de l'écoulement fluvial après le dégagement de la vallée majeure de la Miramichi. Le partage des eaux vers le nord se situe à une altitude de 75 m (250 pieds) et se trouve localisé du côté est de la moraine latérale du lobe glaciaire de la Népisiguit. Une quantité appréciable de sédiments fluvio-glaciaires a été mise en place le long de cette vallée de contact glaciaire. Une vaste surface estuairienne, à la limite de la submersion marine dans la baie de Népisiguit a été construite par le détournement temporaire de la Miramichi Nord-Ouest vers le nord.

Limite de l'invasion marine

Le long de la côte nord de la région étudiée, la limite de la submersion marine maximale passe, d'ouest en est, de 78 m d'altitude (250 pieds) à 40 m (125 pieds), sur une distance de 80 km (50 milles). Il est souvent difficile de localiser précisément sur le terrain la position exacte des limites maximales de submersion car l'action marine à ces altitudes élevées est restreinte. Localement, des lignes d'érosion, des tills remaniés et des surfaces d'aggradation à l'intérieur de vallées permettent de déterminer ces altitudes. La dépression isostatique, résultat possible d'une couverture glaciaire plus épaisse, s'accentue vers l'ouest à un rythme de 0.47 m par km (2.5 pieds par mille).

Dans les zones de submersion plus basses (moins des 2/3 de la valeur de la submersion maximale), une couverture de dépôts de plage et de sédiments lagunaires, généralement mince, est présente sur le till ou la roche en place. L'embouchure de la vallée de la Népisiguit est le seul endroit où la sédimentation marine est importante; des séquences argileuses et sableuses varient en épaisseur de 30 à 100 m (100 à 330 pieds) (valeur maximale suggérée par les forages). La surface rocheuse est souvent inférieure au niveau marin actuel, ce qui suggère une dimension préglaciaire de la baie de Népisiguit, telle que contrôlée par la surface rocheuse, nettement plus grande que celle existant actuellement. La baie de Népisiguit, au maximum de la submersion marine, était un exutoire important de plusieurs rivières formées par les chenaux de fonte des glaciers.

Datations au radiocarbone

Aucun échantillon de matériel organique valable pour une datation au radiocarbone n'a été recueilli cet été. Les milieux les plus prometteurs pouvant fournir un matériel adéquat à la datation sont les larges étendues de tourbe présentes sur l'ensemble des zones d'émergence marine. Un procédé manuel de forage permettrait d'atteindre la base de ces séquences, fournissant ainsi un âge minimal de l'invasion marine et du retrait glaciaire. On prévoit de telles cueillettes pour la prochaine saison. Trois datations publiées permettent d'évaluer l'âge du retrait glaciaire principal de la région 21P. Thomas *et al.* (1973) ont dressé une courbe des changements post-glaciaires du niveau marin dans la région de Shippegan. Des fossiles marins, datés à 12 600 ± 400 ans (GSC-1383), localisés à une altitude de -0.6 m, fournissent la base d'une courbe extrapolée de régression relative de la mer, débutant il y a 14 000 ans, immédiatement après le retrait glaciaire. Des fossiles marins provenant de la région de Jacquet River (limite nord-ouest de la carte 21P; fig. 76.1), localisés à une altitude de 13 m donne un âge minimal de 12 500 ± 170 ans (GSC-1557, comm. pers., Grant, 1976) de la présence marine dans cette région. Une dernière date de fossiles marins de la région de New Richmond (Québec, rive nord de la baie des Chaleurs) suggère la valeur 12 000 ± 180 ans

(GSC-1018, altitude: 0.6 m). Cette datation serait reliée à la présence d'un lobe glaciaire tardif à l'intérieur de la baie des Chaleurs. Ce lobe n'a toutefois aucune influence perceptible sur la déglaciation de la région de la carte 21P.

Conclusions

Les complexes de sédiments glaciaires des hautes terres sont le résultat du retrait tardif des glaces, lié à l'ouverture précipitée de la baie des Chaleurs. Une chronologie complexe de différents mouvements glaciaires définie au moyen des stries glaciaires suggère une succession de plusieurs pulsions glaciaires, dont peu d'évidences sédimentologiques ont été conservées. Le drainage glaciaire régional diffère sensiblement du drainage actuel et a laissé des traces visibles de son influence sur le développement du drainage post-glaciaire et des séquences sédimentaires estuairiennes et marines.

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Project 740054

Nelson R. Gadd
Terrain Sciences Division

Twenty-seven new stratigraphic borings, mainly of clay and silt sediments in the Ottawa Valley portion of Champlain Sea, were made during the 1976 field season. They reveal a fairly consistent suite of sedimentary facies. Visual inspection and photographic record has been made of more than two hundred Shelby tube cores taken at close intervals through the entire suite of clay-silt sediments; laboratory study of representative samples is in progress.

As reported by Fransham and Gadd (1976), at least four principal sedimentary facies that may be recognized through subtle visual differences are present:

1) At the base of the sedimentary column, commonly overlying till, is a varved clay facies with characteristic colour banding and graded bedding; this grades upward into

2) a deep-water marine facies comprising vaguely stratified to massive, dark blue-grey clay and silty clay. This is commonly mottled throughout by unstable black sulphurous material. The mottling in places shows an irregular pattern reminiscent of worm burrows or other organic disturbance of the sediment. This facies is commonly fossiliferous, containing a number of Champlain Sea species (pelecypods, foraminifera, ostracodes). The relatively massive clay facies grades upward into

3) regularly banded, coarse grey silty clay and fine red clay with some bands of sulphurous black material in discrete layers; the relative thickness of the grey-red sedimentary couples (representing coarse-fine sediment couples) increases upward in the section. Grain size of materials also shows a general increase upward in the section.

4) The fourth facies is in the clay to sand range, with sand dominant in the uppermost strata. It is distinguished mainly on the basis of structures such as erosional breaks in sedimentation, distortion of beds due to intraformational slumping, unstratified units possibly representing turbidity flow or subaqueous debris sliding, and by the presence of discrete thin layers of relatively coarse sediment; these layers are made up of coarse silt in the lowermost or clay-silt phase of this unit and of fine to medium sand in the upper sandy phase.

The whole sedimentary package in many cores is overlain by a fluvial sand unit of considerable thickness and possibly also by a fluvial or lacustrine silt unit, the latter being particularly common on some low terraces near modern main streams.

Interpretation of the visual data alone allows a working hypothesis that the basin has received fine sediment under four principal environmental conditions: 1) conditions similar to those of a freshwater glacial lake; 2) deep-water, quiescent marine basin or pro-delta environment (maximum salinity conditions); 3) bottom-set facies of a prograding delta; and 4) upper delta facies of a prograding delta.

From the fairly regular increase upward in the grain size of the sediments, and from evidence of increasing current activity, it is obvious that the energy level of the sequence of sedimentary environments progressed in a regular manner from a low-energy state of the deep-water basin to a relative high-energy state in the fluvial environment of the upper delta. Such a suite of environmental changes is compatible with isostatic rebound of the region and with consequent offlap of Champlain Sea.

In previous work Richard (1975, 1976) has observed abundant macrofossils in till-like material and has postulated that a glacier readvanced through Ottawa Valley at about 11 200 years B.P. As this was about mid-term in the Champlain Sea occupation of the valley, he concluded that fossils incorporated in till-like materials, now lying mainly south of Ottawa Valley, had been picked up from pre-existing fossiliferous material and had been incorporated in till. Because Ottawa Valley lies upglacier from most occurrences of "fossiliferous till", it follows that evidence of glaciation should form a hiatus in the sequence of marine and postmarine sedimentation, and that the glacial event should separate bodies of sediment respectively older and younger than 11 200 years B.P.

Because of this postulate, the author and colleagues were careful to note and sample all lithologic changes in the numerous complete stratigraphic sections of unconsolidated materials encountered in the drilling programs of 1975 and 1976. In no boring was there evidence of such glacial readvance or of a significant hiatus in sedimentation. The borings in Ottawa Valley reveal, on the contrary, that a single, thin till sheet discontinuously overlies bedrock. Since deposition of this till, the Ottawa Valley basin has received the apparently continuous suite of nonglacial sediments briefly described above.

It is assumed by the author that the thick bodies (as much as 100 m (300 feet)) of sediment represent sedimentation during the entire period of marine submergence of this region and therefore should range in age up to as much as 12 800 years B.P., the age established by Richard (1975, 1976) for the highest marine beach southwest of the city of Ottawa. Again, there is no record in these sedimentary suites of a younger glacial event, viz. at 11 200 years B.P. It would seem reasonable, therefore, to entertain working

From: Report of Activities, Part A;
Geol. Surv. Can., Paper 77-1A (1977)

hypotheses other than glacial readvance through Ottawa Valley to explain the presence of fossils in till-like material found outside Ottawa Valley.

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DENDROCHRONOLOGIC INVESTIGATION OF HIGH-WATER EVENTS ALONG HODGSON CREEK, DISTRICT OF MACKENZIE

Project 750077

P.A. Egginton and T.J. Day
Terrain Sciences Division

Introduction

One of the problems associated with assessing the integrity of design structures along Mackenzie Highway is determining the magntitude, frequency, and cause of high-water events on tributary streams of Mackenzie River. Interpretation of high-water events based on indirect evidence (debris lines, scarred trees) at highway crossing sites within 3 to 4 km of Mackenzie River is complicated by the presence of Mackenzie backwater. The cause of backwater from Mackenzie River is generally indeterminate without the aid of extensive stage data at the particular site of interest, as backwater can result from spring breakup flood, ice damming during breakup, or summer rainfall events. Not withstanding this problem, however, this reconnaissance study was designed to investigate the available indirect evidence in the hope of determining relationships between age and height of high-water events along a tributary of the Mackenzie and to use these data to develop a maximum high-water envelope.

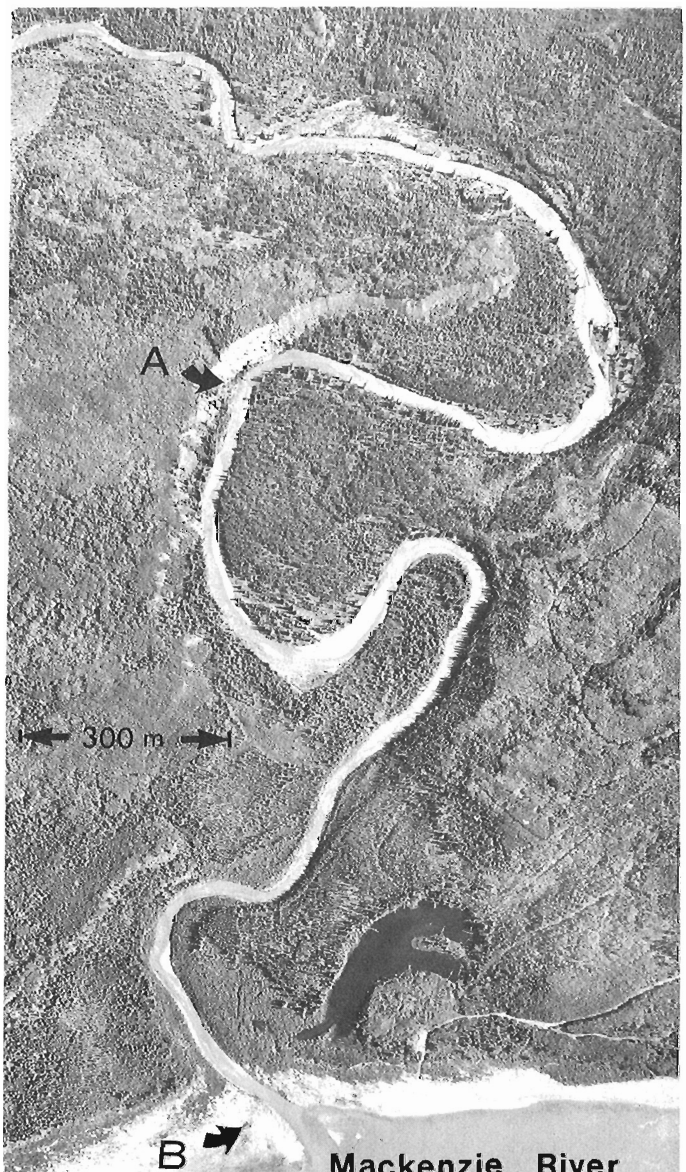

Figure 78.1. Aerial photograph of Hodgson Creek delimiting the river reach surveyed (A to B).

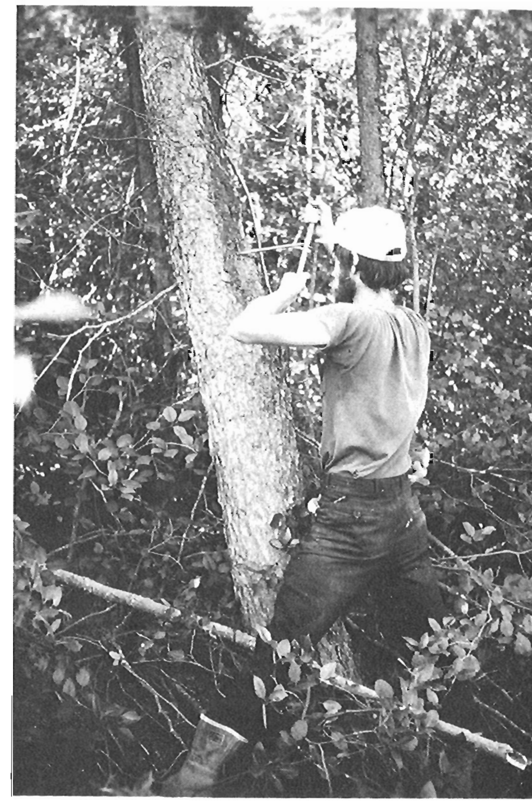

Figure 78.2. An increment borer used to obtain sample cores from a typical ice scar along Hodgson Creek.

Field Area

The tributary stream chosen for study was Hodgson Creek (highway mile 436.5, km 706), just one mile north of the town of Wrigley, Northwest Territories

From: Report of Activities, Part A;
Geol. Surv. Can., Paper 77-1A (1977).

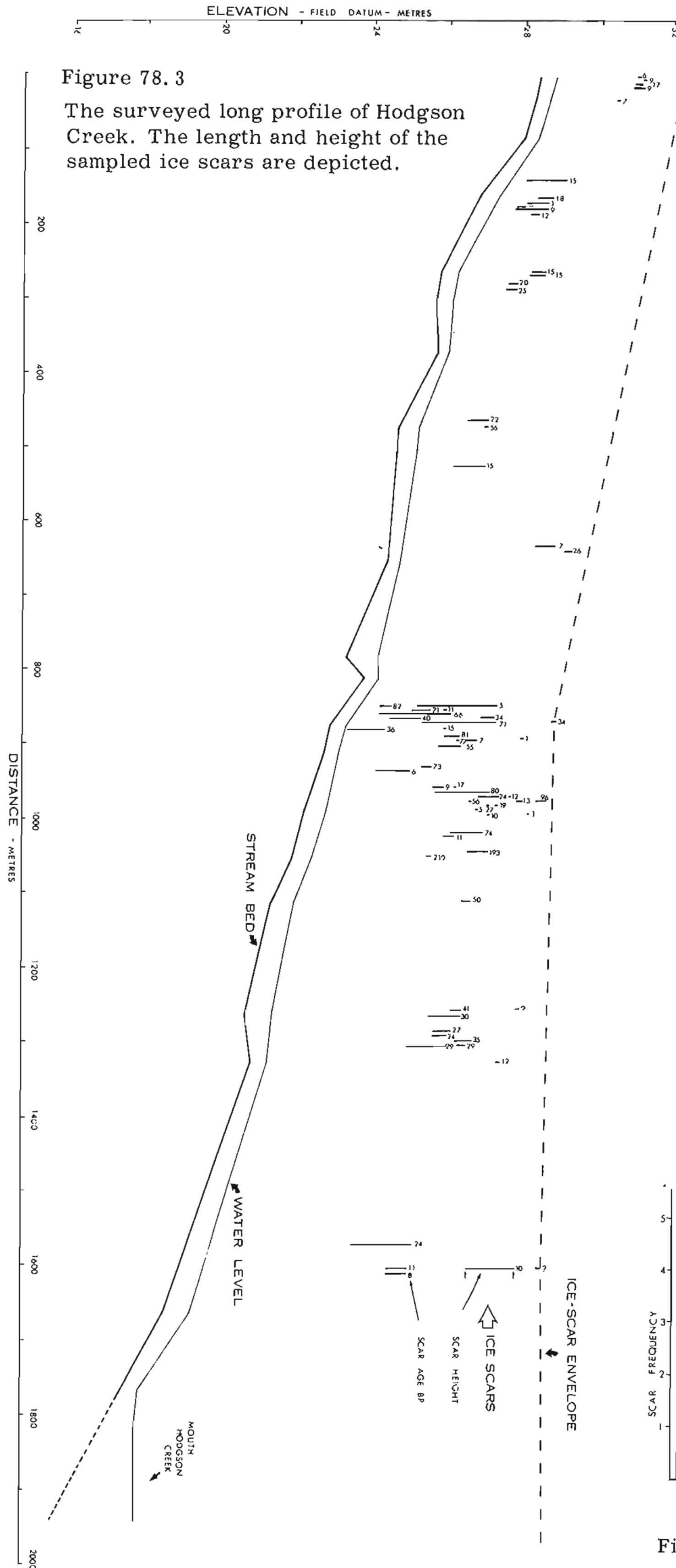

Figure 78.3

The surveyed long profile of Hodgson Creek. The length and height of the sampled ice scars are depicted.

Hodgson Creek at its confluence with Mackenzie River, has a drainage area of 321 km^2. The stream appears to be vertically stable and exhibits an irregular meander pattern (Fig. 78.1).

Field Method

The field study was undertaken along the southern bank of Hodgson Creek, over a length of 2 km (between A and B in Fig. 78.1). Restriction of the study to only one bank and for this distance was dictated by time limitations.

Along the southern bank, evidence of high-water included flood debris (sediment marks on trees, debris hanging in trees), bends in trees, and scars on trees which were the most common. These scars were considered to be formed by the impact of floating ice blocks during high-water. Henoch (1973) dated similar scars and was able to determine the frequency and elevation relationships for high-water events along Mackenzie River.

Once an ice scar was identified (any scars near fallen trees were ignored), an increment borer was used to extract a core from the scar face, through the centre pith, out to the opposite side (Fig. 78.2). By comparing the number of tree rings on either side of the pith, an estimate of the age of the scar can be made. Errors in tree ring dating can occur through not coring the pith, locally absent or false rings, and counting errors (Schulman, 1956).

At the beginning of the study an attempt was made to sample all identifiable scars; however, time restraints resulted in sampling at irregular intervals along the channel. In all, 75 cores were extracted and tied into a survey of the long profile of Hodgson Creek (Fig. 78.3).

Field Data and Discussion

The position of the measured scars and the long profile of Hodgson Creek are shown in Figure 78.3 together with the age of each scar, its length, and its height above stream bed elevation. The longitudinal frequency of the scars is a function of the sampling design and the density of trees at the sampling site. Ages of scars range from 1 year to 210 years B.P. The

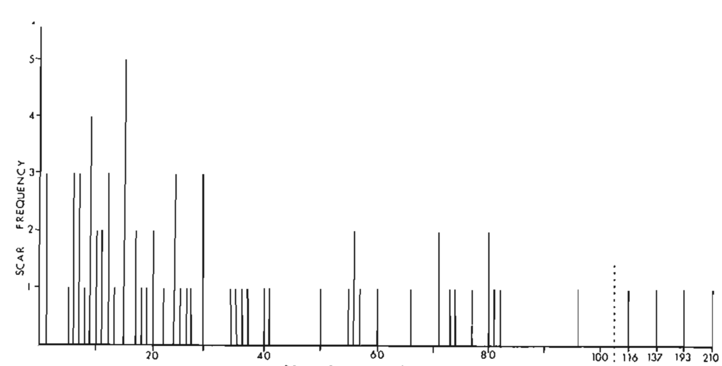

Figure 78.4. The age-frequency distribution of sampled scar on Hodgson Creek.

Figure 78.5.

The coring of a tree near the mouth of Hodgson Creek; the scar is located 9.7 m above the present stream bed. Note the dirt trim line and scarring on adjacent trees (indicated by arrows).

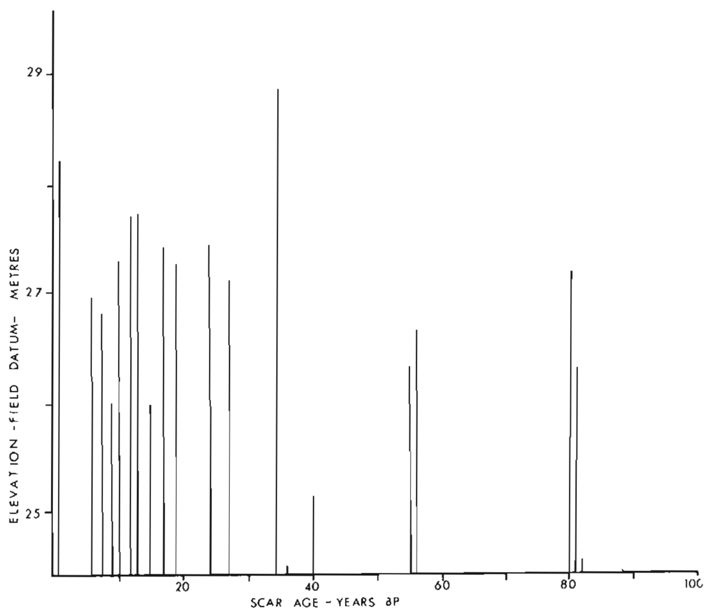

Figure 78.6. The maximum elevation of ice scars of a given age for a limited section (850 to 1000 m, Fig. 78.3) of Hodgson Creek.

age-frequency distribution is shown in Figure 78.4; two thirds of the scars dated are younger than 40 years BP. The skewed distribution does not necessarily result from either a sampling bias (i.e. difficulty in identifying older scars) or a considerable increase in scarring frequency within the last four decades. An examination of aerial photographs (i.e. Fig. 78.1) reveals that considerable portions of Hodgson Creek have been burnt recently, and this is viewed as a limiting factor to tree and ice-scar presence.

The distribution of the ice scars (Fig. 78.3), which generally increases in elevation above river level with distance downstream, indicates that the majority of scars result from Mackenzie backwater events; scarring is evident 9 m or more above present stream bed elevation in the vicinity of the stream mouth (Fig. 78.5). It is impossible, using the limited data collected to contour scars of equal age along the profile. Even grouping scars into 5-year periods (the suspected level of measurement precision) does not preduce identifiable trends.

The variation of maximum scar height above river level with scar age for the south bank between 850 and 1000 m of the long profile (Fig. 78.3) is shown in Figure 78.6. Normal frequency analysis, i.e. return period for ice scarring, is considered inappropriate because records are incomplete.

Conclusions

The sporadic nature of the scar ages may result in part from sample design as only a small portion of the total population was sampled. Scar formation is complex and depends upon the scale and frequency of ice blocks, stream flow, and ice paths. Rapid changes in river levels are associated with backwater events, and scarring can occur over a large elevation range within the same time scale. For this reason, even with a considerably larger sample, it may not be possible to draw comprehensive contour lines through scars of equal age along the long profile of the stream. The problem of interpretation is compounded further by the impossibility of differentiating those scars caused by the spring flood in the tributary, and those caused by ice blocks moving under the influence of Mackenzie backwater. The main positive result of the study is the definition of an upper limit to ice scarring (cf. ice-scar envelope, Fig. 78.3). The envelope is not definitive (higher backwater without ice scarring may occur) and no frequency of occurrence can be attached to it. The envelope is merely a line joining the maximum elevations of observed ice scars.

Acknowledgments

The authors would like to acknowledge the field assistance of M. Monteith.

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VEGETATION TYPES OF NORTH-CENTRAL DISTRICT OF KEEWATIN

Project 760058

S.A. Edlund
Terrain Sciences Division

The vegetation types of north-central Keewatin were studied in conjunction with the terrain mapping project of R.D. Thomas and A.S. Dyke (Project 760012) during the 1976 summer field season. Inventories of vascular and cryptogamic plant species at nearly 200 sites assisted in the delineation of vegetation types and plant communities. Detailed per cent cover measurements of various plant species were made at selected sites, and visual estimates were made of per cent plant cover at the majority of reconnaissance sites. Areas undergoing colonization, such as fluvial and eolian deposits, were studied in detail. Range extensions were recorded for several vascular plant species.

Preliminary correlations between the plant communities and the surface materials and climate are presented in Figure 79.1. More detailed information will be forthcoming.

Surface Materials

Three basic types of materials occur at the surface in the study area: 1) bedrock — Precambrian schists and gneisses, and localized quartzites, granites, and volcanics; bedrock outcrops are not common or extensive; 2) tills and ice-contact materials — bouldery, gravelly, and sandy materials derived from the schist and gneiss bedrock and formed into a variety of glacial landforms. Large angular boulders are commonly a major component of the tills; and 3) thick deposits of silts of marine origin — these deposits are located in the northern part of the study area around Chantry Inlet and Hayes and Murchison rivers.

Climate

Superimposed on this framework of materials are climatic zones. The study area lies within the broad "Low Arctic" climatic zone, indicated by an absence of trees, continuous vegetation cover, and a significant shrub component in the composition of the plant communities. A continental climate is indicated for the whole area, with the presence of Chantry Inlet yielding no discernable maritime influence on the vegetation.

Within this climatic zone, there are three basic subdivisions, reflected by vegetation zones: 1) Shrub tundra zone in which ericaceous and other shrubs are the dominant vascular plants and species diversity is the greatest; 2) Intermediate tundra zone in which shrubs dominate, though diversity and amount is reduced from zone 1. The cryptogamic component of the communities increases in importance; and 3) Lichen-shrub tundra zone, corresponding to Wager Plateau, an upland on which lichens have the greatest per cent plant cover and two species of ericaceous shrubs are the dominant vascular plants. Species diversity is lowest in this zone.

Vegetation Zones

Zone 1. Shrub Tundra Zone

This zone of mixed shrub tundra (zone 1) barely reaches into the study area and reflects a more southerly "Low Arctic" vegetation. It occurs only in the extreme

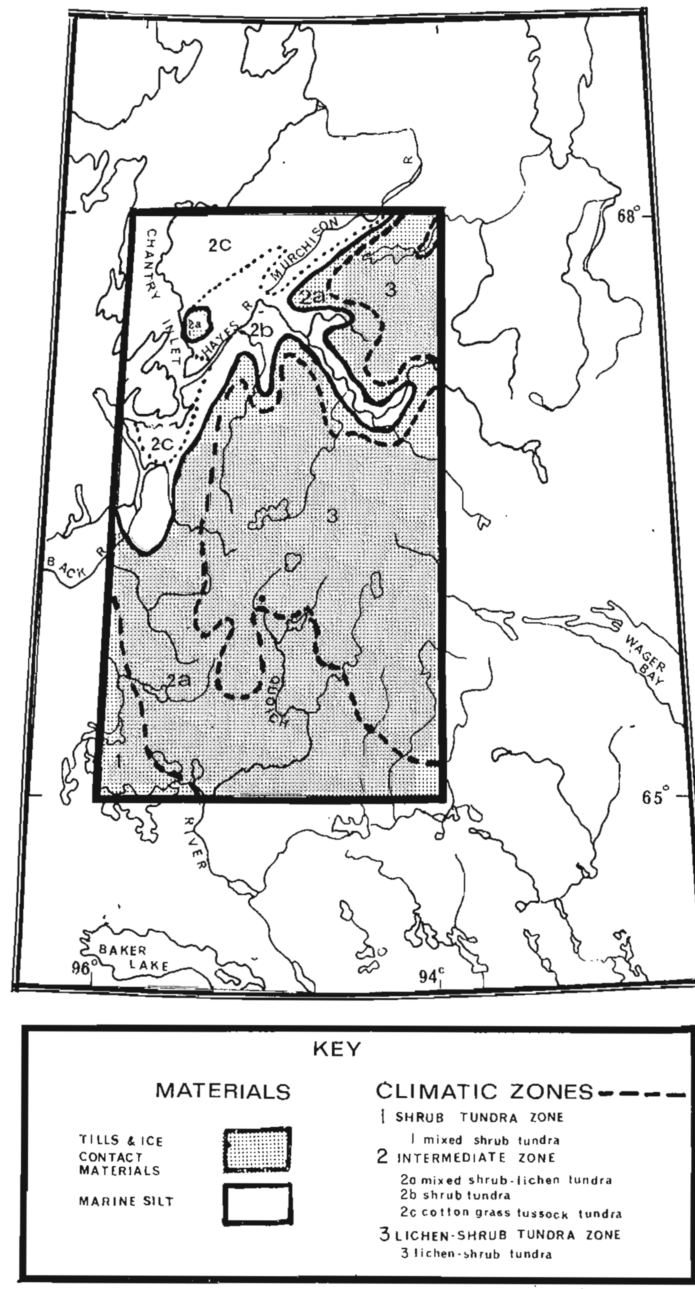

Figure 79.1. Vegetation types of north-central Keewatin.

From: Report of Activities, Part A;
Geol. Surv. Can., Paper 77-1A (1977).

southwestern area where tills and ice-contact materials are the dominant surface materials. Bedrock outcrop is minimal. Elevation of the area is less than 400 feet.

The shrub component of the communities is greater than 60 per cent, and shrub diversity is high. On moderately to well drained materials ericaceous shrubs are dominant. *Cassiope tetragona* (arctic white-heather), *Arctostaphylos alpina* (alpine bearberry), *Ledum decumbens* (Labrador tea), *Vaccinium vitis-idaea* (mountain cranberry), and *Vaccinium uliginosum* (bilberry) are the most common components. *Empetrum nigrum* (crowberry), *Dryas integrifolia* (mountain avens), *Salix arctica* (arctic willow), *Rhododendron lapponicum* (Lapland rosebay), and *Diapensia lapponica* are frequently associated with these communities. Continuous ground cover is provided by a thick lower stratum of lichens such as *Alectoria ochroleuca*, *A. nitidula*, and various *Cladonia* and *Cetraria* species.

Wetland vegetation types, on poorly drained materials in this region, make up a small portion of the total vegetation types. They consist of a variety of *Carex* sp. (sedges) and *Eriophorum* sp. (cotton grass) in meadow and bog communities. Grasses and assorted rushes are found as well. A dense lower stratum of mosses and liverworts is common. Where moss hummocks develop, a small ericaceous shrub component is found on the hummock tops and sides, including *Phyllodoce coerulea*, *Andromeda polifolia*, *Cassiope*, *Ledum*, and *Vaccinium*. Erect and trailing *Salix* sp. (willow) and *Betula glandulosa* (dwarf birch) also occur in some abundance in these wetlands.

Dense thickets of erect shrubs such as *Salix alaxensis*, *S. cordifolia*, *S. niphoclada*, and *Betula* locally occur along streams and in protected valleys.

Zone 2. Intermediate Tundra Zone

This is a transition zone between shrub-dominated communities and cryptogamic communities. The zone lies between 400 and 900 feet elevation in the south and from sea level to 600 feet in the north. Different communities have developed on the different materials. The shrub component of the communities is generally between 30 to 50 per cent. Monocots become more significant in the drier sites, while the shrub component in the wetlands is much reduced.

In the few small areas of bedrock outcrop, crustose lichens cover undisturbed, exposed facets of the rock. Foliose and fruiticose lichens occasionally occur as well. Vascular plants are found only in pockets where fines have accumulated.

Well to moderately drained tills and ice-contact materials within this zone support shrub-lichen communities (zone 2a). *Cassiope* and *Ledum* are the dominant vascular species. The nonericaceous component (*Betula* and *Salix*) is much reduced but is still present. Monocots such as *Carex Bigelowii*, *Luzula nivalis*, and *Hierochloe alpina* are common. An *Alectoria* lichen lower stratum has a greater than 60 per cent cover. Wetland communities are composed of a variety of monocots, carices, and *Eriophorum* sp. The shrub component of these communities is reduced in size and diversity from that of zone 1. There is a well developed lower stratum of moss.

The major silt deposits of the northern portion of the study area support a different type of plant community than those on the coarser grained materials. Along Hayes River silt bluffs are actively eroding. These exposed surfaces are devoid of vegetation. In seepage areas near the base of these cliffs, halophytic grasses such as *Elymus arenarius* ssp. *mollis* and *Puccinellia phryganodes* can be found.

Moderately to well drained stabilized silts have *Salix* - and *Dryas*-based communities (zone 2b). These shrubs are often located in a network of shallow polygon troughs. *Potentilla* sp. is a common shrub in some communities. A lower stratum can be composed of mosses, commonly *Rhacomitrium* sp., or may be absent. Erect shrubs occur in protected areas. Some communities have a substantial monocot and herb component, including carices, *Alopecurus alpinus*, *Juncus* sp., and saxifrages.

The lack of dense lichen mats on the silty materials is noteworthy. *Alectorias*, *Cetrarias*, and *Cladonias* are rarely found.

Ericaceous shrubs, *Betula glandulosa*, and *Salix alaxensis* are usually limited to areas of late-lying snowbeds. There they are mixed with other erect and prostrate dwarf shrubs.

Communities on extensive poorly drained silts (zone 2c) are dominanted by *Eriophorum vaginatum* tussock vegetation (50 to 80 per cent). *Salix arctica* and erect willows may be common components of these communities, as are carices, herbs, and mosses.

Zone 3. Lichen-Shrub Tundra Zone

Extensive lichen-shrub communities occur on the upland plateau at 600 to 900 feet and more. These communities are found exclusively on tills and ice-contact materials. Bedrock outcrop is more common, but is still local. The communities are characterized by a further decrease in per cent cover of shrubs and also in species diversity.

Precambrian bedrock, as in zone 2, is primarily covered with crustose, foliose, and fruticose lichens. Quartzite outcrops, however, have little if any lichen cover, or other vegetation.

The most common communities over this area are Heath-lichen tundra communities (zone 3), found on the bouldery sandy tills and ice-contact materials. The communities are composed of *Cassiope* and *Ledum* (10 to 30%) and *Alectoria* sp. (60 to 90%). *Hierochloe* (less than 10%) -*Alectoria* (90%) communities are found on the driest areas such as outwash terraces, eskers and, kame tops and upper slopes, and on gravel deposits. Moderately to poorly drained areas support monocot-*Cladonia*-moss communities. Wetlands occur occasionally, but are not common; sedges and grasses are common.

A RECONNAISSANCE SURVEY OF SEVERAL PLUTONIC IGNEOUS BODIES WITHIN ONTARIO

Project 740057

P.A. Brown and J.J.B. Dugal
Terrain Sciences Division

Introduction

This study, which forms part of the hard rock program for the geological disposal of radioactive wastes, involved a reconnaissance survey during the summer of 1976 of nine plutonic igneous bodies and one area of gneiss. All the bodies are within Ontario and are located from Kenora in the west to Madoc in the east (Fig. 80.1). A variety of tectonic settings, i.e. pre-tectonic to post-tectonic, and depths of intrusion, i.e. catazonal to epizonal, were encountered. The survey encompassed rock types ranging in composition from granite to anorthosite.

1. Fiest Lake Pluton

The Fiest Lake pluton (49°49'N, 93°51'W) is a foliated granodiorite which intrudes Archean mafic and felsic metavolcanics (Pryslak, 1976). A rapid survey of the northern margin of the pluton indicates that it is a pre-tectonic to syntectonic, biotite granite-granodiorite with numerous xenoliths of an earlier granodiorite-diorite phase. Both rock types have been deformed at least once and a weak tectonic fabric, defined by biotite laths, is developed. This fabric is locally crenulated about a northeast-southwest axis, and minor zones of intense deformation are developed parallel to this axis. A later, variably deformed, biotite quartz monzonite intrudes the granite-granodiorite and surrounding volcanics and obscures the contact relationships.

The fracture system within the batholith is well developed. Four or five fracture sets occur at most outcrops; these have steep dips and strike north-south, northeast-southwest, east-west, east-southeast – west-northwest, and south-southeast—north-northwest. They are straight, 1 to 5 m in length, and have a frequency range of 30 cm to 2 m. Many contain a chlorite-epidote gouge, and slickensides are common. Sheet joints are variably developed throughout the area.

Figure 80.1. Location of study sites within Ontario; numbers refer to sites discussed in the text.

2. English River Gneiss

The English River Gneiss in the Willard Lake area (49°56'N, 93°52'W) is a well banded quartzofeldspathic gneiss. The banding, which strikes north-northeast to northeast, is defined by biotite- and hornblende-rich layers and by amphibolite bands up to 10 cm wide. The gneiss is intruded by granite, which forms pods, and by aplite dykes. A further deformation results in isoclinal folding of the gneissic banding. This is followed by the intrusion of granite and pegmatite stringers and dykes and the development of a weak axial planar fabric which strikes northeast to east-northeast and dips at a high angle to the southeast. Minor shearing is associated with the fabric.

The fracture system within the gneiss is very poorly developed. Two fracture sets occur: one strikes northeast-southwest and parallels the gneissic banding and/or the later fabric, and the other strikes at a high angle to the banding, i.e. northwest-southeast. Minor local fractures strike north-south and east-west. The fractures have a frequency range of 50 cm to 5 m, are tight, straight, and smooth, and locally show minor displacements.

In contrast to the poor development of fractures within the gneiss, the granite pods are intensely fractured. The major fracture sets strike north-south, northeast-southwest, and northwest-southeast, have steep dips, and are generally straight and smooth. The frequency varies from 10 cm to 1.5m, and the length ranges from 2 to 4 m. In addition to these sets the granite contains pervasively developed minor fractures, which parallel the major sets but have a length of 5 to 25 cm and a frequency range of 2 to 10 cm.

From: Report of Activities, Part A;
Geol. Surv. Can., Paper 77-1A (1977).

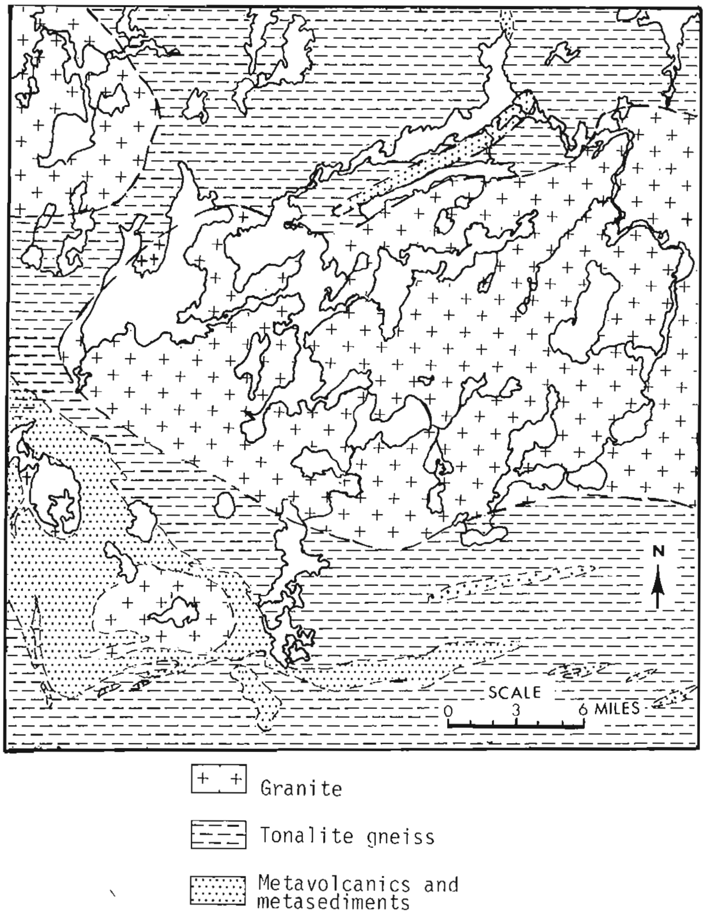

Figure 80.2. Cecil Lake Granite.

3. Cecil Lake Granite

The Cecil Lake Granite (49°30'N, 91°30'W) is a post-tectonic, medium- to coarse-grained biotite granite of batholithic dimensions which intrudes basement gneisses and metavolcanics of the Wabigoon volcanic-plutonic belt (Sage *et al.*, 1974) (Fig. 80.2). The basement has a complex tectonic-plutonic history. To the north of the granite the gneisses are tonalitic to trondjemitic in composition and have a well developed tectonic banding defined by plagioclase-rich and plagioclase-biotite-rich layers. This is intruded by a hornblende diorite-biotite granodiorite phase which contains xenoliths of the banded gneiss and deformed and metamorphosed basic volcanic rocks. Fine grained granodiorite dykes cross-cut this sequence. Subsequent to the intrusion of the dykes the entire sequence was further deformed and a tectonic fabric developed.

The gneisses at the southern contact of the granite, although having a tectonic-plutonic history similar to those to the north, have a different character in that the banding is defined, in part, by amphibolites. The amphibolites are black, medium grained hornblende-plagioclase ± quartz rocks and appear to have been formed much earlier within the tectonic history of the complex than the slivers of deformed and metamorphosed basic volcanic rocks which are infolded(?) into the gneiss and which also occur as xenoliths within the diorite-granodiorite phase.

The Cecil Lake Granite post-tectonically intrudes this gneiss complex. Close to the contact undeformed stringers and dykes of granite and pegmatite cross-cut the gneisses. The contact itself is diffuse in that there is a gradation from gneiss with granite, to granite with gneiss xenoliths, and finally to homogeneous medium to coarse grained granite. Recent gravity studies (Szewczyk and West, 1976) indicate that the granite has a sheet-like form with a modelled depth of 2.1 km.

The main granite body consists essentially of a biotite granite, although local muscovite-biotite phases are present. Xenoliths of banded tonalite gneiss are scattered throughout the body. Generally these are partially assimilated and, where almost completely resorbed, impart a ghost structure to the granite. Dykes and irregular pods and lenses of pegmatite, although present throughout the body, are most common close to the margins. The potassium feldspar within the pegmatites is generally perthitic and locally shows graphic intergrowth with quartz.

Although essentially undeformed, the granite locally contains minor zones of shearing. These are less than 25 cm wide, strike northeast, and dip at shallow to moderate angles to the northwest. The development of fractures is highly variable. The central part of the granite is poorly fractured; two, and locally three, sets are developed. These have steep dips and strike north-south, east-west, and locally either northeast-southwest or northwest-southeast. They have frequencies of the order of 1 to 4 m (locally up to 10 m) and lengths of 3 to 25 m. The fractures are straight with curved splays, are tight, smooth to rough, and do not contain any gouge or filling material. Narrow zones of multiple fracturing, having a frequency of 2 to 4 mm, are locally developed. No slickensides were noted.

The marginal zone of the granite is more highly fractured with three or four and locally five fracture sets at most outcrops. The main sets strike northwest-southeast, northeast-southwest, and east-west. They have steep dips, a frequency range of 25 cm to 2 m, lengths of 1 to 3 m, are tight, straight, smooth to rough, and are only locally filled with chlorite and epidote. Sheet joints are developed throughout the area.

4. Marathon

The area mapped around Marathon (48°48'N, 86°25'W) comprises the eastern part of an arcuate to rounded alkalic complex, the Port Caldwell Syenite (Fig. 80.3). It is a coarsely crystalline-layered igneous body which post-tectonically intrudes gneisses and metavolcanics of Archean age (Milne, 1967).

The eastern margin is defined, on the coast, by syenite and in an inland section by a massive olivine gabbro. The relationship between these units was not observed. The olivine gabbro is overlain to the west by a banded gabbro defined by mafic (hornblende and magnetite) layers 5 mm to 3 cm wide and felsic layers (plagioclase) up to 10 cm wide. The sucession appears to grade (mineralogically) upwards to the west, i.e. hornblende-magnetite-rich bases and plagioclase-rich tops. This in turn is overlain by a well developed banded syenite, which is at a scale of 20 to 50 cm; cross-bedding structures are observed locally. These indicate that the banding youngs to the west and has not been overturned. The banded syenite is overlain by a massive, coarse to medium grained syenite. This unit comprises the bulk of the complex and is somewhat homogeneous except for local variations in grain size and mafic content. Where bedding features were observed, these confirm a younging direction to the west and southwest. In the west-central part of the syenite there are large rafts of metavolcanics that may represent roof pendants and, as such, may indicate that the body has not been deeply eroded.

The syenite is cross-cut by diabase dykes which are up to 10 m wide and strike in a north to northwesterly direction. These are cross-cut by syenite pegmatite dykes. The final magmatic event was the emplacement of fine grained syenite dykes.

The fractures developed within the syenite fall into two groups: (a) those which are related to either the emplacement of the body or its subsequent uplift, or both, and (b) those which are related to an episode of faulting either during or after uplift.

The first group comprises three fracture sets, which strike northwest-southeast and east-west, and which have a variable but generally steep dip. Although irregularly developed throughout the area, they generally have a frequency range of 50 cm to 3 m, lengths of 2 to 20 m, are straight, tight, and smooth, and seldom contain any gouge or filling material. Locally four, five, and six fracture sets were observed, but the additional sets tend to be inconsistently and poorly developed. Sheet joints occur sporadically throughout the area.

The second group has a restricted distribution and tends to occur in zones. These zones show eivdence of breakdown and are both steeply dipping and flat lying. They are characterized by the breakdown of the feldspars, chloritization of the mafic minerals, and pervasive fracturing with a frequency of 2.5 mm. These fractures occur in 5- to 25-cm-wide zones, which are spaced 3 to 8 m apart, and are developed over distances of the order of 70 to 100 m. The fractures generally are filled with calcite and hematite, and slickensides are common.

5. Shawmere Anorthosite

A section from the eastern margin to the centre of the Shawmere Anorthosite (48°15'N, 82°40'W) confirmed that the body (at least along the section) is a deformed, medium to coarse grained plagioclase ± orthopyroxene ± clinopyroxene ± garnet hornblende rock (Thurston *et al.*, 1974). The central section is coarse grained and little deformed and consists essentially of plagioclase with interstitial mafic clots. These clots have an orthopyroxene core surrounded by hornblende and an outer rim of garnet. Towards the eastern margin the anorthosite becomes progressively deformed. This results in a granulation of the plagioclase with concomitant reduction in grain size and a streaking out of the mafic clots which give the rock a banded aspect. The banding is tightly folded, and a northeast-striking axial planar fabric is developed.

The final tectonic (?) event is the breakdown of the anorthosite along 5-mm-to 5-cm-wide zones. These zones strike northeast to north-northeast, dip at moderate to steep angles to the northwest, and have a frequency range of 25 cm to 1 m. They are peculiar in that although there has been complete breakdown within the zones,

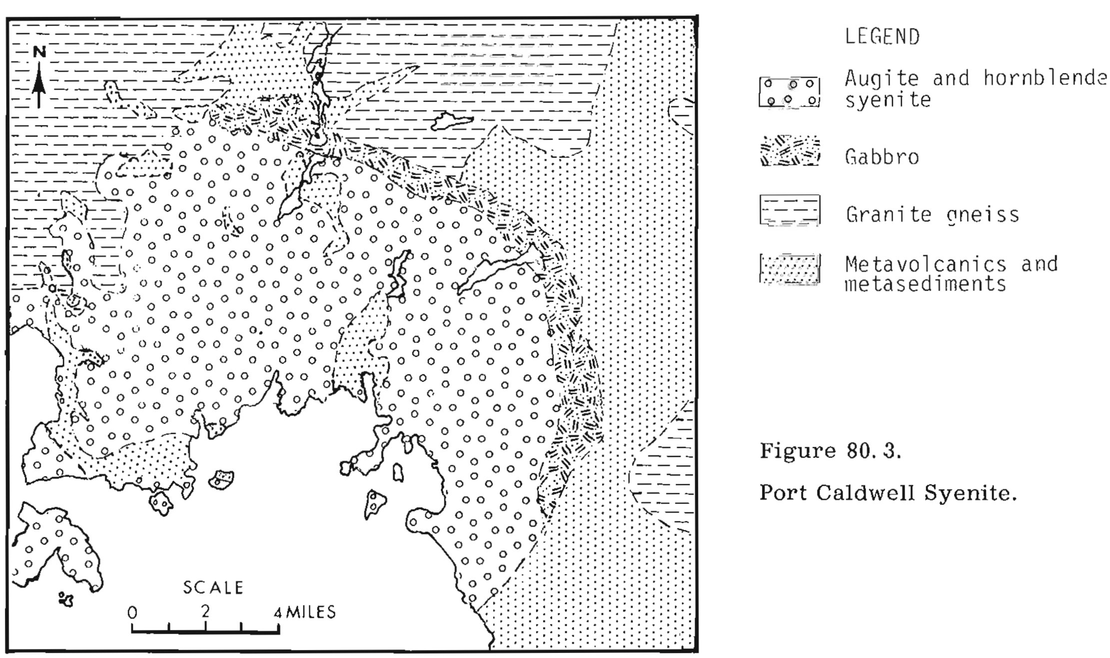

Figure 80.3.

Port Caldwell Syenite.

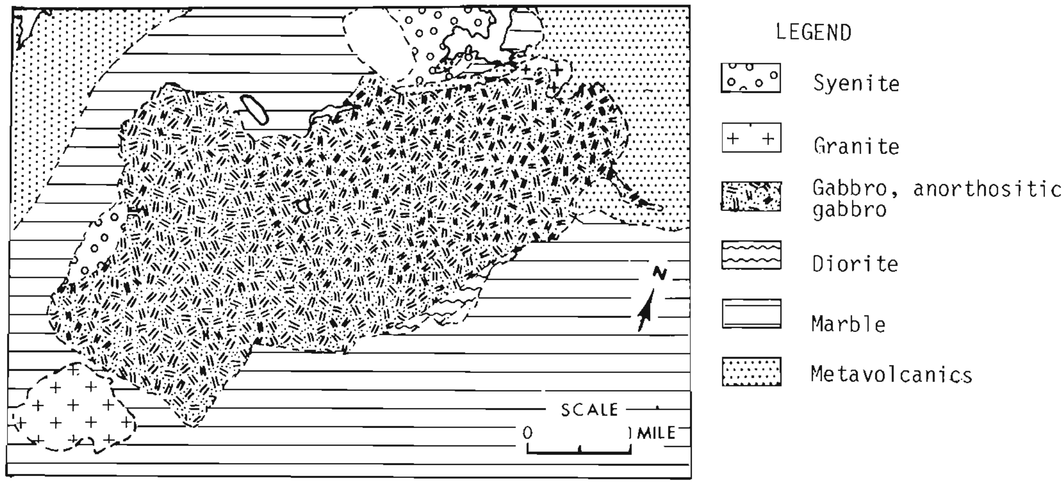

Figure 80. 4. Tudor Gabbro.

there is no displacement across them. This may indicate that they are a product of pure flattening and, as such, may be related to the deformational event which tightly folded the banding.

Fractures are variably developed throughout the body. In general three sets are present which have moderate to steep dips and strike north-northwest—south-southeast, west-northwest—east-southeast, and east-northeast—west-southwest. Locally they have frequencies of the order of 15 to 25 cm, but generally the frequency is about 1 m. They are straight, smooth, tight, and 2 to 5 m in length.

6. Tudor Gabbro

The Tudor Gabbro (44°43'N, 77°40'W) is a layered gabbro-anorthosite complex which syntectonically to post-tectonically intrudes marbles and mafic metavolcanic rocks of the Grenville Supergroup (Fig. 80.4). It is elliptical in plan and elongated in a northeasterly direction. The body consists essentially of medium- to coarse-grained gabbro and uralitized gabbro with anorthositic phases and a marginal diorite phase (Lumbers, 1969). The banding, which is locally well developed, is defined by sharply bounded mafic-rich and felsic-rich layers varying from 5 to 25 cm in width. Magnetite-ilmenite-rich horizons define the banding in the north-central part of the intrusion. Clots of mafics, rather than bands, were observed locally. Throughout the body the banding is irregular and discontinuous. The dominant strike varies from northeast to southeast, with moderate to steep dips both to the north and south. Way up criteria, found at only one locality, indicate that the banding has not been overturned.

Dykes of fine grained diabase and coarse pegmatite cross-cut the banding; the pegmatites also occur as discontinuous pods and lenses within the banding. These dykes were not observed to cross-cut the rafts of schistose metasediments, metavolcanics, and marbles which are locally found within the body.

The complex is altered extensively as evidenced by the uralitized pyroxenes and saussuritized feldspars. Shear zones are irregularly developed throughout the body and strike in northwest to northeast and east-west directions. Within these zones, which may be up to 5 m wide, the gabbro is further broken down to a chlorite-sericite schist.

The fracture system within the complex appears to be related to, or has been extensively modified by, the development of the shear zones. In general three or four, but locally as many as six or seven, fracture sets are present at any one outcrop. Close to the shear zones the fractures have frequencies as low as 5 to 10 cm, generally show slickensides, are filled with calcite ± chlorite, and are curved rather than straight. Away from the shear zones the fractures have a frequency range of 5 to 75 cm, are straight, tight, and smooth, and only locally are filled with calcite. Fracture length varies from less than 1 m to greater than 20 m.

7. Lingham Lake Complex

The Lingham Lake Complex (44°45'N, 77°30'W) is a composite gabbro-granodiorite body which syntectonically to post-tectonically intrudes basic metavolcanic rocks of the Grenville Supergroup (Fig. 80.5). Compositionally the complex varies from pyroxenite through gabbro and anorthositic gabbro to diorite and granodiorite (Lumbers, 1969). The pyroxenite is a minor constituent and occurs close to and along the western boundary of the body. The gabbro-anorthositic gabbro varies from coarse- to fine-grained and also occurs as a marginal phase. A fabric defined by oriented plagioclase laths is variably developed throughout this unit; it strikes north-south and dips steeply to both the east and the west, although predominantly to the east. Locally a crude igneous layering parallels the fabric.

The central and eastern parts of the body consist essentially of medium grained diorite, quartz diorite, and granodiorite with either hornblende or biotite as the mafic phase. The diorite is locally intrusive into the gabbro, and an intrusive breccia is developed. A tectonic fabric is well developed throughout the eastern part of the complex. Fine grained mafic xenoliths, contained within the granodiorite, are flattened within the plane of the fabric, and narrow zones of intense deformation, resulting in comminution, are locally developed. Diabase, aplite, and granite pegmatite dykes, which intrude the granodiorite, also contain this fabric. Surrounding metasedimentary and metavolcanic rocks contain a fabric which is subparallel to the fabric within the intrusive complex.

Four fracture sets occur at any one outcrop and strike northwest, north, northeast, and east, and have moderate to steep dips. The fractures have a frequency range of 5 cm to 1 m, are straight, tight, and smooth, and vary from 1 to 5 m in length. Calcite and hematite locally occur as filling material; slickensides are rare.

Northeast-striking local shear zones breakdown the gabbro to a chlorite-sericite schist. Due to the poor outcrop in low-lying areas, the nature and extent of these zones could not be assessed.

8. Weslemkoon Granite

The Weslemkoon Granite-Granodiorite (45°00'N, 77°25'W) is a deformed, in part gneissic, body of batholithic dimensions (Evans, 1964; Lumbers, 1968). Sections from the margin to the centre of the body show that composition varies with poorly deformed biotite granite at the margins and a biotite granodiorite inwards. The latter is locally intensely deformed, and a fabric defined by biotite laths is developed. This is succeeded by a biotite-hornblende granodiorite-tonalite which locally is cut by a mafic-poor granite. A compositional banding, defined by biotite-rich and biotite-poor layers and also by amphibolites, imparts a gneissic aspect to parts of the batholith.

Fractures are variably, but generally well developed throughout the body; three and locally four or five sets were observed at most outcrops. These strike north-northwest—south-southeast, northeast-southwest, east-west, and north-south, west-northwest—east-southeast, and have moderate to steep dips. They have a frequency range of 20 cm to 3 m, although generally the spacing is of the order of 1 m. The length varies from 1 to 10 m; they are straight, tight and smooth, and seldom contain any filling material. Sheet joints are developed locally throughout the area.

9. Elsevier Granite

The Elsevier Granite (44°40'N, 77°20'W) is a composite granite-granodiorite-trondjemite body which intrudes mafic volcanics of the Grenville Supergroup (Meen, 1944; Hewitt, 1964). A section through the body indicates that it consists essentially of medium grained equigranular granodiorite-trondjemite. A tectonic fabric, defined by biotite, is developed throughout the body. Granite to granodioritic compositions are prevalent towards the margins.

Fractures are variably developed within the body, and generally four sets were observed at any one outcrop. These strike northwest-southeast, north-south, northeast-southwest, and east-west, and have moderate to steep dips. Although the frequency range is 5 cm to 3 m, the general spacing is 75 cm to 1 m. They are straight, tight, and smooth and vary from 2 to 20 m in length. Quartz and epidote locally occur as a filling material within the fractures; slickensides are not uncommon.

10. Mount Moriah Syenite

The Mount Moriah Syenite (44°42'N, 77°25'W) is a subcircular, coarse grained syenite complex, which post-tectonically intrudes metavolcanics of the Grenville Subgroup (Meen, 1944) and is unconformably overlain by Ordovician (Black River and Trenton groups) limestone (Fig. 80.6). The major part of the body comprises a medium to coarse grained equigranular augite and/or hornblende syenite with local biotite-rich phases. The mafic content is generally of the order of 15 per cent and rarely exceeds 20 per cent.

In the extreme southwest of the body an arcuate mass of medium- to coarse-grained, mafic-poor syenite

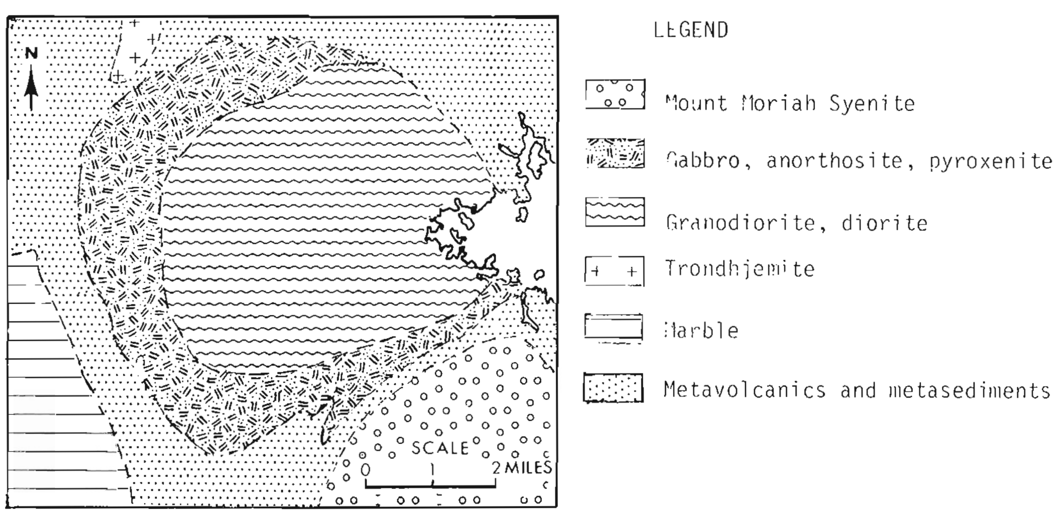

Figure 80.5. Lingham Lake Complex.

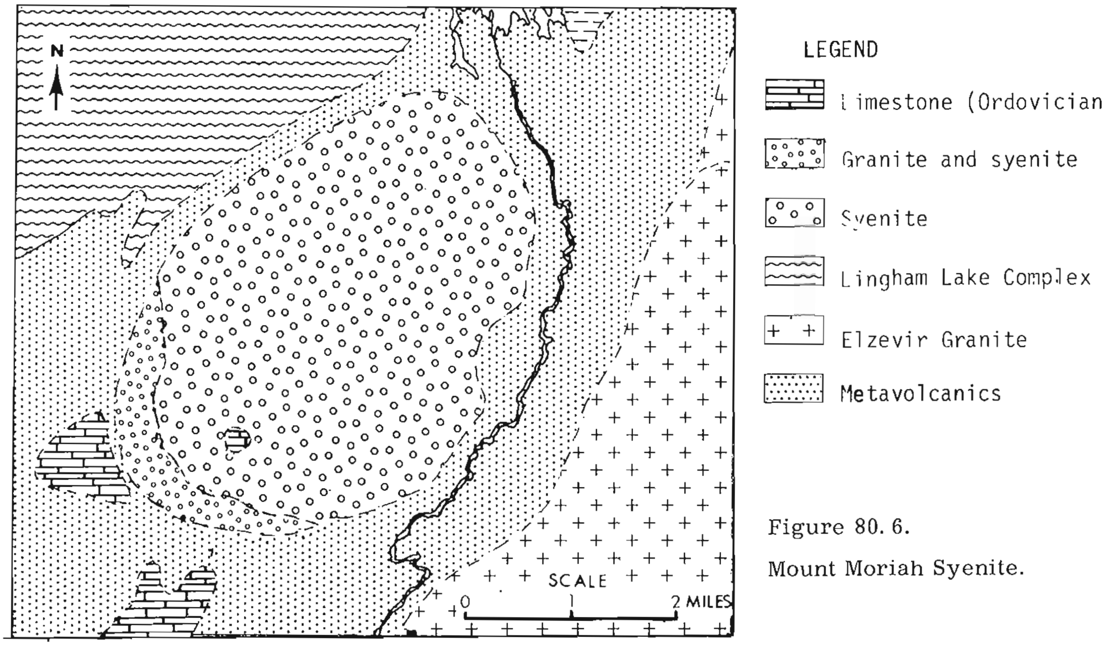

Figure 80.6.
Mount Moriah Syenite.

is separated from the main part of the body by a sliver of amphibolite. This part of the syenite contains a minor quartz monzonite to granite phase which occurs both as discontinuous and irregular pods and as dykes. Dykes of this composition also cut the main body of the syenite and locally have an arcuate outline. Porphyritic to fine grained syenite dykes, which are cross-cut by the granitic dykes, generally trend at high angles to the margin of the syenite and define a crude radial pattern.

A weak igneous fabric, defined by oriented potassium feldspar crystals, is locally developed in both parts of the complex. The fabric parallels the margins and is steeply dipping both outwards and towards the centre of the body. No igneous layering was observed.

Four fracture sets were observed at any one outcrop within the syenite; three dip at moderate to high angles, and the fourth is generally a sheet joint. Locally two or three additional fracture sets are present, but they tend to be poorly and irregularly developed. The main sets show a radial pattern rather than having a consistent orientation throughout the body. They have a frequency range of 50 cm to 2 m, lengths of 2 to 20 m, are straight, smooth, tight, and seldom contain any gouge or filling material; slickensides are rare.

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Project 740057

J.E. Gale, J.J.B. Dugal, J.S.O. Lau, and K.G. Raven
Terrain Sciences Division

Introduction

Atomic Energy of Canada (AECL) and the Geological Survey of Canada (GSC) have undertaken a joint project (hard rock) to determine if igneous (intrusive) or metamorphic rocks are suitable host rocks for the storage of solidified radioactive wastes in a mined repository. This project was initiated at the beginning of the 1975-76 fiscal year. The main objective of this hard rock project is to select and evaluate a number of sites within the Ontario portion of the Canadian Shield (Gale *et al.*, 1976) that appear to meet the requirements of a radioactive waste repository. This paper is a brief review of current activities within the project and some discussion of the direction of future activities.

Current activities

Figure 81.1 is a flow chart for current project activities covering fiscal years 1975-76 and 1976-77.

Site selection

Emphasis in the early part of this project was placed on the selection of sites within Ontario by analysis of aerial photographs and the geological literature. Ten sites were selected and field checked during the 1976 field season (Brown and Dugal, 1977). Five of these areas (Ignace area, 49°22'N, 91°45'W; Marathon area, 48°47'N, 86°25'W; and three areas near Madoc, Ontario: Tudor Gabbro, 44°45'N, 77°39'W, Lingham Lake Gabbro, 44°46'N, 77°29'W, and the Mount Moriah Syenite, 44°44'N, 77°26'W were examined briefly (two to three days of field checking at each site) to assess their geotechnical-hydrogeological suitability.

This field checking revealed the following:

(1) The Ignace area is unsuitable because of the extensive cover of glacial material, resulting in approximately five per cent bedrock exposure (Sage *et al.*, 1974) in the five- to six-square-mile area of interest. The cover makes it very difficult to assess subsurface hydraulic and mechanical properties from surface measurements. Without good bedrock exposure initial geotechnical surveys in this area would be expensive, relying heavily on geophysical surveys and requiring extensive drilling to check geophysical anomalies. The exposures that were observed showed that the bedrock is massive with fracture (joint) spacings of up to 30 m in the centre of the batholith, but much closer fracture spacings were observed around the margins.

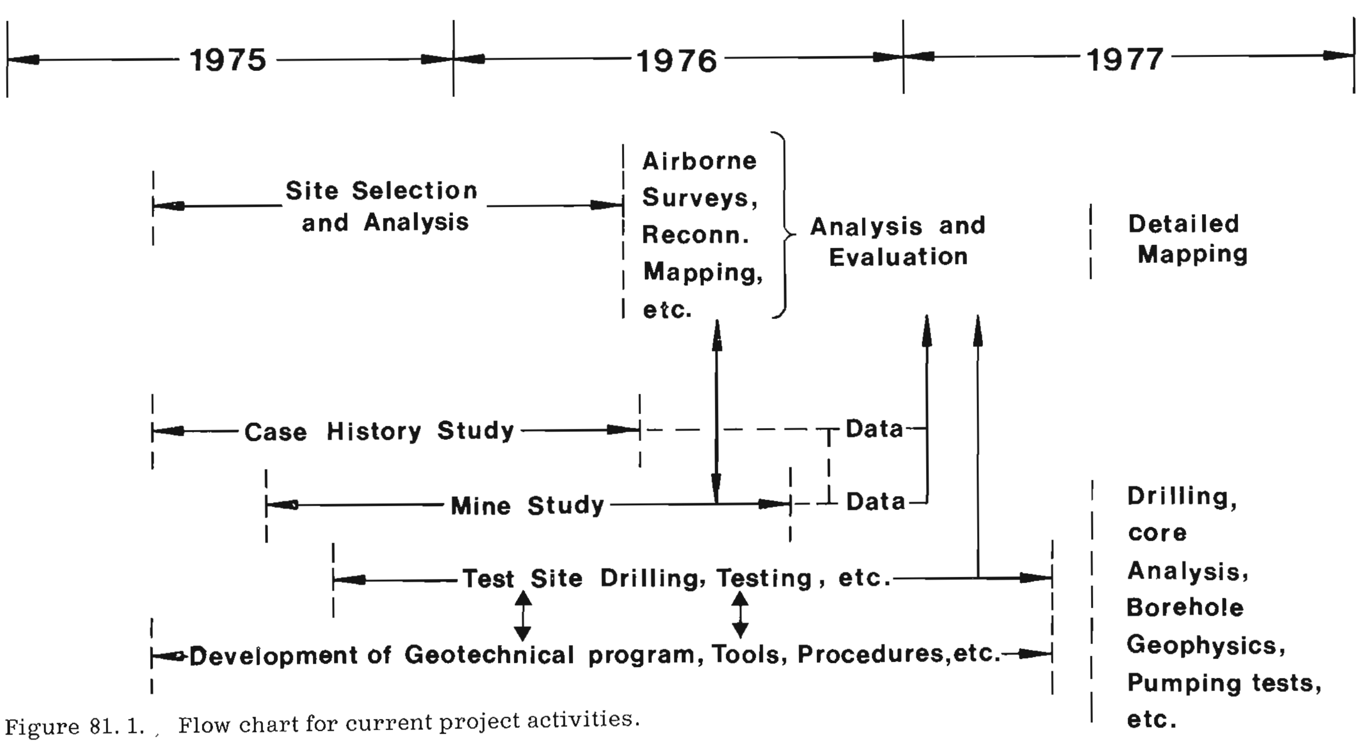

Figure 81.1. Flow chart for current project activities.

From: Report of Activities, Part A;
Geol. Surv. Can., Paper 77-1A (1977).

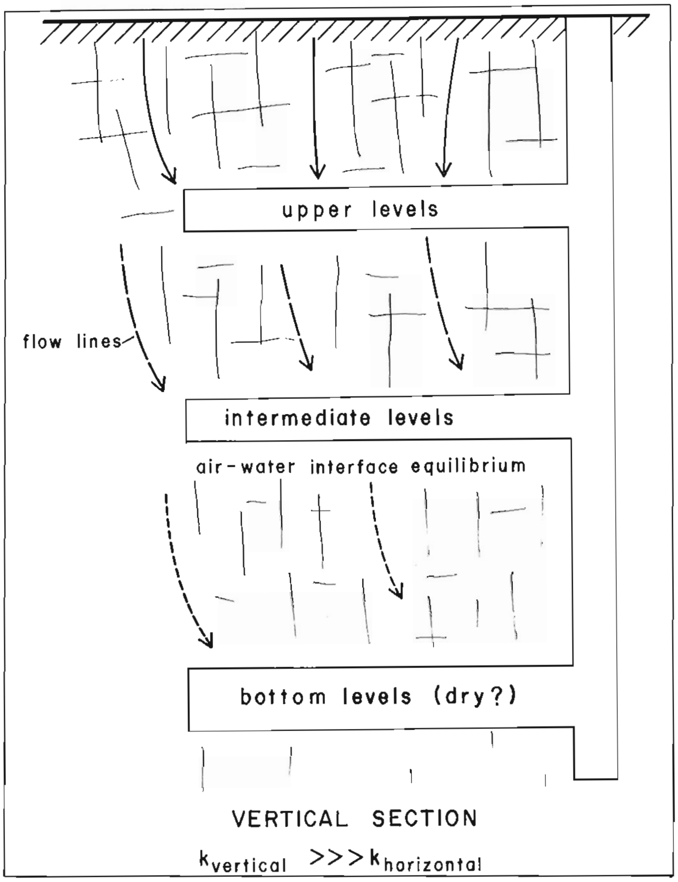

Figure 81. 2. Schematic drawing showing the upper levels of a mine intercepting the recharge to the rock mass.

(2) The Marathon area was mapped by Milne (1967). This mapping shows major east-west and northwest-southeast trending faults with definite offsets. In addition, both topographic maps and aerial photographs show major linears following similar trends in the area of interest. Bedrock exposure in the area is good (greater than 25 per cent in some places) with high topographic relief (from 160 to 440 m above mean sea level). This combination of high relief and a well developed system of major linears and faults, which show a high degree of interconnection, probably would result in the development of deep circulating groundwater flow systems. This would be an unfavourable hydrogeological environment in terms of waste isolation in the mined cavern concept. Also, the complex mineralogy and very coarse (up to 7-8 cm) grain size of the feldspars present in the syenite probably would exhibit unfavourable responses to thermal stress loads and strength characteristics which were considerably lower than those determined for finer grained and more isotropic syenites.

(3) The Tudor Gabbro (Lumbers, 1969) shows a well developed system of linears on aerial photographs and good bedrock exposure. The possible hydraulic and mechanical significance of the linears is not immediately apparent from aerial photographs. Field checking revealed the presence of several 3- to 6-m-wide zones

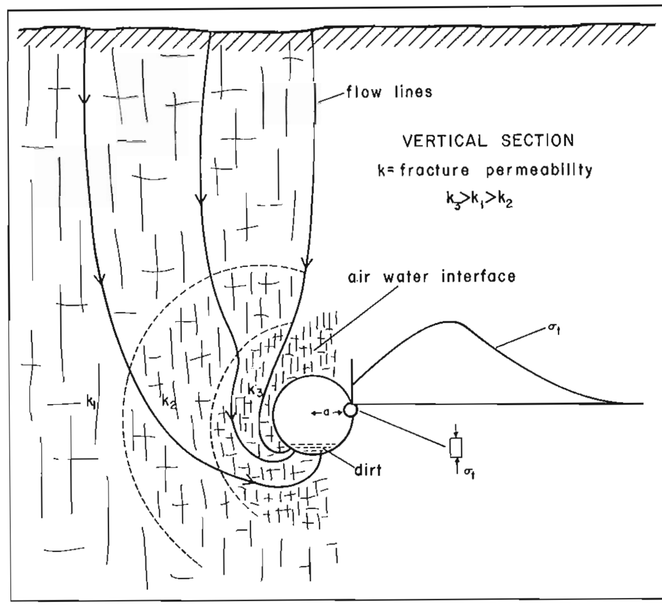

Figure 81. 3. Schematic drawing showing the interaction of stress and relaxation effects and its effects on the flow system.

of brittle deformation along the northeastern edge of the gabbro. More detailed field checking (Brown and Dugal, pers. comm., 1976) has shown that these zones are widespread throughout the body. The presence of these zones and their degree of interconnection and strength suggest that they will persist with depth and thus will present an unfavourable set of hydraulic and mechanical conditions for the construction of a radioactive waste repository.

(4) Aerial photographs and topographic maps show that the area underlain by the Lingham Lake Gabbro (Lumbers, 1969; Hewitt, 1964) is covered extensively by swamps and surface water. A one-day field check indicated that the linears present on aerial photographs are related to zones of brittle deformation. Because of the ground surface conditions in this area and the possible presence of many zones of deformation, geotechnical evaluation of this area would be very expensive and many areas of uncertainty would remain even after detailed work.

(5) The Mount Moriah Syenite (Meen, 1944; Hewitt, 1964) has not been assessed in any great detail at least with respect to the hydraulic and mechanical significance of the linears visible on aerial photographs. The two brief one-day visits revealed that this area is well exposed, fractures are widely spaced, and at least two of the major linears present in this body may be dykes or lenses of the country rock instead of zones of brittle deformation. The topography is low to moderate (from 259 to 370 m above mean sea level), and the major linears do not appear to be highly interconnected. Because of the above site characteristics, it should be possible to determine the hydraulic and mechanical properties of the major features present in the Mount Moriah Syenite to a high degree of certainty and with a reasonable expenditure of resources. Additional surface assessment work is planned during November 1976.

Case History and Mine Study

It is obvious from a review of the site selection work to date that in order to be able to assess quickly the relative merits of potential repository sites in other parts of Canada a suitable geomechanical model must be developed relating surface measurements to subsurface hydraulic and mechanical characteristics. The limiting factor in almost all previous attempts, in addition to the lack of a suitable model, has been a lack of surface and subsurface observational data as well as a suitable theoretical framework for the interpretation and correlation of such data. The current study (Fig. 81. 1) of large underground excavations (case history study) and underground mines (mine study) should provide some valuable observational data to assist with the development of such a model.

The preliminary results from the case history and mine studies are given by Raven (1977). The initial results suggest that, although there is an apparent correlation between structural features and groundwater inflows, many other factors contribute to the presence or absence of water in underground openings. As shown in Figure 81. 2 the upper levels of a mine may intercept the recharge to the rock mass and hence the lower levels of the mine may be able to dewater the surrounding rock mass quickly. In addition, evidence from studies of groundwater in crystalline rocks have shown that there is a logarithmic variation of permeability with depth in fractured rock masses. This variation of permeability with depth, when superimposed on the interaction of the stress and relaxation effects around mine openings (Fig. 81. 3) may result in the development of a free-surface condition, and the resulting reduced inflow can be removed by the mine ventilation system and hence no continuous seepage through the joint system was observed. The significance of the factors shown in Figures 81. 2 and 81. 3 in controlling groundwater flow into underground openings will be evaluated using finite element codes of groundwater flow.

White Lake Test Site

The White Lake experimental test site (Gale *et al.*, 1976) is located on the southeast side of the White Lake batholith (Quinn, 1952). Figure 81. 4 shows the fracture

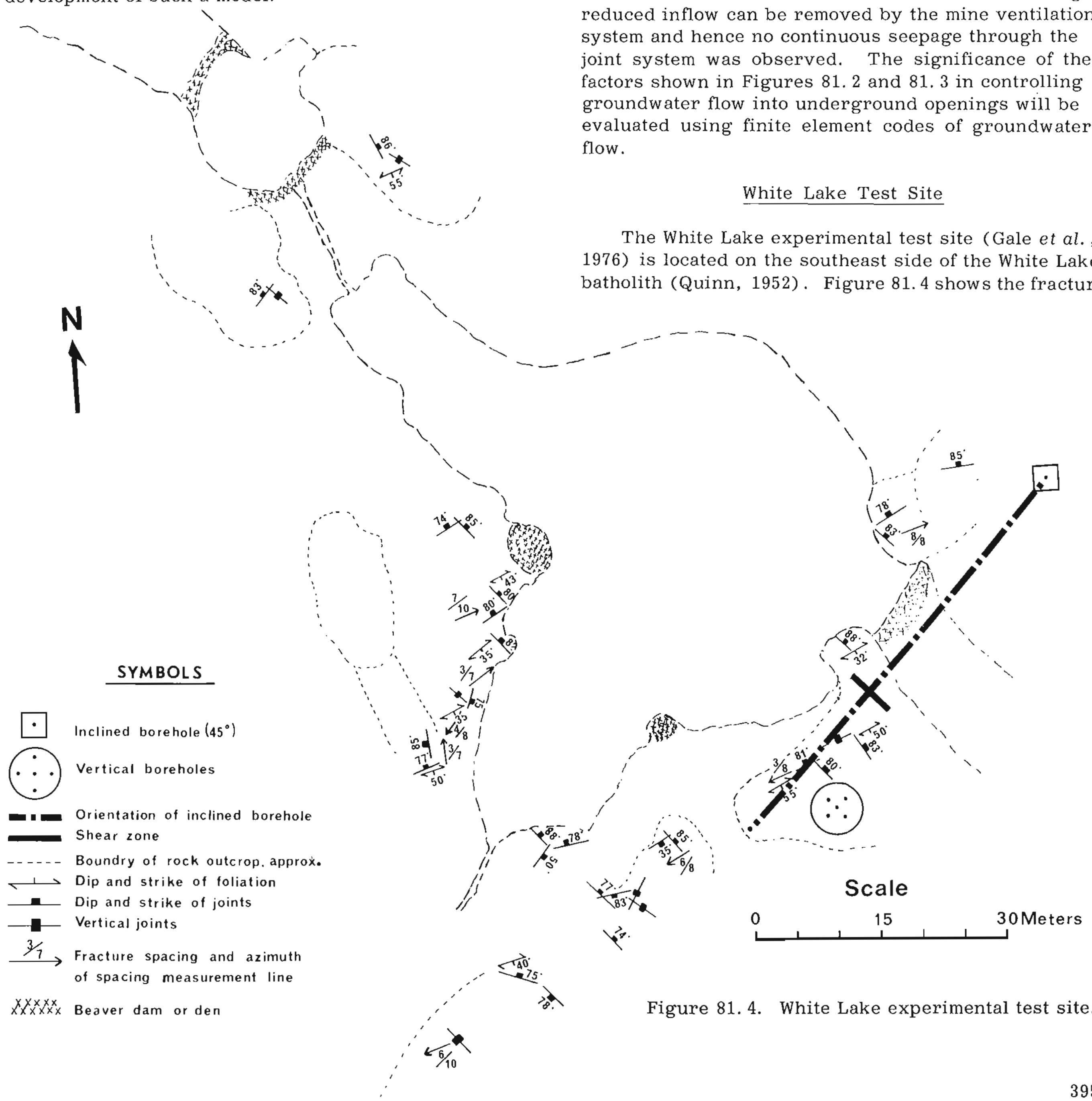

Figure 81. 4. White Lake experimental test site.

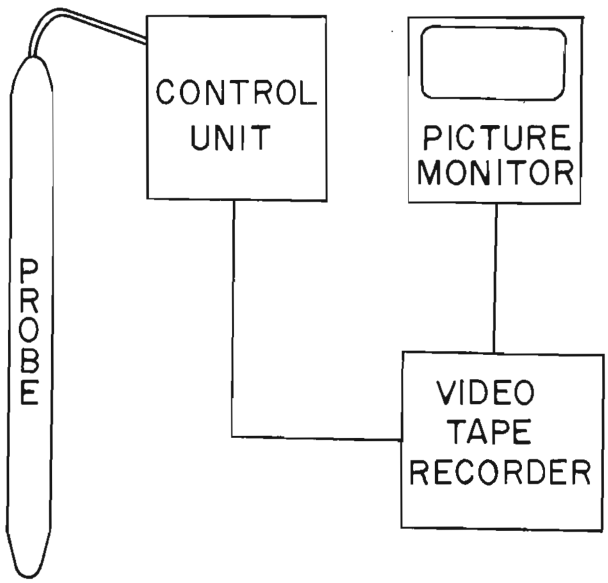

Figure 81. 5. The borehole TV camera operational system.

orientations in the immediate area of the test site and the location and orientation of the five, NX, vertical boreholes (centre borehole ≈37 m deep, four boreholes on the circumference of a 3 m circle, ≈30 m in depth) and the inclined (45°) corehole, 79. 5 m in length.

The logging of the drill core from the six boreholes and the borehole TV camera surveys of the five vertical boreholes and the first 57 m of the inclined borehole have been completed. In addition, borehole geophysical logging (caliper, density, gamma-neutron, E-log) of the six boreholes has been carried out by Roke Oil Enterprises Limited of Calgary. *In situ* hydraulic testing of the fractures intersecting the boreholes will be carried out during October and November 1976. Additional detailed surface and borehole geophysical surveys will be carried out at the White Lake test site by other groups within the Geological Survey of Canada.

The boreholes at White Lake were drilled by the Department of Public Works using standard diamond-drilling equipment. It is expected that future drilling programs that are a part of this radioactive waste containment program will utilize more advance diamond-drilling equipment; this will enhance drill core recovery

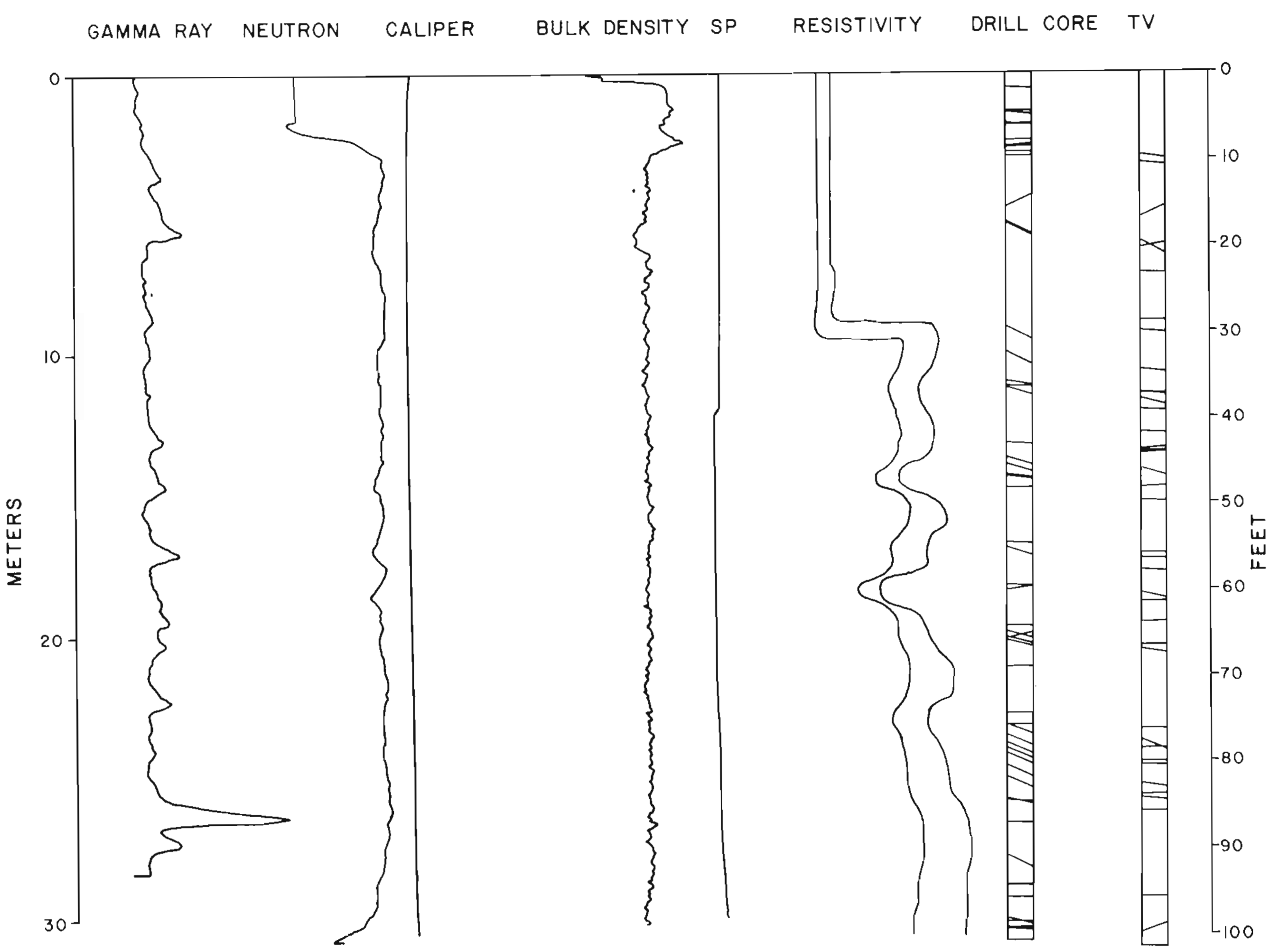

Figure 81. 6. A comparison of geophysical logs with the drill core log and the TV borehole log.

and core reconstruction, especially in zones of crushed or broken rock.

At the White Lake site an attempt was made to reconstruct the core using the well-developed gneissosity that is present in the core as a guide. Using this reconstructed core, the relative orientation of the fractures and breaks in the drill core were measured. These relative orientations then were converted to true strikes and dips using spherical trigonometric relationships (Lau and Gale, 1976).

The borehole TV camera surveys (Fig. 81.5) provide a direct measure of the strike and dip plus the location of the fractures intersecting the borehole and an estimate of the size of the fracture opening. The observations made with the borehole TV camera have been recorded on video-tape and are available to interested parties. The data obtained from the drill core and the borehole TV camera will be combined with the injection test results and surface measurements of the fracture system in order to determine the directional permeabilities of the fractured rock mass. Figure 81.6 shows the type of information that can be obtained from the drill core, and borehole TV camera and borehole geophysical logging. It is obvious that using one source of data alone could present a very misleading or incomplete picture whereas the use of the three sources of data gives a much more reliable guide to the interpretation of the mechanical properties plus a much more accurate guide to, and assistance with the interpretation of, an injection test program.

Future Activities

It is obvious that during the next several years a series of new sites will be selected by members of the Geological Survey of Canada; these sites will have to be evaluated. A three-stage hydrogeology program has been proposed to evaluate these sites: (1) "a preliminary assessment of a given site" involving a limited (5000 to 10 000 ft.) drilling and testing program designed to evaluate the hydraulic and mechanical properties of the major features at a given site; (2) a second stage "in-depth" assessment consisting of an extensive (20 000 to 40 000 ft.) drilling program combined with detailed injection and pump testing, *in situ* dispersion tests, installation and monitoring of an array of piezometers, studies of surface water hydrology, plus the use of age dating techniques and numerical models to assess the groundwater flow system; followed by (3) a third and final stage that could consist of approximately 20 000 feet of drilling designed to fill gaps in the record and partly to provide design data. In addition to the regular water tests, emphasis in this third stage will be placed on setting up a system of peizometers which will provide background or base values so that any effects of repository construction and operation on the groundwater flow system can be monitored.

Such a three-stage program would require about three years to complete. Each stage would be dependent upon favourable results being obtained from the previous investigations.

Some of the basic questions that this hydrogeology program is designed to answer with regard to the suitability of a proposed radioactive waste repository site are:

1) What are the hydraulic characteristics (porosity and permeability) of the most prominent structural features and the more common fracture sets at a given site?

2) What is the three-dimensional configuration of the groundwater flow system?

3) What are the dispersion characteristics of the fracture system and the potential for mass transport through the fracture system?

4) What is the nature of the interaction of the surface water system with the groundwater flow system? What is the nature of the recharge mechanism in fractured media?

5) What will be the short and long term effects of the underground excavation and changes in the thermal regime on the permeability of the rock mass and the configuration of the groundwater flow system?

6) What is the effect of rock type on the stress-dependent permeability relationship for flow in fracture systems?

7) What is an acceptable level of groundwater flow, during repository construction, operation, and after its operating life, into and through a radioactive waste repository?

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82. GRANULITES, ANORTHOSITES AND COVER ROCKS NORTHEAST OF BAKER LAKE, DISTRICT OF KEEWATIN

Project 760028

Mikkel Schau and L. Hulbert[1]
Regional and Economical Geology

Introduction

A large variety of gneisses of medium to high grade, an abundance of basic or anorthositic plutons, the eastward termination of the Dubawnt Group, the presence of many faults, projected or inferred (Donaldson, 1965; Wright, 1967; Reinhardt and Chandler, 1973; Skippen, pers. comm.; LeCheminant, pers. comm., 1975), complicated aeromagnetic signature (Geological Survey of Canada, 1972), and the presence of a high amplitude, low wavelength, positive Bouguer anomaly (Gibb and Halliday, 1975), suggests that the region northeast of Baker Lake is one in which the long complex history of a part of the northern Churchill Structural Province may be unravelled.

The northern half of map-sheet 56 D/1 was mapped at various scales by a combination of compass and pace traverses across lithologies, as well as following selected marker lithologies and structures. Access to area was mainly by foot, although boat and helicopter facilitated travel to more remote regions. The sketch map (Fig. 82.1) and table of formations (Fig. 82.2) illustrate the main features.

General Geology

A northern gneissic granitoid terrain, a central "supracrustal" sequence, a southern region composed of gneissic and supracrustal groups, with unconformably overlying Dubawnt Group rocks (Donaldson, 1965), and a gabbroic anorthosite body (Wright, 1967) lying along the boundary between the southern and central region, constitute the mapped area (Fig. 82.1). A previously unknown anorthosite complex, gabbro stocks, a diatreme, and a foliated fluorite-bearing granite, were found during the mapping.

Pre-Dubawnt rocks are metamorphosed to amphibolite grade in the north and northeast, and granulite grade in the central and southern parts of the area. Locally the pre-Dubawnt rocks have experienced retrograde metamorphism and are now epidote amphibolite facies. Celadonite, pumpellyite, and prehnite are late degradation and vein products of the higher grade rocks near the Dubawnt unconformity.

Northern Region

The northern region consists chiefly of northeast trending, variably dipping, granitoid gneisses. They are well-veined, layered paragneisses, orthogneisses, and migmatites and show signs of multiple deformations. All contain abundant complex mesoscopic structures, including curvilinear fold axes, conical folds and refolded isoclines. Biotite-rich bands are interlayered with felsic and pegmatitic units, as well as by amphibolite layers of variable thickness. Deformed and fragmented amphibolite dykes cut the complex. In the northwest corner the gneisses are intruded by a fluorite-bearing porphyritic biotite granite with a northeasterly foliation. A small massive granodiorite stock (not shown on Fig. 82.1) is present in the central part of the northern region.

Massive to well foliated, locally garnetiferous, metagabbro stocks form generally east-west zones. They consist locally of plagioclase-phyric, pegmatitic of leucogabbroic phases and contain blocks of orthopyroxene-rich gabbro-norite. Large blocks of gneiss have been stoped into these stocks. The emplacement of the stocks postdates the gneisses and was either coeval or later than the anorthositic intrusions mentioned later. More than one generation of metagabbros may be present in this region because the intensity of foliation within the stocks varies considerably.

Central Region

A well-layered sequence of mafic and, rarely, acid granulites with steeply-dipping, east-west-striking attitudes of remarkable lateral persistence constitutes the major part of the central region (unpatterned on Fig. 82.1). A unit (marker unit 3 of Fig. 82.1) 40 m thick extends for at least 15 km along strike. Most of the material of the central region is rusty weathering, relatively massive, rock composed of two pyroxenes, feldspar ± garnet, ± extremely flattened quartz, ± hornblende, ± brown mica, ± pyrite, ± graphite, and rarely ± calcite, in various proportions. Examples of modal and chemical analysis of silicic, magnesian and calcareous units are shown in Table 82.1 (Anal. 1, 2, 3). Some layers are garnetites or pyroxenites. The central region rock suite belongs to the medium pressure granulite metamorphic facies (Carmichael, 1974).

The determination of the nature of the protolith of these granulites is a difficult problem. The fact that these granulites have been depleted in some elements (*see* Geophysical Surveys), as well as the presence of small orthopyroxene "concretions" postdating some deformation, suggests that some metasomatism has occurred; thus the interpretation of bulk chemical compositions in terms of original compositions will be difficult. The intense flattening of quartz crystals, rock units and folds, makes any deductions as to the structure of the layers prior to their flattening, suspect.

Layering in this unit is in part defined by graphite-rich schists and pyritic horizons and rare carbonate-rich layers. Locally, layering is cut by an anorthositic complex (intersected by the line B-B in Fig. 82.1)

[1]Department of Geology, University of Saskatchewan, Regina, Saskatchewan.

From: Report of Activities, Part A;
Geol. Surv. Can., Paper 77-1A (1977)

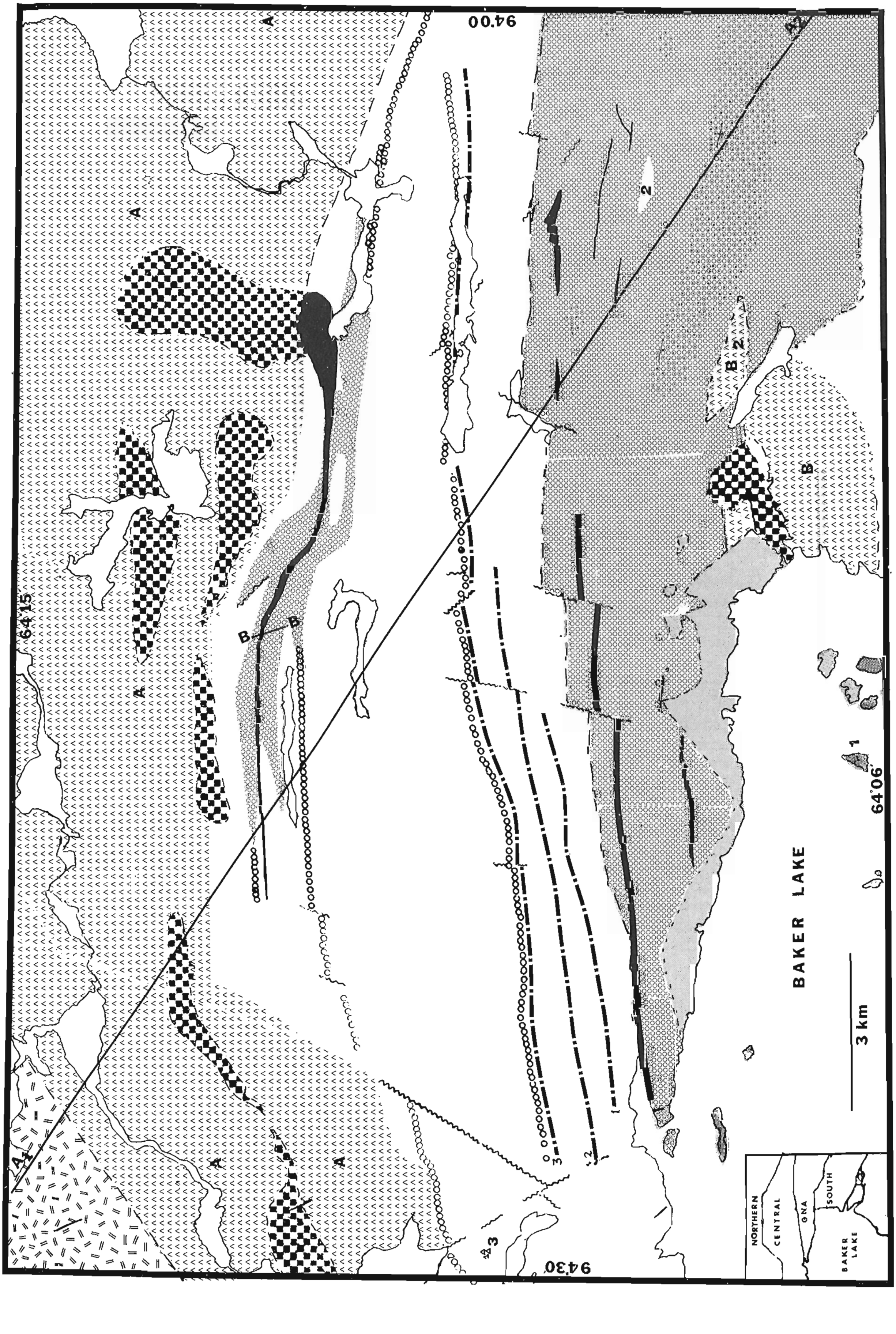
94°00
64°15
64°06
94°30
BAKER LAKE
3 km
A
B
NORTHERN
CENTRAL
QNA
SOUTH
BAKER LAKE

Figure 82. 1.

Figure 82.1. (opposite)

Geological sketch map of map-sheets 56 D/1 and 56 D/2. Location of cross-sections marked by A^1-A^2 and B^1-B^2.

DUBAWNT GROUP
- 3 DIATREME BRECCIA
- 2 DYKES
- 1 LACCOLITHS OR SILLS

GABBRO, USUALLY METAMORPHOSED, MAY INCLUDE OLDER AMPHIBOLITES

FOLIATED FLUORITE GRANITE

GABBROIC-NORITIC ANORTHOSITE COMPLEX
- 4 FERROGABBRO INTRUSIVES
- 3 COARSE-GRAINED, UNSHEARED
- 2 XENOLITHS
- 1 PLAGIOCLASE GNEISS (SILL?)

LAYERED GRANULITES
- 3 QUARTZ-RICH, GR-2PX, GRANULITE MARKER
- 2 2PX-GRANULITE MARKER
- 1 GR-HYP-FP-GNEISS MARKER

GRANITOID GNEISS COMPLEX
- B GRANULITE METAMORPHIC GRADE
- A AMPHIBOLITE GRADE (MAY INCLUDE METAGABBRO AND DYKES?)

although in general, layering in both the anorthositic complex and country rock is similar. The layering predates the flattening because it is locally folded and sheared, high-grade metamorphic minerals are aligned along foliation rather than layering, and pegmatites, themselves with flattened fabric, cut across layering at small angles. This assemblage (excluding the clearly igneous units) is interpreted as sedimentary because the layering is so persistent along strike. The extreme compositions may represent metasomatically enhanced variations within a complex, relatively shale-poor, sequence.

The gabbroic anorthosite is cut by extensive, magnetite-rich ferrogabbro dykes and stocks (shown as a solid black tadpole-shaped area in Fig. 82.1). These rocks have been affected by the same pervasive deformation and metamorphism as the country rock granulites.

After metamorphism, the rocks of the central region were intruded by biotite pegmatites, diatremes and Dubawnt mica-phyric dykes. The diatreme is a breccia consisting mainly of country rock broken, twisted, and cemented with calcite, quartz and powdered rock, and near the southern edge, by a Dubawnt dyke containing partially-resorbed granitoid xenoliths. This is apparently the eastern extension of a similar zone in map-sheet 56 D/2 (A. Miller, pers. comm.).

South Region

Three main groups of rock constitute this region: 1) old gneisses (granitoid gneiss complex B, of Fig. 82.1), 2) layered "supracrustal" metamorphites (shown as blank area on southern edge on Fig. 82.1), and 3) the Dubawnt Group. The old gneisses are two-pyroxene-bearing, rusty weathering complexly folded granitoid gneisses. The dark layers consist of two-pyroxenes ± feldspar (antiperthite) aligned in a steeply-dipping east-west direction. The complex mesoscopic structures seen in outcrops of this group (i.e. the granitoid gneiss complex) are in contrast to the continuity of layering seen in the second group (i.e. the layered "supracrustal" metamorphites). The latter are similar to granulites of the central region.

The Dubawnt Group (Donaldson, 1965; Wright, 1967) lies unconformably on rocks of the south region and on the gabbroic-noritic anorthosite complex, which is discussed later. The basal units consist of locally derived, coarse conglomero-breccias overlain by conglomerates which contain quartz-muscovite and granitoid gneiss pebbles which are not locally derived. These units are overlain by pink, trough cross-bedded, pebbly, arkosic units and locally, ripple-marked red mud or siltstones. Thin beds of volcanic breccias whose fragments are mainly biotite porphyries are locally and sporadically interbedded with these sediments. Sills or laccoliths of mica syenites (red mica traps of Tyrrell, 1898) such as those on Chain or Jessiman Islets, are present at the contact between arkoses and mudstones. They intrude and deform the sediments in a complex manner reminiscent of soft sediment deformation.

Gabbroic-Noritic-Anorthosite Complex

A gabbroic-noritic-anorthosite complex (Wright, 1967) extends at least 22 km along strike and attains widths up to 6 km (*see* index map Fig. 82.1) (Table 82.1, Anal. 6). It has streaky fabric due to deformed and aligned feldspar megacrysts up to 5 cm in maximum diameter set in a matrix of deformed or recrystallized pyroxenes, hornblende, and/or garnet. The colour index generally varies from 15 to 25, although glomeroporphyritic mafic layers and rarer oxide-rich zones (Table 82.1, Anal. 7) are present. Generally white, but occasionally red (Table 82.1, Anal. 5), blocks of anorthosite with rounded to serrated edges are found in the complex. In one locality anorthosite dykes were noted within the body.

The complex is intrusive, as shown by xenoliths of gneissic and layered country rock and local cross-cutting relationships along its southern contact. Locally preserved compositional and textural layering, rarely preserved cumulate textures, and on a finer scale, pyroxenes with exsolution lamellae, suggest that the complex had a cumulate history. Dykes of ferrogabbro (Table 82.1, Anal, 8, 9, 10) cut the body.

Labradorite is the main feldspar of the complex and it suggests that total pressure was not high at the time of igneous emplacement (Yoder, 1969; Green, 1969). The presence of "ilmenite-magnetite" minerals, enriched in some layers and in the later ferrogabbros, suggests that water pressures were never high (Osborn, 1969). The body therefore has many features characteristic of the labradorite-type end member of the anorthosite massif types (Anderson and Morin 1969; Romey, 1968) and was probably emplaced relatively near the surface.

The sheared northern margin of the gabroic-noritic-anorthosite complex proper consists of a garnet-rich, well-layered, extremely feldspathic unit of uncertain origin.

The complex was variously affected by penetrative deformation so that relatively undeformed blocks up to 6 by 1.5 km of coarse grained gabbroic-noritic-anorthosite are surrounded by medium grained, streaky, gneissic or sugary varieties, similar to those described

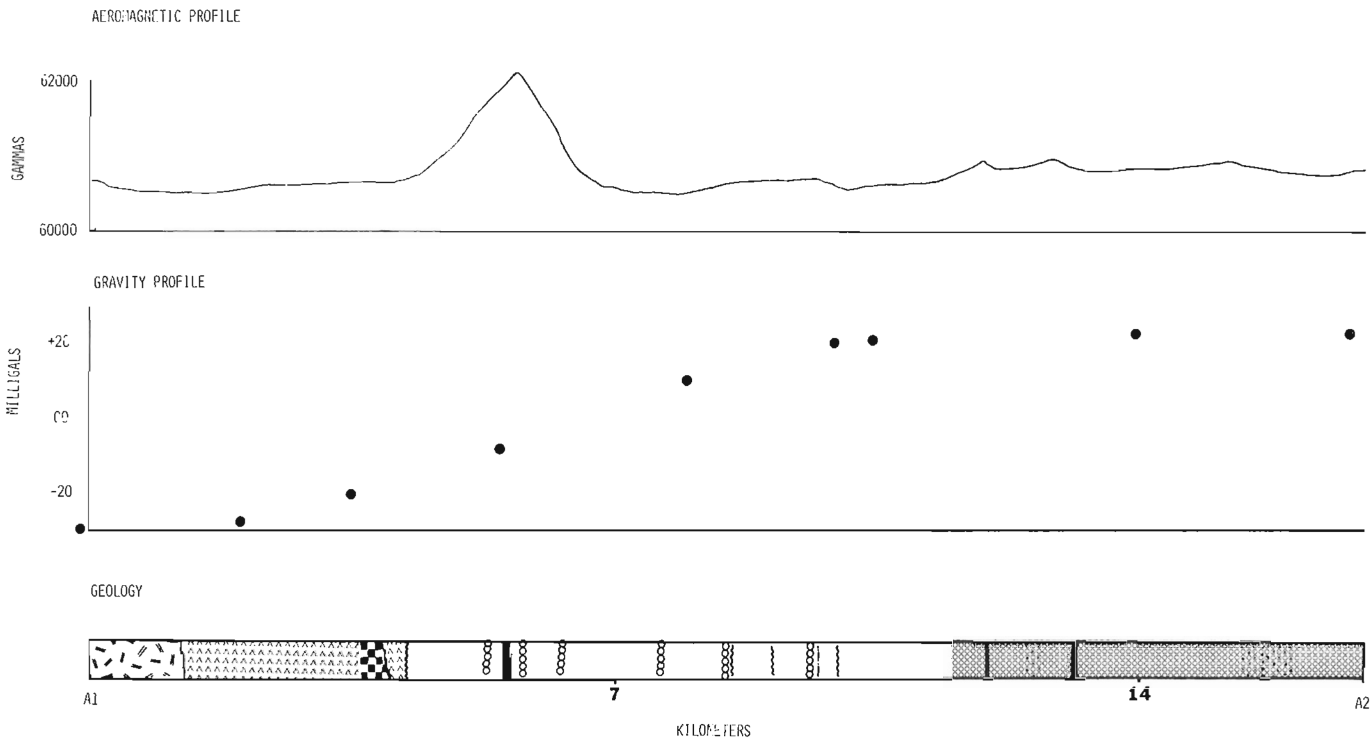

Figure 82. 3. Aeromagnetic and gravimetric profiles compared with geology.

by Kehlenbeck (1972) for Grenville anorthosites. The complex is well foliated, feldspar crystals have been rotated and ferrogabbro dykes occasionally form small, tight folds. Local areas remain where composition and textural layering is at low angles to the steep foliation.

Deformational variations can be seen in a single hand specimen where plagioclase megacrysts vary in thickness normal to foliation from 2 cm to 2 mm in 10 cm. A ferrogabbro dyke 50 m wide pinches to 50 cm in a ductile fault zone.

The penetrative deformation, which is parallel to that seen in the layered sequence, combined with mineral assemblages characteristic of granulite facies (2 pyroxenes, garnet, feldspar) indicate that the complex was brought to crustal levels where it was exposed to conditions characteristic of the granulite metamorphic facies. The presence of layering, as well as other evidence suggesting an igneous cumulate history in a low water pressure environment, indicates that the complex originally was of great lateral extent. Such sill-like bodies are usually emplaced in shallow tectonic environments. Rare minor structures in supracrustal rocks and the gabbroic-noritic-anorthosite complex suggest that an anticlinal axis occurs in the southern part of mapped area. Here the complexly folded "2 pyroxene" gneisses are found, whereas to the north the less complex granulite supracrustals occur where a syncline is to be expected. The gabbroic-noritic-anorthosite complex "sill" may have been emplaced near or along an ancient unconformity which separated "granulite" layered sequence from older "2 pyroxene" gneisses.

Figure 82. 2

Table of formations and correlation between regions.

NORTHERN REGION	CENTRAL REGION	GNA COMPLEX	SOUTHERN REGION dykes + laccoliths
Dubawnt dykes	Dubawnt dykes Diatremes Dubawnt weathering	dykes mud + siltstones conglomerates Dubawnt Group	mud + siltstones volcanic breccia conglomerates Dubawnt Group
PROFOUND UNCONFORMITY			
	medium to low grade metamorphism		
	biotite pegmatites	biotite pegmatites	biotite pegmatites
	INTRUSIVE CONTACT		
GABBRO	GABBRO	GABBRO	GABBRO
RELATION UNCERTAIN	INTRUSIVE CONTACT		
	Deep-seated granulite metamorphism		
DISCONNECTED AT THIS TIME	Ferrogabbro dykes G.N. Anorthosite sills with local compositional layers	Ferrogabbro dykes G.N. Anorthosite complex with local layers and possibly gabbro	
FOLIATED FLUORITE GRANITE	INTRUSIVE CONTACTS		
	LAYERED SEQUENCE PROBABLY METASEDS.		LAYERED SEQUENCE PROBABLY METASEDS.
AMPHIBOLITE DYKES GRANITOID GNEISSES OLD AMPHIBOLITES	NO BASE		CONTACT NOT SEEN
			ORTHOPYROXENE GNEISSES

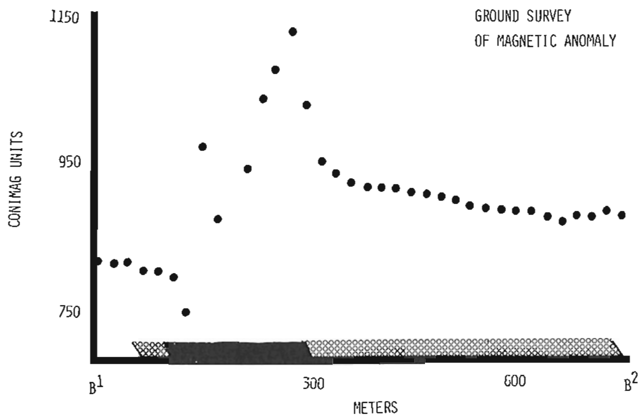

Figure 82. 4. Ground surveyed profile of major magnetic anomaly. Note close correspondence of anomaly to solid black ferrogabbro. The total relief of the anomaly using factors supplied by CONIMAG is about 14 800 gammas.

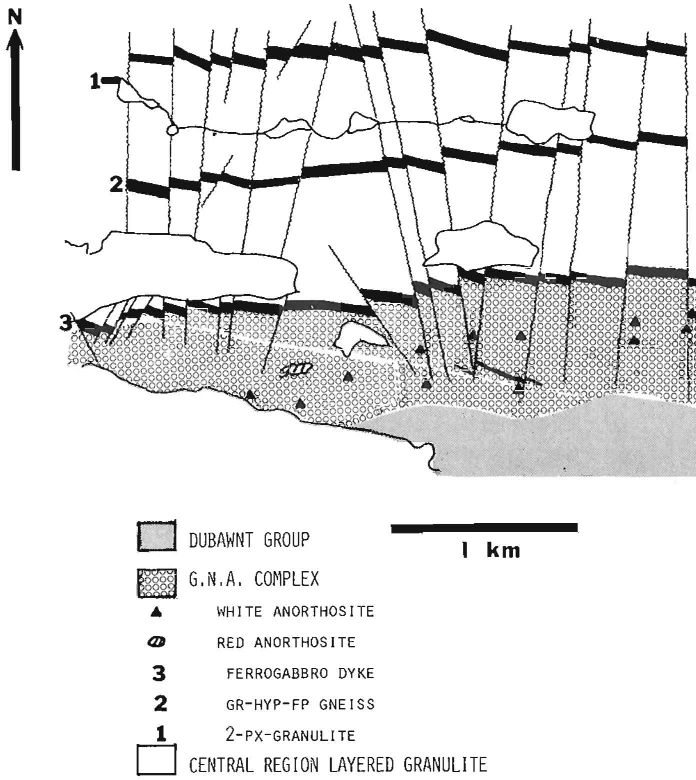

Figure 82. 5. Detailed map of part of central region and gabbroic-noritic-anorthosite complex showing the effect of late faulting.

Mineral Occurrences

Ferrogabbro and oxide-rich layers that contain high Ti and V (Table 82. 1, Anal. 7-10) are present in the anorthosite complexes. Graphite schists are present in granulite terrain enabling the structure to be traced using geophysical methods. Iron sulphides are ubiquitous, though not abundant, and are responsible for rusty weathering of granulites. Coper sulphides are sparsely associated with gabbros and are locally sparsely disseminated in anorthosites and amphibolites. Purple fluorite is present in the granitic intrusion in the northwest corner. Uranium showings are associated with diatremes in an area to the west of map-area (Laporte, 1974) but no such showings are associated with similar diatreme breccia of the present map-area.

Geochemical anomalies of Cu and Pb might be expected in small swampy lakes because the area is one into which weather balloons carrying electronic equipment and batteries, have been dropping for many years.

Geophysical Surveys

Ground-based surveys of gamma-ray spectroscopy were done using Scintrex GIS-3 and TV-2 spectrometers. The majority of the determinations on bedrock of high-grade metamorphic units were near or at lower detection limits. Sands, gravels and boulders throughout the mapped area gave higher readings, which correlated well with the derivation of this material from granites and granitoid gneisses of the northern region. Airborne gamma-ray spectroscopy, which indiscriminately integrates values derived from bedrock and cover, is probably more useful in tracing the glacial dispersion of debris from radioactively high units such as the purple fluorite-bearing granite, than for correlating "depleted" units in anorthosite-granulite complexes. The main conclusion from the ground-based surveys is that the "granulite terrain" is depleted in radioactive elements. This is a well-known feature of granulites (Lambert and Heier, 1968; Eade and Fahrig, 1971).

Ground-based magnetic studies were performed in areas of aeromagnetic highs (Geological Survey of Canada, 1972). A positive anomaly of large amplitude dominates the region mapped (Fig. 82. 3), although an east-west pattern is superimposed within the central region. The main east-west anomaly was traversed several times; an example of a profile (*see* line B-B, Fig. 82. 1) taken with a portable magnetometer (Conimag) is given in Figure 82. 4 and shows a direct correlation with a magnetite-bearing (30%) ferrogabbro dyke. Ferrogabbro dykes in the gabbroic-noritic-anorthosite complex display a much lower response on aeromagnetic maps although their iron content is high. Oxides of the southern complex are now mainly hematite which has replaced ilmenite and magnetite, as indicated by study of opaques in polished sections of ferrogabbro, and by measurements of magnetic susceptibility (Table 82. 1). The reason for the oxidation is not known, but may be related to the irregularly distributed but intense chemical weathering associated with the pre-Dubawnt unconformity.

A positive gravity anomaly of 50 mgal is associated with the gabbroic-noritic-anorthosite complex (Gibb and Halliday, 1975). Steep gradients of up to 5 mgal/km roughly coincide with the contacts of amphibolite grade gneisses with the granulite and anorthosite complex

Table 82.1

Analytical Data for Pre-Dubawnt Rocks (map-sheets 56 D/1 and 56 D/2)

	"Granulites"			"Gabbroic-noritic-anorthosite Complex"						
	1[1]	2	3	4	5	6	7	8	9	10
Major Elements*										
SiO_2	69.9	49.5	46.1	46.5	52.7	52.2	36.1	48.7	40.6	42.8
TiO_2	.56	.26	.15	.26	.14	.23	7.8	1.92	6.0	3.7
Al_2O_3	15.6	19.0	28.9	25.5	23.4	21.4	5.8	12.7	12.8	10.7
Fe_2O_3	.8	.1	.2	1.0	1.9	.5	28.2	9.4	7.9	6.2
FeO	2.7	5.2	3.2	5.1	3.7	6.2	6.8	7.7	18.8	16.4
MnO	.03	.09	.04	.09	.07	.09	.28	.20	.32	.31
MgO	1.54	11.3	2.71	5.16	4.6	4.54	5.43	5.72	3.0	6.0
CaO	1.50	10.6	13.2	13.1	9.76	10.9	6.56	8.85	7.5	8.8
Na_2O	2.7	1.7	2.5	2.0	3.6	3.3	.3	1.7	1.8	1.0
K_2O	4.34	.39	.47	.44	.71	.59	.32	.98	.2	.2
P_2O_5	.09	.04	.02	.07	.04	.05	.07	.25	.34	.24
S	.08	.1	.08	.1	.06	.05	.01	.05	.1	.1
CO_2	-	.3	.1	.0	.5	.1	.6	.7	.0	.4
H_2O	.4	2.0	2.6	1.3	.8	.7	2.2	2.3	1.3	3.8
Minor Elements(%)**										
Zr	.027	.0031	.0026	.0037	.0023	.0031	.016	.013	.029	.017
Sr	.021	.028	.034	.020	.033	.029	.0056	.017	.0076	.011
Zn	.004	.003	.002	.006	.004	.006	.017	.011	ND	ND
Ba	.110	.010	.011	.013	.021	.019	.011	.029	.0038	.0039
Be	<.00030	<.00030	<.00030	<.00030	<.00030	<.00030	.00058	<.00030	.0036	.0040
Co	.0012	.0033	.0014	.0028	.0024	.0034	.014	.0087	.0094	.0097
Cr	.0058	.067	.0032	.0052	.0069	.0043	.0092	.012	.0031	.0032
Cu	.0016	.0045	.00071	.0013	.0032	.0020	.014	.012	.078	.088
Ni	.0013	.012	.0044	.0079	.0079	.0074	.024	.011	.010	.010
V	.0059	.0097	.0065	.0075	.0060	.0071	.15	.052	.13	.11
Y	.0030	<.0020	<.0020	<.0020	<.0020	<.0020	.0062	.0073	.0076	.0075
Yb	.00021	<.00020	<.00020	<.00020	<.00020	.00020	.00029	.00041	.00053	.00047
Sc	.00090	.0021	.0037	.00085	.00072	.0014	.0049	.0037	.0052	.0034
Magnetic Susceptibility***										
cgs units	$<1.x10^{-6}$	$3.x10^{-6}$	$<1.x10^{-6}$	$6.x10^{-6}$	$6.x10^{-6}$	$3.x10^{-5}$	$1.5x10^{-4}$	$5.x10^{-5}$	$3.6x10^{-3}$	$1.5x10^{-3}$
Specific Gravity ******										
	2.75	2.87	2.74	2.91	2.75	2.82	3.34	3.02	3.24	3.10
Modes*****										
Plagioclase	25	50	75	70	85	79	20	35	20	35
Kind	O	B	L	B	L	L	L	L	L	L
Perthite	30	-	-	-	-	-	-	-	-	-
Quartz	28	-	5	-	-	-	-	tr	2	-
Orthopyroxene	2	34	-	-	8	7	5	15	30	15
Clinopyroxene	2	15	-	-	5	6	30	10	10	16
Amphibole	-	tr	15	19	tr	5	-	9	tr	4
Biotite	tr	tr	-	-	tr	tr	-	tr	-	-
Apatite	tr	tr	-	-	tr	tr	tr	tr	tr	-
Garnet	10	-	4	10	-	tr	-	tr	-	-
Chlorite	tr	tr	tr	-	tr	tr	-	-	tr	5
Talc	-	tr(?)	-	-	-	-	-	-	-	tr
Magnetite (M)	tr	-	tr	-	tr	tr	5	5	10	tr
Ilmenite	-	-	tr	-	tr	1	14	5	10	15
Opaque dust in Silicate (M or H)	-	-	tr (M)	-	tr (M)	1 (M)	5 (H)	10 (M)	7 (M)	tr
Equally secondary Hematite (M)	-	-	tr	-	tr	1	20	10	10	10
Pyrite	tr	-	tr	tr	tr	-	tr	tr	tr	tr
Chalcopyrite	-	-	-	-	tr	-	tr	tr	tr	tr
Graphite	2	-	-	tr	-	-	-	-	-	-
Carbonate	-	tr	-	-	-	-	-	-	tr	-
Prehnite	-	-	tr	tr	-	-	-	-	-	-
Pumpelleyite	-	-	tr	tr	-	-	-	-	-	-
Saussurite	-	-	tr	tr	-	-	-	-	-	-
White mica	-	-	-	-	tr	tr	-	-	tr	tr

Footnotes to Table 82.1

* Major elements by GSC XRF methods. GSC analytical laboratories.

** Minor elements by GSC spectroscopic methods, Ag, As, B, Ce, La, Mo, Pb, Sb, Sn less than detection limits on direct reader. Pt not detected using 2659Å line, i.e. less than 150 ppm. GSC analytical laboratories. Accuracy ∿15% of quoted value. ND – not determined.

*** Magnetic susceptibility determined (Christie and Symons, 1974) on analyzed powders (for guidance $2 \times 10^{-3} \sim 1\%$ wt magnetite (Nagata, 1961).

**** Specific gravity determined on whole-rock specimens.

***** Modal analyses by estimating percentage from thin section and polished section, checked by x-ray diffraction. These estimates should be treated with caution when the material is coarse grained. Opaque minerals were identified in polished sections and their approximate proportions are only estimated. O = oligoclase, L = labradorite, B = bytownite.

No norms are published since they are so dependent on oxidation state of rock and observed oxidation is thought to be secondary.

[1] Sample 1: SMA6-0017A, a white weathering, porphyroblastic garnet +2-pyroxene granulite with extremely flattened quartz ribbons and feldspar augen. Zone 15, UTMN 7116170, UTME 429830.

Sample 2: SMA6-0040A, a massive, medium grained, rusty weathering 2-pyroxene mafic granulite unit from granulite terrain. Zone 15, UTMN 711-6010, UTME 430380.

Sample 3: SMA6-0099A, a hornblende-bearing feldspathic gneiss from granulite terrain. Zone 15, UTMN 711-5960, UTME 429430.

Sample 4: SMA6-0113A, a medium grained feldspathic gneiss of possible igneous origin in lower granulite or upper amphibolite grade. Zone 15 UTMN 7117710, UTME 429850.

Sample 5: SMA6-5006A, a blood-red anorthosite from the gabbroic-noritic-anorthosite (GNA) complex. Zone 15, UTMN 7112880, UTME 431430.

Sample 6: SMA6-5068A, typical streaky, gabroic-noritic-anorthosite from GNA complex. Zone 15, UTMN 7113160, UTME 432350.

Sample 7: SMA6-5010C, oxide-rich layer from GNA complex. Zone 15, UTMN 7112760, UTME 431180.

Sample 8: SMA6-5010B, ferrogabbro from GNA complex. Zone 15, UTMN 7112760, UTME 431180.

Sample 9: SMA6-5016, ferrogabbro from GNA complex. Zone 15, UTMN 7113140, UTME 431180.

Sample 10: SMA6-5046, ferrogabbro along strike of above dyke from GNA complex. Zone 15, UTMN 7113250, UTME 431680.

(Fig. 82.3). Rock density data shows a correlation between the exposed rocks and the gravity anomaly. The difference in surface rock density between the high-grade complex (specific gravity 2.70-3.30) and the granite gneisses (specific gravity 2.61-2.72) is great enough to explain the observed anomaly.

Integration of the seismic, gravity, and magnetic data with the geological information now available, will allow selection of the "best" fit model of the crust in this region. The association of anorthosites and high-grade gneiss with relatively high gravity anomalies in the Baker Lake region was suggested by Heywood (pers. comm.) and Gibb and Halliday (1975). Thus, the presence of "anomalous crust" (Gibb and Halliday, 1975) from Baker Lake to Southampton or Coats Island becomes a key feature in models of tectonic evolution of the northern Churchill Province. It is of special interest to note that Archean supracrustal units in belts north of the Baker Lake region (for example, the Prince Albert Group, Schau, in press) have few features in common with Archean supracrustal units south of it, such as the Kaminak Group (Ridler and Shilts, 1974).

Structural History

The tectonic history of the region is complex and is not fully known, but it can be divided into two contrasting phases. The first phase was associated with granulite metamorphism and affected anorthosite complexes, layered rocks and gneisses deformed during a still older phase. It involved flattening and extensive recrystallization, with some local, minor, mesoscopic folding, and possibly with the development of folds whose closure extends well beyond the limit of the area mapped. For example, layers as thin as 2 m can be traced along strike for over 20 km, with no closures evident.

The second phase is one characterized by brittle, high level fracturing, local hydration and degradation of the high-grade metamorphic rocks. The exact timing of the various fracture sets is not known, but north-south shears of both sinistral and dextral movement, some filled with biotite-pegmatites, preceded easterly and west-northwesterly-directed shearing (Fig. 82.5). The apparent offsets appear to diminish northward and there is a suggestion that scissor faults were active for some time. Shearing in most of these directions has affected the Dubawnt Group. Since pegmatites do not cut this group, but the shears do, shearing is in part along old reactivated planes. The net effect of all the shears on the rocks is surprisingly little, allowing the tracing of marker horizons along strike. One example of the shear systems is shown in Figure 82.5. Here, tracer beds in supracrustals and some ferrogabbro dykes are mapped to show the stepping effect of the shearing yielding a slightly more northerly strike for the unit than that observed in the individual segments. Units offset on this scale are continuous for tens of kilometres.

Tectonic History: Problems to be solved

The model presented below is also a statement of problems to be solved.

1) The complexly deformed gneisses are probably early Archean basement to the Archean "supracrustal pile" (for example, the unpatterned layered granulites of Fig. 82.1) into which gabbroic-noritic-anorthosite bodies were emplaced at shallow depths.

2) The whole sequence was dehydrated, decarbonated and depleted after having been deeply depressed into the crust and deformed along east-west tectonic axes (Archean (Kenoran?) tectonic regime?).

3) Granulite blocks were subsequently raised along faults and gabbros were emplaced along the faults.

4) Later uplift, brittle deformation and emplacement of pegmatites took place during the earlier part of the Hudsonian(?) Orogeny. Continued uplift and erosion took place.

5) Continental clastic and, rare, but explosive, alkalic volcanic rocks of the Dubawnt were deposited, intruded and deformed in the jostlings following the "Hudsonian orogenic event".

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Project 760025

W.W. Heywood
Regional and Economic Geology Division

Field work during the 1976 field season completed the mapping of the northern half of the Amer Lake area (NTS 66 H) for publication on the scale of 1: 250 000. No field work is planned for 1977; the southern half is scheduled for completion in 1978. The area is part of Operation Baker reconnaissance mapping (Wright, 1967), originally published on the scale of one inch to eight miles. The present investigations are intended to upgrade the knowledge of the Aphebian metasediments, the Archean(?) metavolcanic rocks and a wide variety of Aphebian and/or Archean basement granitic and gneissic rocks.

Basic metavolcanic rocks outcrop in a narrow band that extends westerly from 96°15'W, 65°28'N, and as isolated remnants associated with diorite on the southern border of this outcrop band. These rocks are massive, very fine grained, locally amygdaloidal, moderately deformed metavolcanics of andesitic to basaltic composition. Tuffaceous layers and minor interbeds of felsic volcanics are locally present. Banded iron-formation consisting of thin layers (0.1 to 0.5 cm) of magnetite alternating with thin siliceous layers are exposed over a thickness of 10 m at the eastern end of the volcanic band. Doubtful pillows suggest that the volcanics are stratigraphically below the Aphebian quartzites.

Low weathering, scattered outcrops of metamorphosed diorite are intimately associated with discontinuous remnants of basic volcanics south of the main band of volcanic rocks. The diorite weathers dark green to almost black and is dark mottled green on fresh surfaces. These rocks are medium- to coarse-grained, massive to moderately foliated, and consist of altered plagioclase in a matrix of fine grained altered mafic minerals. In places the diorite contains altered inclusions of basic volcanic rocks and ultrabasic rocks.

Moderately to intensely metamorphosed peridotite occurs as small outcrops associated with the diorite and metavolcanic rocks. It ranges from fine- to coarse-grained and is generally extensively altered with the development of soapstone in places. Near the east end of the volcanic belt serpentinite contains a small amount of asbestos. Brown weathering discontinuous outcrops of very fine grained ultrabasic rock, in part schistose, is interlayered with the basic volcanic rock. In one place a spinifex texture is preserved.

The name 'Amer Group' is informally applied to the generally east-west trending belt of sedimentary rocks (Hurwitz Group of Wright, 1955) of probable Aphebian age in the central part of the area. Quartzite, feldspathic sandstone, conglomerate, slate, carbonate rocks and derived schist and phyllite are represented.

The lowest unit consists largely of white quartzite that forms prominent ridges throughout the area. The quartzite is generally massive, but in places it is thin to thick bedded, pure white, or less commonly, pink. It is composed almost entirely of fine- to medium-grained sugary to glassy quartz. Quartz pebbles and cobbles form conglomerate lenses as much as 3 m thick. Primary structures such as bedding, cross-bedding and ripple marks are not well preserved; however, a moderate to well developed fissility probably represents bedding. Minor impure quartzite, cross-bedded feldspathic quartzite and quartz-muscovite schist, are interbedded with the white quartzite. Dolomitic limestone overlies the white quartzite unit. In general it is a recessive unit and outcrops are widely scattered along linear belts. In many places black slate, phyllite and schist overlie the dolomitic limestone. These are followed by pink to reddish feldspathic quartzite and sandstone. Layers containing carbonate detritus are interbedded with the feldspathic quartzite. The amount of black slate increases and entirely constitutes the uppermost units.

All units described above are intruded by late granitic rocks.

The 'Amer Group' rocks are intensely folded and faulted. Northwesterly, predominantly left lateral faults, displace the Amer Group sediments and all older rocks. Within the gneissic areas to the north of the 'Amer Group', the fault traces are commonly marked by lineaments. Thrust faults and overturned folds characterize the northeastern part of the 'Amer Group' and are present in the vicinity of Amer Lake. To a great extent these have formed through shearing along overturned limbs of recumbent folds; some faults appear to become folds along strike. The boundary between the 'Amer Group' and the mafic volcanic group is marked by several high angle faults.

Gneisses of varying composition, texture and structure underlie most of the northern half of the map-area. Although grey, medium grained gneisses with irregular foliation and layering are typical, these rocks range from well-layered paragneiss to granitoid gneiss of unknown origin. Paragneiss zones are commonly heterogeneous. Thinly banded rocks are interlayered with amphibolites or amphibolite-rich zones and rusty garnetiferous schist and gneiss. The general aspect suggests a sedimentary origin. Migmatites are widely distributed. Quartz monzonite to granite plutons intrude all gneissic rocks. In part these are of Aphebian age.

Pink weathering, generally leucocratic granite or granodiorite ranges from fine grained to pegmatitic. Near Amer Lake they are intrusive into the lowest 'Amer Group' quartzite units. Elsewhere, similar granitic rocks are associated with the older gneisses and may be much older. Separation of these two granitic rocks has not been possible.

The northwestern corner of the area is underlain by plutons of pink and white porphyritic granite. Outcrops

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are widely scattered in a broad sandplain crossed by bouldery eskers. The pink granite is porphyritic, fresh and undeformed, with abundant feldspar phenocrysts up to 5 cm long. The white granite, generally porphyritic, ranges from equigranular, medium grained to pegmatitic. It is characterized by the development of faint to moderate schistosity.

Augen gneiss forms a band 3 to 10 km wide along the eastern boundary of the plutonic granites in the northwestern quarter of the area. In part these gneisses appear to be derived from the white porphyritic granite, and elsewhere appear to be porphyroblastic phases of the grey granitic gneiss.

A northeast trending fault extends from 65°30'N on the western boundary of the map-area, to the northeast corner of the map-area. The fault is characterized by extensive cataclasis with extensive mylonitization in a zone as much as 5 km wide. Mylonitized remnants of 'Amer Group' quartzite, dolomitic limestone, and slate or shale are present, as well as the granitoid gneisses.

Diabase dykes are widely distributed but not abundant in the area mapped. They are as much as 20 m wide, but most range from 2 to 4 m wide. None were traced for more than a few hundred feet. Two trends are apparent, north-northeast and northwest. The north-northeast dykes are commonly amygdaloidal.

Lamprophyre dykes, 1 to 2 m wide, were observed as intrusive into feldspathic quartzite and metavolcanic rocks. They are fine grained, rusty weathering rocks with abundant biotite phenocrysts as much as 10 mm in diameter.

A few small occurrences of sulphide mineralization are present in the metavolcanics. A small occurrence of quartz magnetite iron-formation is associated with the basic metavolcanic rock at the eastern end of the volcanic belt. Near the centre of the map-area, south of Amer Lake, magnetite iron-formation is associated with biotite schist and amphibolite. Uranium mineralization in 'Amer Group' sediments has been described by Laporte (1974).

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GEOLOGY OF THE SHAKESPEARE-DUNLOP LAYERED GABBRO-ANORTHOSITE INTRUSION, ONTARIO

E.M.R. Research Agreement 1135-D13-4-30/76

R.S. James[1] and B.J. Harris[1]
Regional and Economic Geology Division

Introduction

This intrusion occupies an area of 12 km^2 and is exposed immediately west and north of Agnew Lake (in Dunlop and Shakespeare townships) 6 km north of Webbwood, Ontario. It is the largest of several intrusions of its kind in these two townships and unlike the remainder, which form narrow sill-like bodies along the Southern province-Superior province boundary, this intrusion lies 3 km to the north and is totally enclosed within Archean granites which it intrudes (Card and Palonen, 1976). Robertson (1976) and Cape (1973) have described intrusions in the Massey area which are clearly identical to these sill-like boundary intrusions. These gabbro-anorthosite bodies should be distinguished from the Sudbury gabbro-Nipissing diabase intrusions which are obviously younger in age.

The present study represents the initial part of a project the objectives of which are to describe in more detail than is presently available the geology, geochemistry and economic potential of the gabbro-anorthosite intrusions at the base of the Huronian sequence in the Sudbury-Elliot Lake region.

Previous mapping of the Shakespeare-Dunlop intrusion (Card and Palonen, 1976; Card and Innes, 1971) suggests that it is a thick, funnel-shaped, sill-like or lopolithic mass. This conclusion is based on the attitudes of magmatic layering in various parts of the intrusion (Fig. 84.1). Layers vary in thickness from a few centimetres to near 100 m, and are due to changes in the proportions of felsic and femic minerals, grain size variations, and textural changes. Plagioclase feldspar (probably labradorite) and augitic pyroxene are the major rock-forming minerals. However, augite is rarely observed; actinolite and blue-green hornblende replace pyroxene and are the common femic minerals. Rock types in the intrusion range from pyroxenite to anorthosite; the intermediate members of this series are by far the most abundant.

Results of 1976 Field Activities

Data compiled during two months of field mapping in the western half of the intrusion are presented below. Three major zones are recognized, two of which have been divided into subzones. We interpret the outer and western margin of the intrusion to represent the lowest exposed stratigraphic level. In the following section zones and subzones are described in order of increasing stratigraphic level. Their distribution is shown in Figure 84.1.

Zone I - Anorthosite Zone

This zone represents the border and stratigraphically lowest portion of the intrusion. It never exceeds 800 m in thickness (plan view) and normally is nearer 500 m. Five subdivisions of this zone have been recognized; three of these units (Ic, d, e) are probably, in part, stratigraphic equivalents.

I-a: Glomeroporphyritic Anorthosite, Anorthositic Gabbro. Anorthosite consists of aggregates of glomeroporphyritic plagioclase feldspar and interstitial pyroxene. The glomerophenocrysts are spherical to ovoid (2-10 cm in maximum dimension); within the central portion of each, feldspar poikilitically encloses pyroxene. Anorthositic gabbro contains fewer glomerophenocrysts set in a medium grained gabbroic matrix. Although individual outcrops normally exhibit a uniform percentage of glomerophenocryst to matrix gabbro, a sufficient variety in percentages between outcrops suggests that local crystal settling has operated within this unit. At only one locality has this been clearly observed; there a 3-m-thick unit grades from coarse grained glomeroporphyritic anorthosite at its top to a more mafic anorthositic gabbro at its base; a second unit of the same type in sharp but irregular contact underlies it. We conclude that crystal settling, operative once the layer had formed, is responsible for the observed variation in rock type.

Associated with these rocks are breccia dykes which are normally 0.5-2.0 m wide, but occasionally form zones 60-80 m in width. They consist of rounded, ovoid fragments (normally up to 1.5 m in diameter) of anorthosite, anorthositic gabbro, gabbro (all of local derivation) and basement granite, set in a fine grained matrix. Textures in the matrix suggest flowage at elevated temperatures. The orientation of the breccia dykes and zones approximately parallel the contact between the anorthositic rocks and Archean basement. We believe that the breccias may have formed during emplacement of the intrusion, which implies that the anorthosites were largely solid prior to that time.

I-b: Anorthosite and Anorthositic Gabbro. These rocks are massive and coarse grained. They form ridges 40-70 m in height in the western marginal area of the intrusion. Compositional layering is almost entirely absent and the glomeroporphyritic character of the previous unit is not present. At one locality clear deformational features in the vicinity of the contact with basement rocks were recognized. Here, narrow (1-2 m) bands of pyroxenitic gabbro/pyroxenite within thick and massive units of anorthositic gabbro form a 20-30-m-wide zone in which were observed drag-folded lenses of pyroxenitic gabbro up to 0.5 m

[1]Department of Geology, Laurentian University, Sudbury, Ontario

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in length within anorthosite, and vice versa. Mapping suggests that this is a contact facies of the intrusion.

The deformed material must have been solid but also quite plastic during the formation of these features.

I-c: Anorthositic Layered Sequence. Rock types vary from anorthosite to pyroxenite. Outcrops representing this subzone lie inside the marginal portions of the intrusion and probably stratigraphically overlie units I-a and b. Anorthositic gabbro and gabbro form thick (30-50 m) massive layers hosting: (1) layers about 10 m thick which have a 0.5 to 1.0-m basal pyroxenite zone overlain gradationally by pyroxenitic gabbro, gabbro, anorthositic gabbro and finally anorthosite and (2) massive layers of medium to coarse grained anorthosite 0.3 to 2 m in thickness. Both types of layering dip at angles from 20°-30° towards the centre of the intrusion; strike directions approximately parallel its margins.

In the southeastern part of lot 3, concession 6, Gough Township, disseminated pyrrhotite, pyrite and chalcopyrite occur in the basal pyroxenitic zone of a differentiated layer which is in shear contact with a diabase dyke that strikes 075° and dips 80°N. Traces of mineralization occur intermittently over a strike length of 300 m parallel to the pyroxenite-diabase intersection. Estimated total sulphide content is near one per cent. Pyrrhotite is the dominant sulphide mineral.

I-d: Northern Layered Sequence. Alternating layers, 100 to 200 m thick, consisting of coarse grained gabbro/anorthositic gabbro comprise this zone. Layer boundaries appear to parallel the margins of the intrusion. Rarely the same type of layering is observed on a much smaller scale (i.e. 1 to 2-m-thick coarse gabbro layers in medium grained gabbro). This unit overlies I-a; the contact appears to be gradational over a distance of 20 to 40 m.

I-e: Southern Porphyritic Gabbro. This unit forms the marginal facies in the southwestern portion of the intrusion. Its maximum thickness (plan view) is 600 m. The dominant rock type consists of plagioclase phenocrysts 1 to 3 cm in length in a medium grained, massive gabbroic matrix. Layers of coarse grained gabbro from 1 to 40 m thick are present but not abundant; layer contacts are sharp and continuous over 100 m where observed. Minor zones (layers?) in the stratigraphically lowest part of this unit appear to consist of anorthositic and pyroxenitic gabbro and

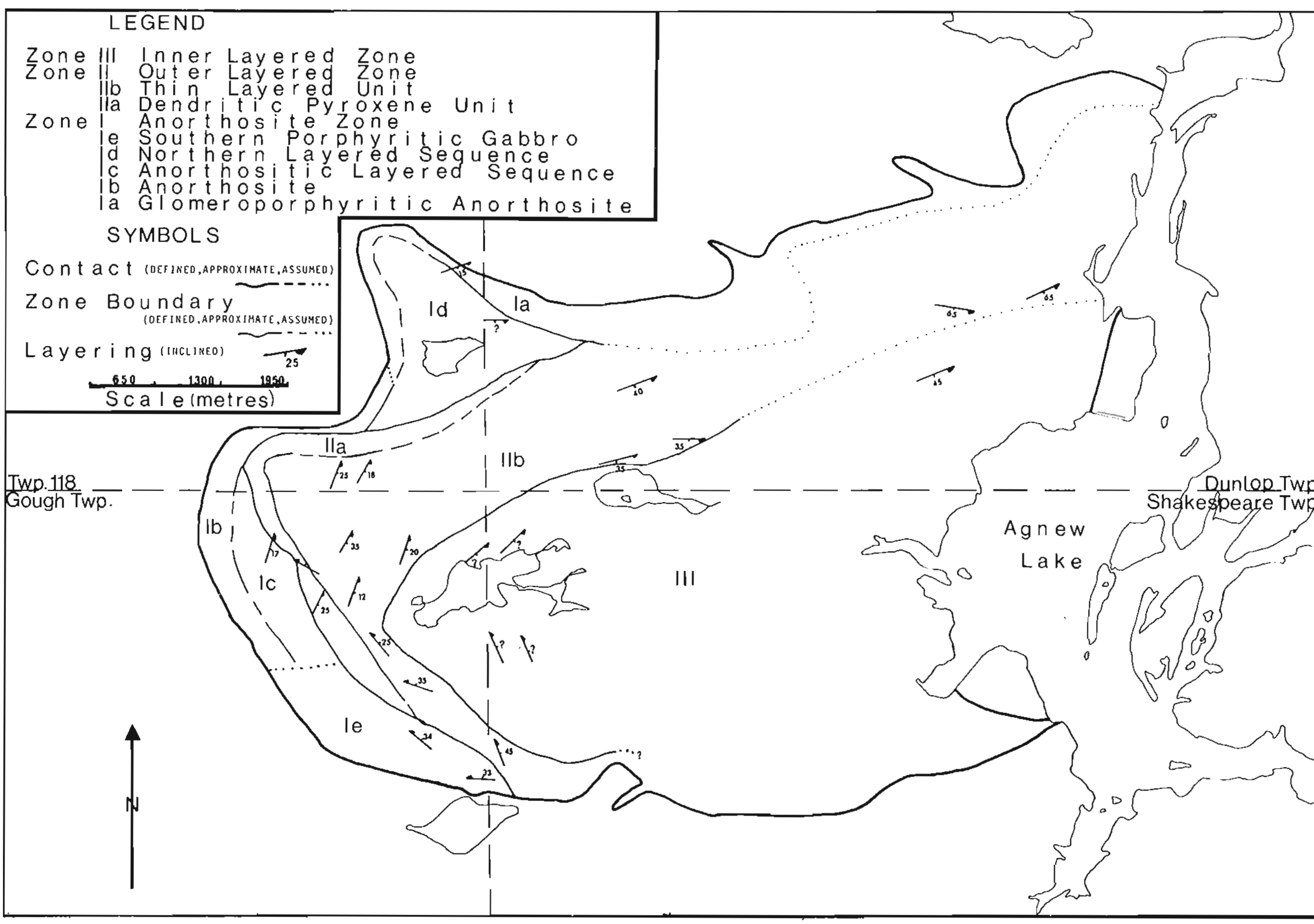

Figure 84.1. Geology of the Shakespeare-Dunlop gabbro-anorthosite intrusions.

suggest a correlation between a portion of this unit and I-c. Massive porphyritic gabbro is in sharp contact with Archean granitoid rocks at the outer margin of the intrusion in this area; no chilled margin has been observed.

Outer Layered Zone - Zone II

This zone stratigraphically overlies all of the rock units described in zone I. Coarse grained gabbro, porphyritic (plagioclase) gabbro and medium grained gabbro are the major rock types. They form thick layers (30-150 m) which are in sharp contact with each other and are continuous along strike for hundreds of metres. This zone has been subdivided into two units on the basis of the character of much thinner layers hosted by the previously described major rock types.

II-a: Dendritic Pyroxene Unit. This unit forms the lower member of the Outer Layered Zone; its maximum thickness in plan view is approximately 300 m. The characteristic feature of this unit is the presence of coarse grained pyroxenitic gabbro as layers, ovoid patches and rarely as narrow dykes in which the pyroxene occurs as continuous and often curved, prismatic crystals up to 10-15 cm in length; aggregates of these crystals form interlocking dendritic patterns. Thick layers of medium grained gabbro together with minor coarse gabbro act as host rocks. Layers containing the dendritic pyroxene are most abundant in the lowermost portion of this unit. They range from 4-20 m in thickness and are continuous over a strike length of at least 100 m. Internally, textures are uniform throughout and boundaries with adjacent rock types are sharp. Ovoid patches (0.5 to 1 m in diameter) of the same material have finer grained margins. Adjacent to the layered zones they are abundant and randomly distributed; in addition they have been observed at much higher stratigraphic levels than the layered material. The upper boundary of this zone has been drawn where the patchy material disappears. Its lower contact with the Anorthosite Zone (unit I) is not exposed.

II-b: Thin Layered Unit. This unit forms the upper portion of zone II. In plan view it may be nearly 1500 m thick in the northwestern portion of the intrusion; a measured section in the west-central part of the complex indicates a thickness of only 150 m. The characteristic feature of this unit is the presence of thin layers 1 cm to 1 m thick within much thicker units of a contrasting rock type. Medium grained gabbro, pyroxenitic gabbro, anorthosite, and coarse gabbro in order of decreasing abundance, constitute nearly 85 per cent of the thinly layered horizons. Pyroxenite, anorthositic gabbro, and rarely porphyritic gabbro are much less abundant. Thick layers of porphyritic gabbro, coarse grained gabbro (present in a ratio near 2:1) and minor (> 10 per cent) medium gabbro are the major rock units in this zone. Only the latter two rock types host the thinly layered horizons, and of these, 70 per cent of the layering is within coarse grained gabbro.

Contacts between the thin layers and host rock are sharp; the layers are normally continuous over at least 50-100 m and dip towards the centre of the intrusion at angles from 10°-40°, averaging 26°. The layers are normally distinguished from host rock on the basis of mineral proportions; grain size differences occasionally defined a layer. In a few instances mineral proportions varied regularly within a layer in the vertical dimension (i.e. pyroxenitic gabbro at base to anorthositic gabbro at top), indicating that crystal settling was operative once the layer was in place. The upper contact of this unit with the Inner Layered Zone is drawn at the point where the thinly layered horizons disappear; this contact is sharp where it has been observed.

Inner Layered Zone: Zone III

This zone occupies the central core area of the intrusion and is characterized by thick repetitious layers of coarse grained gabbro which is occasionally anorthositic, medium grained porphyritic (plagioclase) gabbro, and medium grained gabbro. Volume percentages of these rock types are 48, 17, and 31 per cent respectively based on a measured section of the lower 630 m of this zone. Pyroxenitic gabbro (3%) and pyroxenite (1%) represent the remaining recognized rock types. Normal thicknesses of layers for the three main rock types are: coarse gabbro, 50-80 m, porphyritic gabbro 30-65 m; medium gabbro, 25-125 m. Normally, porphyritic gabbro overlies with a sharp contact coarse grained gabbro; 0.5-1 m diameter patches of the coarse gabbro in the porphyritic phase are commonly observed adjacent to the boundary of the two units. Contacts between medium and coarse grained gabbro are similarly sharp; patchy coarse gabbro occurs in the medium grained phase where the latter rock type is the younger unit. Sharp contacts characterize the relationship between the medium and porphyritic gabbro; no clear evidence of crystal settling as a mechanism linking these two rock types, was observed.

In our measured section medium grained pyroxenitic gabbro and pyroxenite occur together as layers 18 m and 5 m thick; the former rock type overlies the latter and the contact is gradational. Medium grained gabbro underlies the pyroxenite; the contact between the two was not observed; coarse gabbro overlies and is in sharp contact with the pyroxenitic gabbro. The pyroxenite-pyroxenitic gabbro association is reminiscent of the lower portion of layers in the inner part of the anorthosite zone (unit I-c); tentatively we suggest that this type of layering has formed by crystal settling of pyroxene *in situ*.

Summary

The following generalizations are possible:

1. Structures in the lower portion of the Anorthosite Unit (i.e. I-a and b) suggest that the intrusion was emplaced after portions of this unit had solidified.

2. The boundaries between individual zones in Figure 84.1 and between layers within zones II and III normally parallel the margins of the intrusion. Layers exhibit shallow dips (25° average); dip directions outline a trough within the centre of the complex — its axis plunges at a shallow angle to the east. These features were probably produced by one or more post-intrusion deformational events.

3. The Inner and Outer Layered Zones consist of thick repetitious units of medium and coarse grained gabbro and porphyritic gabbro. Within the latter zone thin-scale layering is prominent. Anorthosite is a common rock type only in the marginal zone of the intrusion. In this same zone pyroxenite and pyroxenitic gabbro are much more abundant than at higher stratigraphic levels. Gravity data (Popelar, 1971) indicates that at a depth there exists a significant mass of high density (pyroxenitic?) material.

4. Layering has resulted from cumulus action in a primary environment. Both plagioclase and pyroxene have been primary precipitates and intercumulus phases. Variation in grain size of the thick layers may be due as much to cumulate growth as to cooling rates. Patches of coarse, porphyritic and dendritic material found at stratigraphically higher levels than corresponding layers may represent residual pore material trapped during the cumulate process. The repetitious, almost cyclic nature of the layering in this intrusion, may well be evidence for repeated influxes of fresh magma into the system.

This report has been compiled from field observations and megascopic descriptions of rock types. Detailed petrographic and geochemical studies of typical material from each zone in the complex is now in progress.

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THE GEOLOGY OF THE HACKETT-BACK RIVER GREENSTONE BELT – SECOND PRELIMINARY REPORT

Project 750008

R.A. Frith, W.K. Fyson[1], and J.D. Hill[2]
Regional and Economic Geology Division

Introduction

Field work during the summer of 1976 completed the field component of this project aimed at producing a 1: 100 000 map and report for open file release, and a final 1: 250 000 map and report for the Nose Lake (76F, E/2) and Beechey Lake (76G, W/2) map-areas. More detailed work was carried out along the cross-sections illustrated in Figures 85.1 and 85.2 and in the vicinity of the Hackett River Dome (Figs. 85.1 and 85.3).

Assistants involved in this year's field mapping included: J.D. Hill, who is studying the granitic rocks of the area as part of a Ph.D. thesis; J.A. Percival, who is studying the metamorphism and petrogenesis of the basal metasediments in the Hackett River area as part of an M.Sc. thesis at Queen's University; and J. Ostler, who is studying the sedimentation and petrogenesis of rocks along Back River, south of the map-area, as part of a M.Sc. thesis at Carleton University. During the season W.K. Fyson came to study structural relationships around the Hackett River area, and T. Ewing (with K. Condie), a M.Sc. student from the New Mexico Institute of Mining and Technology, collected volcanic rocks to study their bulk and rare-earth geochemistry.

The authors gratefully acknowledge the co-operation during the season of B. Shepard and L. Saliken, Brascan Resources; W. Padgham and L. Beaumont, Department of Indian Affairs and Northern Development; and B. Mioduszewska and A. Boronowski of Comino.

Regional Geology

Bryan, *et al.* (1975), Jefferson, *et al.* (1976 a, b, c) and Padgham, *et al.* (1975 a, b, c) have mapped parts of the area and described their observations in marginal notes. Frith and Hill (1975) briefly discussed the major rock units of the area and outlined their stratigraphy. Further work this season confirmed most of the stratigraphic relationships and has provided new data for a more comprehensive review.

The "Basement" gneisses (unit 1) of the Hackett River Dome were interpreted to be the core of a mantled gneiss dome. Initially the core may have been intrusive, but now only deformed granitic gneiss is present. Younger monzonite and alpite rocks intrude these genisses. Anthophyllite-bearing inclusions similar to the surrounding metasediments are present as structural inliers or as roof pendants in the centre of the dome. The contact with the metasediments is faulted or gradational, with younger monzonitic rocks obscuring the relationship.

The metasediments (unit 2) that appear to overlie the "basement" gneisses of unit 1 may be divided into three units: a quartzofeldspathic lower member; a middle conglomeratic member; and an upper sericitic member. Sillimanite grade metamorphism and intense deformation obscures primary features, but bedding is preserved in some of quartzofeldspathic members and some of the conglomeratic members contain boulder trains. The individual clasts are mostly unrecognizable, but where deformational shadow zones are present pebble-like shapes and gneissic cobbles were identified. The conglomeratic members are locally referred to as the "Gob" because of the characteristic quartz clasts that are sporadically present over most outcrop surfaces. The clasts are not generally in contact with each other. The matrix is made up of mafic and aluminous minerals that locally include anthophyllite, biotite, chlorite, muscovite, garnet, sillimanite or kyanite, staurolite and cordierite, with lesser amounts of plagioclase and quartz. The magnesium- and aluminum-rich composition of the matrix suggests that the initial sediment may have been volcanogenic and derived from basic volcanic rocks. The presence of granitic clasts in the conglomerate implies a bi-model provenance, and the sorting suggests that the rock is a paraconglomerate.

The volcanic rocks (unit 2) make up the Hackett River greenstone belt. They conformably overlie the sedimentary rocks (unit 2) which in turn overlie (unconformably?) the "basement" gneisses (unit 1). The volcanic rocks in the northern half of the belt appear to be homoclinal, facing toward the overlying sedimentary rocks (unit 4). The belt thickens toward the west, where it becomes migmatized in the vicinity of Mara River (Fig. 85.1). The belt is truncated on the west by a granitic gneiss complex (Section B-B^{111}, Fig. 85.2). The southern part of the belt is wider and is broken up by intrusions of diorite and granodiorite that produce hybrid agmatitic rocks of uncertain petrologic affinity. Here the widest part of the belt contains exhalite facies rocks on both sides, which may delineate paleobasin margins (section C-C^{1} Fig. 85.2). The southern part of the belt narrows to an outward-facing anticlinorial structure, as illustrated in section D-D^{1}, Figure 85.2.

The nature and composition of the original volcanic pile is unknown, but it is suspected that the more acidic upper part of the pile is vestigial and the lower more mafic part of the sequence has been removed by tectonic processes. The more mafic parts were probably downwarped and replaced by granitic basement gneiss

[1]University of Ottawa, Ottawa, Ontario.

[2]University of Western Ontario, London, Ontario.

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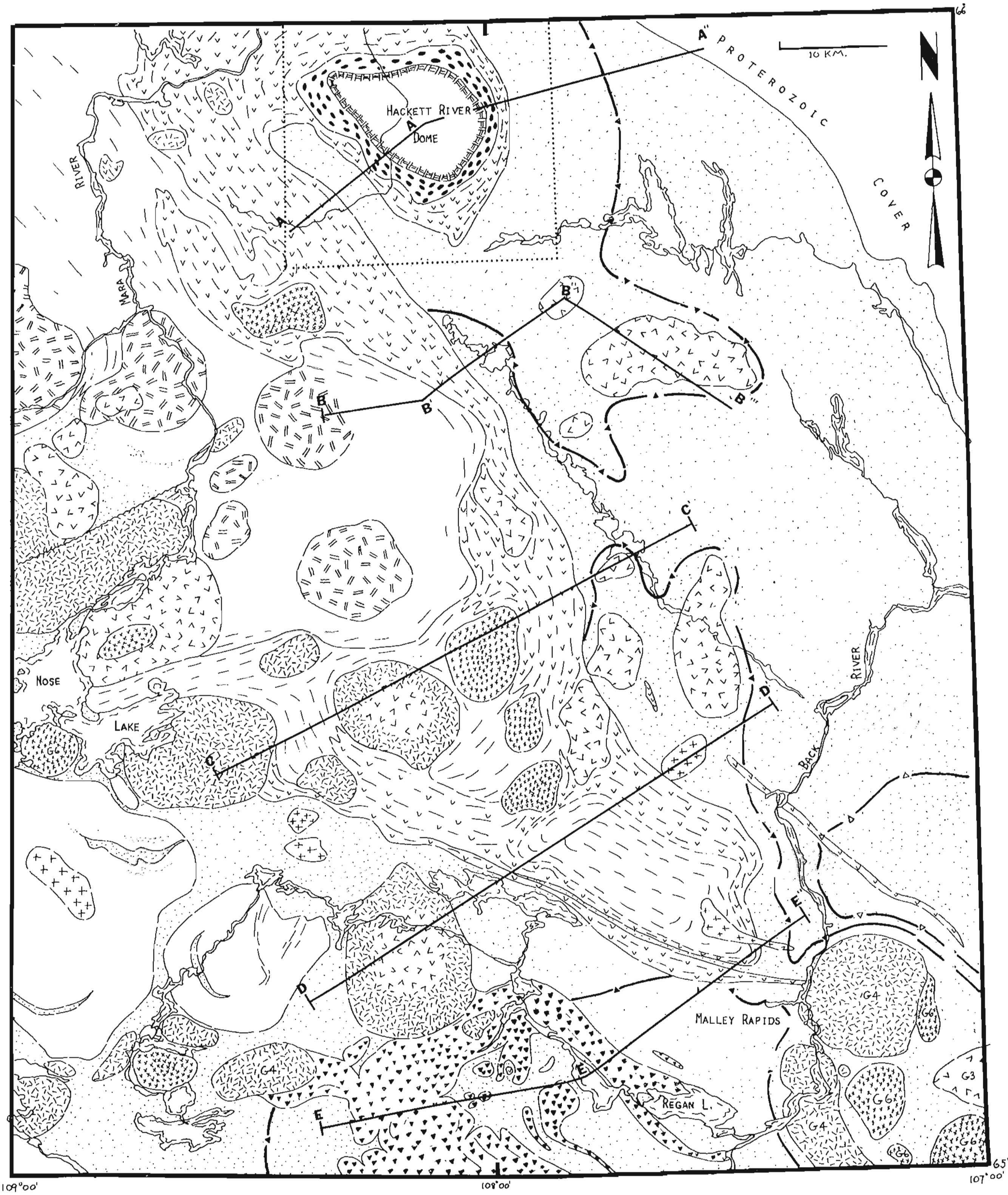
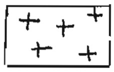
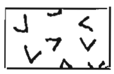


Figure 85. 1. Diagrammatic sketch map of the geology of the Beechey Lake, W/2 and Nose Lake, E/2 map-area. Areas of detailed work are outlined by the dotted lines and by cross-sections illustrated in Figure 85. 3. Geology is adapted in part from mapping by the Department of Indian Affairs and Northern Development.

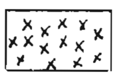
 (M) Migmatite derived from metasediments (M4) or volcanic rocks (M3)

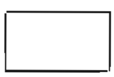
 (G1) Coarse grained undeformed pegmatitic granite or quartz monzonite, commonly muscovite-bearing

 (G2) Porphyritic or porphyroblastic quartz monzonite, granodiorite with locally aplite

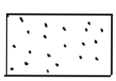
 (G3) Quartz monzonite stocks, some forming central cores with granodioritic margins (G4)

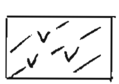
 (G4) Intrusive stocks of granodiorite, locally porphyritic

 (G6) Dioritic, medium- to coarse-grained, locally with amphibolite inclusions

 (6) Granitic gneiss complex consisting of indifferentiated quartz monzonite tonalite, pegmatite, aplite and lesser amounts of units 1-5

 (5) Back River volcanic belt of andesitic flows, pillow lavas, breccia and rhyolitic flows, tuff and breccias

 (4) Metagreywacke meta-argillite and minor iron-formation, lamprophyre dykes

 (3) Hackett volcanic belt of pillowed basalt, andesite, agglomerate, tuff and rhyolitic pyroclastics, flows, tuff and breccia

 (2) Metasedimentary "Gob" unit of quartzofeldspathic gneisses, quartz pebble paraconglomerate and muscovite, quartz schist

 (1) Biotite granodioritic gneiss with secondary quartz monzonite, pegmatite, aplite and "roof pendants" of unit

Isograd - staurolite andalusite

Isograd - biotite

complex and magmatic plutons that moved upward. The remaining belt is made up of about one third acid volcanic rocks such as rhyolite flows, agglomerates, tuffs, welded tuffs and breccias, and two thirds andesitic or basaltic agglomerates, tuffs and pillowed flows.

Exhalite facies rocks (Ridler, 1973) are located at or near the volcanic-sedimentary interface, which generally corresponds closely to the subaerial, submarine transition. Three exhalite facies rocks are recognized, although intermixing is normal: iron facies (banded iron-formation); carbonate facies; and sulphide facies. The predominance of one type over another appears to be controlled by the proximity and type of volcanic rocks. The chemical sedimentation of exhalite rocks is normally mixed with volcanoclastic, sedimentary and pyroclastic rocks.

There is mineralization of commercial interest near Hackett River at the Bathurst-Norsemines deposit and near the narrow midpoint of the belt at the Yava deposit (Frith and Hill, 1975). The Hackett River deposits occur in five known zones (MacNeil, 1976) along a

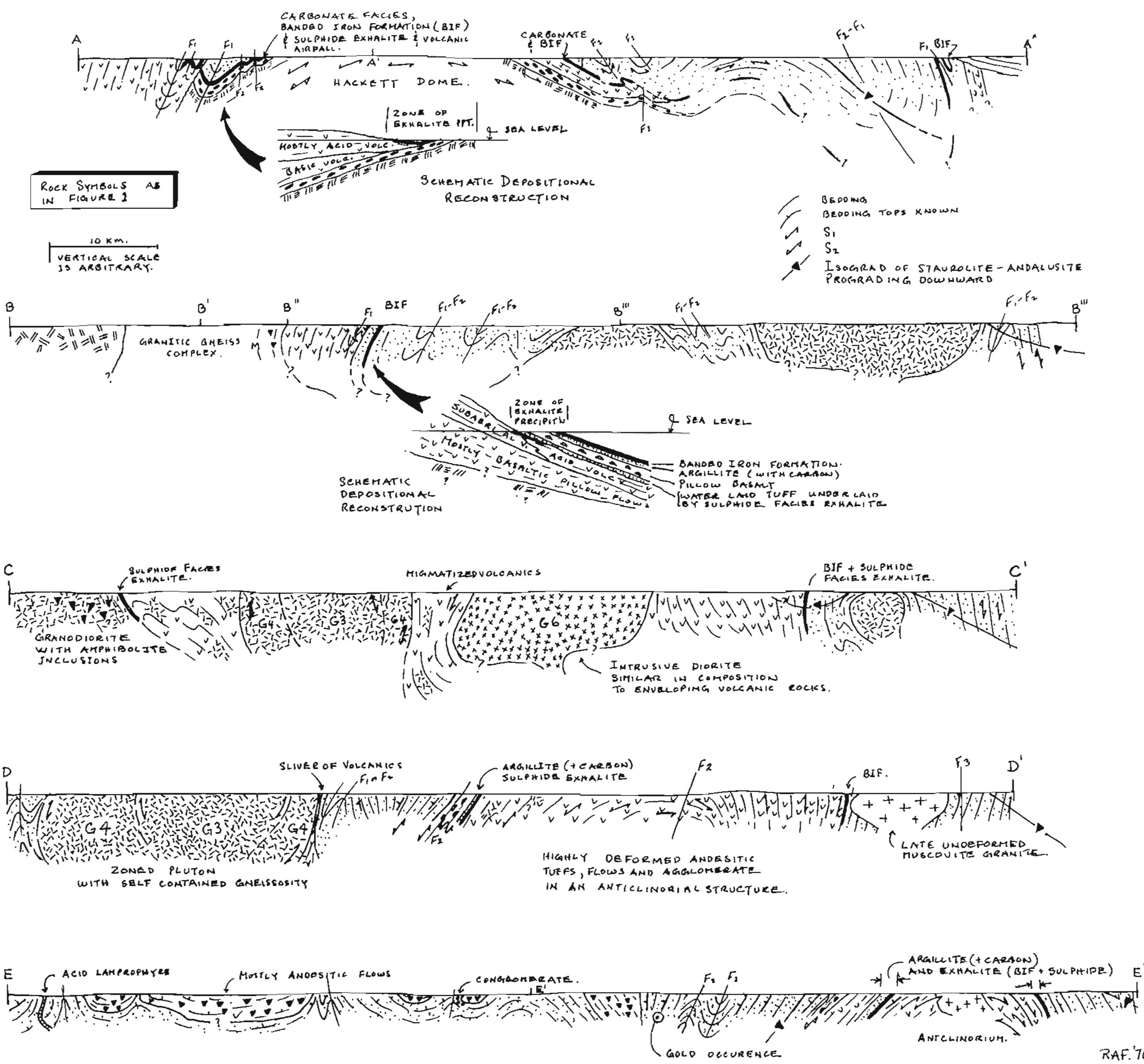


Figure 85. 2. Diagrammatic profiles of the Hackett and Back River greenstone belts. The depositonal environment of the Bathurst-Norsemines and Yava mineral deposits is illustrated.

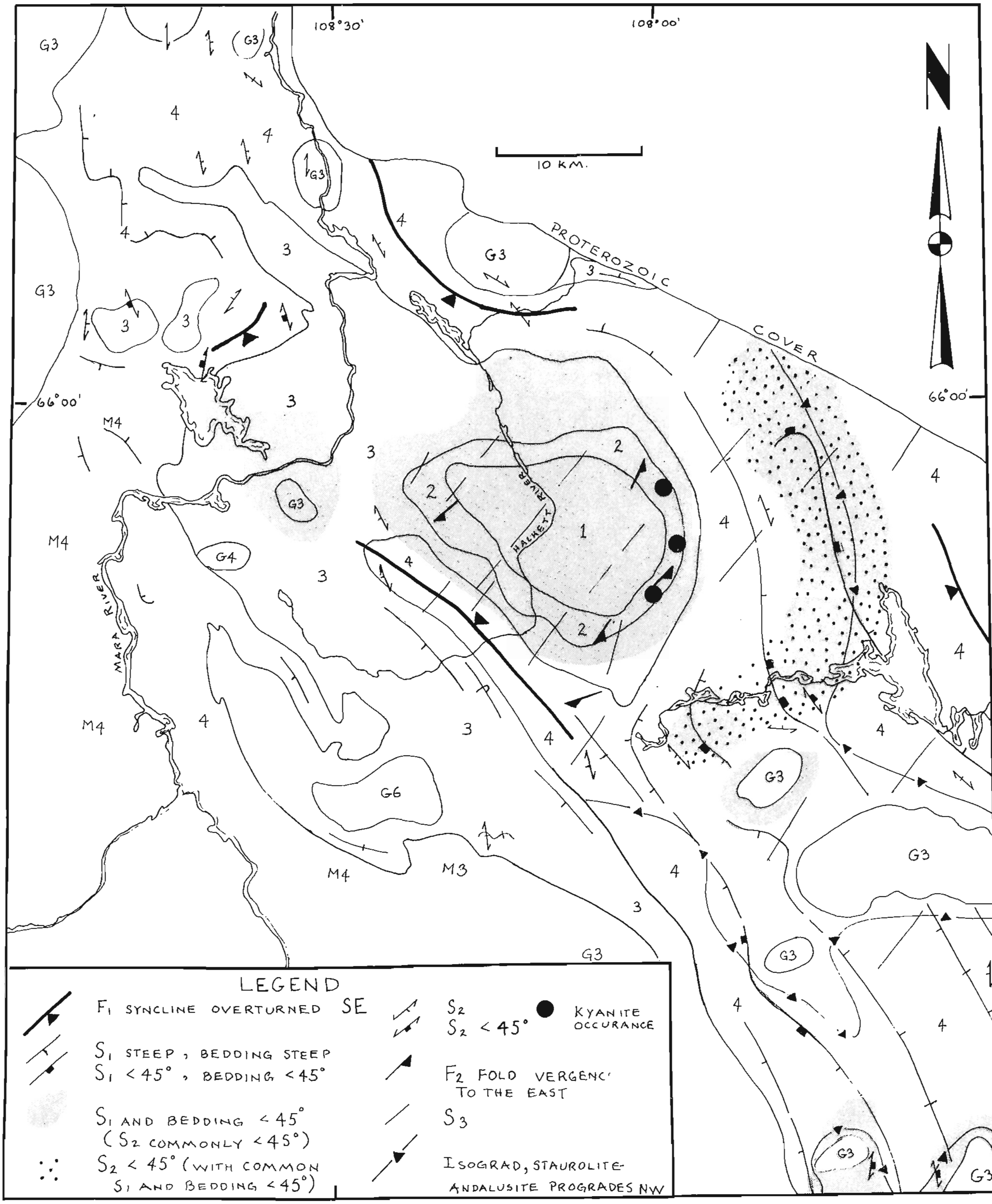


Figure 85. 3. Structural sketch map of the Hackett River area to illustrate the domal uplift of post-S_1 and S_2 axial planar surfaces and its relationship to rock types and metamorphic isograds. Rock unit numbers as in Figure 85. 1. Geology is adapted in part from mapping by the Department of Indian Affairs and Northern Development.

Table 85.1

Characteristics and time relations of structures and metamorphism of the northern part of the Hackett River greenstone belt around the Hackett dome

Deformation	D_1	D_2	D_3
Fold style	Large-scale F_1 isoclines upright northwest trending or overturned in towards dome.	Small to large open F_2 folds; steep axial surfaces strike northwest to north. In dome, folds are tight, axial surfaces are flatter and vergence is southeast.	Small open upright F_2 folds trend northeast.
Axial surface foliation	S_1 bedding plane muscovite and quartz; inclusion trails in porphyroblasts; segregation foliation in dome.	S_2 widespread, mostly schistosity with aligned biotite, muscovite and segregation foliation.	S_3 is discontinuous as muscovite in crenulations, fractures.
Mineral growth: sillimanite, kyanite, staurolite, andalusite, cordierite biotite muscovite			
Diapiric (x) and pluton (+) emplacement			

mineralized horizon that is at least 15 km long. The mineralized horizon is located mainly at the volcanic-sedimentary interface, but the volcanic rocks in this region are very thin and low-dipping. Exhalite facies rocks are abundant and volcanic airfall pyroclastic rocks are intimately associated with the sedimentation. At Camp Lake "A" zone it was observed that the sulphide facies directly overlie the metasedimentary rocks of unit 2. A schematic reconstruction of the depositional environment is illustrated in section $A\text{-}A^{11}$, Figure 85.2.

The Yava deposit is similar in some respects to the Bathurst-Norsemines deposit, except that exhalite precipitation occurs within the top part of the pile rather than at the sedimentary interface. The mineralized horizon is steeply dipping and underlain by a thin veneer of pillowed basalt which overlies a thick sequence of welded tuff and subaerial basaltic flows. The mineralized horizon is in turn overlain by thin water-worked tuffs and an approximately 100-m-thick pillowed basalt sequence. Graphitic argillite and banded iron-formation overlie the basalt. The deposit appears to be related more to mafic than acidic volcanism. However, lateral variation characterizes this part of the belt and rhyolitic flows and breccias are present both to the north and south. A depositional reconstruction of the Yava deposit region is illustrated in section $B\text{-}B^{111}$, Figure 85.2.

The southern part of the belt is more highly metamorphosed than the central part, but volcanic textures are well preserved around the edge of the belt. Exhalite facies rocks and carbonaceous argillite are common in the volcanic-sedimentary transition zone. Thin interlayered volcanic belts southwest and northeast of the main belt are made up of basaltic or andesitic pillow basalt, tuffs and agglomerates, previously correlated with the Back River greenstone belt (Frith and Hill, 1975). These are now thought to be a later phase of the Hackett belt. Good top determinations in the overlying sedimentary rocks are used to draw the $E\text{-}E^{11}$ profile of Figure 85.2 illustrating this stratigraphy.

The anticlinorial structure that characterizes the southern part of the belt consists of highly deformed and presumably higher metamorphic grade andesitic tuffs, agglomerates, and less abundant amphibolitic interlayers. Frith and Hill (1975) reported that carbonate-bearing (exhalite facies) rocks occur within the structure, which we now consider to be inliers or keels. More detailed work on the southern flanks of the structure showed that the rocks are dominantly acidic tuffs. Garnet is present in the lower grade rocks, but toward the centre of the structure the volcanics become more andesitic and develop a characteristic knobby appearance due to plagioclase porphyroblasts.

The metasediments (unit 4) that overlie the Hackett River greenstone belt are typical of the turbidite mudstones of the Yellowknife Supergroup. Near the volcanic contact exhalite facies banded iron-formation and carbonaceous shale are usually present. The turbidites that overlie them are thick to thinly bedded and range in composition and grain size from coarse sandy greywacke to fine argillite. Most of the rocks along the Back River within the study area have not been metamorphosed. The isograds on Figure 85. 1 show that metamorphism in some areas is less than biotite grade.

A northwest trending sedimentary basin is proposed, bordered on the east by the Hackett River greenstone belt, and by another hidden greenstone belt beneath the Proterozoic Goulburn Group. Near this Goulburn unconformity, iron-formation (Section A-A[11], Fig. 85. 2) is exposed which may be extended by aeromagnetic extrapolation under the Goulburn. The basin is presumed to open toward the southeast. Along the northeast stretch of the Back River (Fig. 85. 1), a very thick sequence of northeast-facing metasediments was mapped. The true thickness is not known due to slab-like thrusting of sedimentary rocks along fold hinges which has caused some repetition of beds. However, the thickness may still be as much as 3000 m. Broadly speaking, the sediments may be divided into a lower, middle and upper sequence. The lower units are argillaceous mudstones, the middle are argillaceous to sandy greywackes, and the upper are mostly coarse bedded, sandy greywackes.

The Back River greenstone belt (unit 5) conformably overlies the mudstone turbidites of unit 4. In the contact zone thin carbonate-rich clastic sandstone overlying fine grained argillaceous turbidites of unit 4 is in turn overlain by pillowed andesite. Quartz and volcanolithic pebbly conglomerate are locally present along the contact. In one place, large conglomerate boulders (up to 50 cm diameter) underlie the volcanic succession. Minor sulphide facies exhalite rocks were observed at the contact. Within the volcanic sequence subaerial rhyolite domes both massive and brecciated are found, which locally shed sediment into basins, the margins of which are marked by pebbly, sometimes carbonate-rich volcanogenic sediment. Also, small areas of turbidites with iron-formation that overlie the volcanic sequence are present, but not shown on Figure 85. 1. The volcanic rocks of this belt have been further described by Frith and Hill (1975) and Lambert (1976).

The granitic complexes (units 6, G2 and M) occur on the west side of the Hackett River greenstone belt and as expansive ovoid bodies in the southwest corner of the map-area. The complexes include migmatites (unit M) derived from units 4 and 3. Muscovite-rich granitic to monzonitic masses intrude the complex, and aplitic dykes are abundant. The granitic complex around Mara River (Fig. 85. 1) is made up of migmatites derived from units 3 and 4 (M3 and M4). However, much of the area is made up of heterogeneous monzonitic to tonalitic gneiss of uncertain origin. It is tempting to think that the gneisses are a hybrid derived in part from older granitic basement rocks, but where this relationship is more certain, as at Hackett River, the gneisses are far less granitized in appearance.

The porphyritic or porphyroblastic bodies (G2) found within the granitic complex around the Mara River have indistinct contacts with their host rocks. The megacrysts of potash feldspar are variable in size and abundance. The megacrysts make the rock appear massive and homogeneous, but close observation of the matrix shows that the proportion of biotite, quartz and plagioclase varies significantly. The megacrysts may have formed by some metasomatic process involving some upward diapiric movement.

The plutonic stocks (units G6, G4, G3, G1) were intruded into metamorphosed sediments of unit 4 and the volcanic belts of unit 3 and 5. The oldest types are diorite (G6) in composition and spatially related to the Hackett River greenstone belt. Within this belt the boundary relationships are not always clear. Agmatitic migmatite and inclusion-rich border phases are common. Granodiorite (G4) and monzonitic (G3) rocks form round, presumably tadpole-shaped intrusions (Figs. 85. 1, 85. 2) that are regionally abundant, extending from this study area almost to Great Slave Lake. The Malley Rapids granodiorite body intrudes unit 4 and the metamorphic aureole is a hornfels, but the thermal effects are noted only within a 200 to 500-m zone, suggesting that the intrusion is epizonal to mesozonal. The stocks have a faint self-contained gneissosity. Some of the bodies are zoned with granodiorite margins and monzonitic cores.

Late muscovite-bearing granitic plutons (unit G1) intrude all older rock types. They are more irregular in shape, and widely but irregularly distributed throughout the area. Figure 85. 1 shows only the larger bodies. The granites appear to be undeformed except by a late northeast cleavage. The muscovite granites are volumetrically insignificant but in outcrop they tend to stand out in topographic relief. They are generally coarse-grained and massive. Pegmatitic phases are common and usually contain muscovite books, some being 10 cm wide or more.

Structural deformation and metamorphism

The isograds drawn on Figure 85. 1 surround areas of granitic intrusion, highly deformed volcanic rocks, as well as the porphyroblastic metasediments. The isograd positions are thought to be tectonically controlled by vertical movements centred principally around tectonic "highs" such as the Hackett River Dome, the volcanic anticlinorium northwest of the Malley Rapids, the Malley Rapids group of intrusions, and possibly, uplifts in the Nose Lake and Mara River region. In one area studied in some detail (Hackett River dome) these uplifts were found to postdate the first and second formed fabric found in the supracrustal rocks.

The structural trends within the sedimentary regions of the map-area are northwest. At Malley Rapids, these trends are deflected by the intrusion of a granodioritic stock. The bedding (S_0) and an early fabric (S_1) are refolded along northwest-striking axes. A crenulation cleavage (S_2) is associated with this folding. However, on the limbs of the second folds, sinistral and dextral S_1 and S_0 intersections show that F_1 isoclines are present, even though closures are usually not observed. A northeast crenulation cleavage (S_3) is developed throughout the area, but it is undeflected by igneous intrusion, suggesting that it postdates units G2-G6. This same cleavage is unconformably overlain by the Proterozoic Goulburn Group, setting an upper limit to its age.

The porphyroblasts developed in the metasedimentary rocks (cordierite, andalusite or staurolite) were formed after the first period of folding. Cordierite porphyroblasts are locally cleaved by both S_2 and S_3.

The Hackett River region contains one of the few occurrences of kyanite in the Slave Province. It was first reported by Padgham, *et al.* (1976b). Another locality along Hiukak River east of Bathurst Inlet is the only other known locality (Fraser, pers. comm.). In thin section the kyanite is overgrown by cordierite and muscovite, which suggests that high temperature but high pressure conditions, followed low pressure-high temperature conditions. The metamorphic minerals around the Hackett River Dome have a more complicated mineral paragenesis that may be due to a longer time of metamorphism over a wider range of conditions. More detailed mineral study is needed to work out these parameters. Table 85.1 incorporates a preliminary mineral paragenesis as it relates to structural development in the Hackett River Dome.

Contact metamorphism occurs around some of the larger granitic stocks. In the Malley Rapids area contact relationships were studied in detail and the aureole of hornfels was shown to contain granular to decussate porphyroblasts of garnet, staurolite and/or andalusite that postdate S_1 and S_2 fabrics.

Structural development of the Hackettt River Dome

The F_1 synformal axes outlined in Figure 85.3 show that the Yellowknife Supergroup sedimentary rocks were folded around a tectonically positive "basement" high. The F_1 folds, such as the Hackett River syncline (Jefferson, *et al.* 1976a) are upright or overturned. The early fabric (S_1) developed at this time may be observed as a mica foliation in the central core gneisses and in the overlying sediments as a muscovite-quartz bedding plane cleavage. The cobbles in the paraconglomerate unit are also flattened along S_1.

The isoclinal F_2 fold axes strike northwest to northeast obliquely to the F_1 fold axes, modifying and deflecting the F_1 axes around the Hackett Dome. The F_2 axial planar surfaces (S_2) are shallow-dipping around the dome and around nearby plutons (shaded areas, Fig. 85.3). Some S_2 surfaces are even horizontal. F_2 fold pairs around the limbs verge consistently southeastwardly, implying that uplift of the dome postdated (and did not cause) F_2 folding. The F_2 axes steepen away from the dome in the southwest, but to the north the axes do not steepen as expected. This is not completely understood.

The S_2 surfaces east of the dome are generally parallel to both the bedding and S_1, and dips of all surfaces have been flattened, as shown in Figure 85.3 and section A-A[11], Figure 85.2. This is in contrast to the region north of the dome where S_2 is steep and S_1 is shallow. This may be explained by assuming that the early deformation was due to regional lateral compression and gneiss doming of a lighter granitic basement. The lateral compression produced near-vertical axial planes away from the dome, but above it vertical compression flattened axial planes (Dixon, 1975). Higher temperatures during the second deformation gave rise to more horizontal than vertical compression above the dome.

An F_3 folding has steepened most axial planar surfaces along northeast-trending axes. S_3 fabric is observed discontinuously in the dome and in the supracrustal rocks as a jointing or crenulation cleavage. In the more argillaceous rocks muscovite has developed parallel to S_3.

A structural-time synopsis of the Hackett River Dome is shown in Table 85.1.

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ADDENDUM – December 1, 1976

p. 416 Figure 85. 1, section A-A"; eastern plutons patterned G4 should be patterned G3.

p. 419 Figure 85. 3, caption; "illustrate ..... relationship" should read "illustrating where S_1 and S_2 axial planar surfaces are shallow-dipping and the relationship to"

p. 421 Second par., line 2; "east" should read "west".

p. 421 Second last par., line 11; "post-date" should read "accompany both deformations that produced"

p. 422 Last section, first par., line 5; after "overturned" add "toward the Hackett River dome suggesting subdued relief during folding."

THE WINDSOR GROUP OF CAPE BRETON ISLAND, NOVA SCOTIA

Project 740110

H. H. J. Geldsetzer
Regional and Economic Geology Division

Introduction

The Windsor Group of Cape Breton Island was the object of a detailed stratigraphic investigation during the summer of 1976. Particular attention was paid to the lower part of the sequence which is locally the host rock for base metals and various industrial minerals. This work is the first phase of a continuing study of the sedimentary and diagenetic development of Carboniferous to Triassic strata in the Appalachian Region. The main objective of this report is to describe the distribution of sedimentary facies during early Windsor time and to shed some new light on the correlation of Lower Windsor rocks by means of lithostratigraphic evidence. Biostratigraphic data based on spores (Utting, 1977) and conodonts from the Lower Windsor are being evaluated in order to supplement existing biozonations by Bell (1929) and Mamet (1970).

The Windsor sediments of Cape Breton Island were deposited in a marine environment with approximately north-south trending depositional zones. The facies changes across this depositional strike are illustrated by composite sections from four key areas (Table 86.1).

Bell's zonation of the Windsor rocks on the basis of macrofaunas (Bell, 1929) placed the laminite and overlying massive evaporite of the Lower Windsor Group into a biostratigraphic vacuum, called the subzone A. The lack of datable fossils left this zone open to constant time-stratigraphic speculations. The two end members of this speculation series seem to be publications by Mamet (1970) who considers the subzone A as time-equivalent to subzone B and Schenk (1969) who regards the laminite as a time-transgressive response to a slowly transgressing sea. Base metal occurrences in the Lower Windsor such as copper at the Windsor-Horton contact (Binney, 1975) and lead-zinc in the Lake Enon area, the Middle River area and elsewhere in Cape Breton Island, require a solution to this stratigraphic dilemma in order to facilitate further exploration work.

Basal Windsor Carbonates

The basal Windsor sediments of Cape Breton Island are represented by two major carbonate facies, a western stromatolitic laminite (commonly referred to as the A-1-Limestone) and an eastern fossiliferous carbonate sequence. The two facies join along an approximately north-south line running from Cape North through Ingonish, Cape Dauphin, Eskasoni to the southern coast about 10 km east of St. Peters (Fig. 86.1). A strong deviation from this north-south line in the region of the southern Bras d'Or Lake is due to an anticlinal structure in the underlying rocks.

The stromatolitic laminite rests generally on clastics of the Horton Group and locally on basement as along the Strait of Canso near Mulgrave, north of Mabou Harbour and along the southeast shore of St. Ann's Bay. Whereas the first two occurrences appear to have been local basement highs projecting through a rather low-relief region underlain by Horton clastics, the last occurrence lies along the western edge of a region which rose gently to the east and was underlain by pre-Carboniferous basement and Horton sediments of variable thickness.

The fossiliferous carbonate facies of the basal Windsor is found in this eastern region. Variations within the eastern facies are related to pre-Windsor paleogeographical structural trends. Thus, massive coquinoid brachiopod accumulations follow northeasterly striking granitic anticlinal ridges which plunge to the southwest under Horton sediments. The coquinoid carbonates are not restricted to a granitic subcrop but spread locally onto the Horton cover. A reverse condition occurred along the southeast shore of St. Ann's Bay were stromatolitic laminite spread beyond the normal subcrop of Horton sediments on to one of the northeasterly striking granitic ridges. The basal Windsor surrounding the anticlinal ridges, and in the Ingonish-Dingwall area, consists of bedded micrite (mudstones) which are locally fossiliferous (packstones) and almost invariably characterized by various algal structures forming beds of small hemispheroids, oncolites or flat-pebble horizons deposited as a result of storm-reworking.

The facies distribution above leads to the obvious question: Are the western and eastern facies time-equivalent? A partial solution is provided by the interpretation of under- and overlying sediments.

Underlying sediments

It has been proposed that the Grantmire Formation and coarse red beds similar to the Grantmire Formation which underlie most of the eastern carbonate facies, were still being deposited during Early Windsor time. For this reason the Grantmire Formation and similar red beds were mapped as Early Windsor age (Bell and Goranson, 1938; Kelley, 1967; Weeks, 1954). The coarse, generally unsorted and subangular texture and the thickness of these red beds suggest considerable relief, ready availability of terrigenous debris, and high-energy conditions during deposition. A careful examination of the basal Windsor carbonates at virtually all recorded localities in Cape Breton Island failed to discover a single locality where this terrigenous debris mixed with the basal Windsor carbonates. Admittedly, the basal few centimetres contain locally sandy admixtures, a feature which is due to winnowing of finer matrix from the underlying conglomerates by the transgressing sea. As it is highly unlikely that both provenance and dispersal agents of the red beds

From: Report of Activities, Part A;
Geol. Surv. Can., Paper 77-1A (1977).

Table 86.1

Stratigraphic sections of the Windsor Group in Cape Breton Island

Lithologic Divisions (Faunal Zones)	Port Hood Island Mabou Mines	North Shore Skir Dhu	Lake Enon area	Point Edward
Upper Windsor (C to E, part of B)	algal, partly fossiliferous carbonates and fine clastics with minor gypsum beds (550 m)	algal, partly fossiliferous carbonates and fine clastics, minor gypsum beds in lower part	algal, partly fossiliferous carbonates and fine clastics with gypsum beds (230 m)	algal, partly fossiliferous carbonates and fine to coarse clastics (300 m)
Lower Windsor (A, part of B)	massive evaporite (~ 300 m)	massive evaporite (~ 300 m)	collapse breccia or evaporites (70 m)	
	collapse breccia	thin laminite breccia		
	stromatolitic laminite (10 m)	stromatolitic laminite (2 m+)	algal, fossiliferous carbonates or coquinoid carbonate (10 m+)	
underlying units	volcanic basement or Grantmire-like red beds	Grantmire red beds	intrusive basement or Grantmire red beds	Grantmire red beds

suddenly disappeared it must be assumed that the deposition of red beds had stopped long before the initial transgression of the Windsor sea. Evidence for the foregoing assumption is well documented along the eastern edge of the western laminite facies as at Irish Cove, Grand Narrows and the south end of St. Ann's Harbour where such mixing with terrigenous debris would have been most likely to occur. At these localities coarse red conglomerates are overlain by thin grey sandy patches which appear to fill shallow depressions on a consolidated surface. Both the sandy patches and the conglomerate are blanketed by the laminite. Apparently, the Cape Breton Island area had already reached a mature stage of denudation at the onset of Windsor deposition. Small inselbergs of basement rocks probably projected through rather flat-lying Horton sediments to the west, whereas the eastern region had risen slightly exposing basement rocks over large areas and exhuming old structural features. Even though the lack of terrigenous debris in the basal Windsor carbonates precludes continued deposition of Grantmire-like red beds into Early Windsor time, it does not prove that the western and eastern carbonate facies are time-equivalent.

Overlying sediments

The western stromatolitic laminite facies is generally overlain by a calcium sulphate body which is about 300 m thick (based on seismic and outcrop data). However, the eastern carbonate facies displays more complex relationships with respect to the overlying evaporites. Along the western edge as at Ingonish and Dingwall, the basal Windsor carbonates dip beneath massive evaporites of unknown thickness. At Lake Enon, along the southern part of the facies boundary, extensive drilling has penetrated up to 65 m of calcium sulphate above the basal fossiliferous carbonates whereas only a few kilometers away, the position of the evaporite is partially or entirely represented by a collapse breccia which contains considerable amounts of basement debris. Farther east at Point Edward and near Louisbourg no evaporites or collapse breccias occur. Instead, the sections there consist of relatively thin fossiliferous, but mainly algal carbonates and clastics of probably Late Windsor age correlative with similar carbonate-clastic sequences farther to the west (Table 86.1).

There is evidence, therefore, that the evaporite blanket extended beyond the major facies boundary probably wedging out against the gently rising land area to the east. At a hypothetical gradient of only 0.5° the 300-m-thick evaporite body would pinch out after only 35 km. The laminite facies and the western edge of the eastern fossiliferous facies were deposited prior to the evaporite blanket. Strong evidence in favor of this concept is a coquinoid carbonate southwest of St. Ann's Harbour that rests on basal laminite and which definitely underlies the massive evaporite.

It could be reasoned that the eastern fossiliferous facies postdates the western laminites accepting that both facies predate the evaporite body. Such an interpretation is difficult to defend on depositional grounds, because the fossiliferous carbonates should have spread completely across the underlying laminites. Their present restriction to the area east of the laminites, except for the occurrence near St. Ann's Harbour, is most likely due to salinity differences which controlled the sharp boundary between the two facies. The mere preservations of laminar stromatolites in the Phanerozoic requires abnormal marine environmental conditions. As blue-green algae are normally consumed by a variety of predators, mostly gastropods and polychaete worms (Garrett, 1970), a hypersaline environment would have kept such predators away. The water depth must have been shallow as stromatolites are built by blue-green algae, i.e. stromatolites form within the photic zone which extends down to approximately 40 m. The even laminations suggest constant subtidal conditions below the eroding capability of waves. Thus, the laminites were probably deposited in a wide, uniformly shallow and hypersaline sea grading eastward into a marine environment with seawater of normal salinity where the predatory activity of a normal marine fauna accounts for the non-preservation of blue-green algae. The normal salinities probably existed over a narrow zone roughly parallel to the eastern coastline where longshore currents provided a constant supply of normal seawater and nutrients. Slight

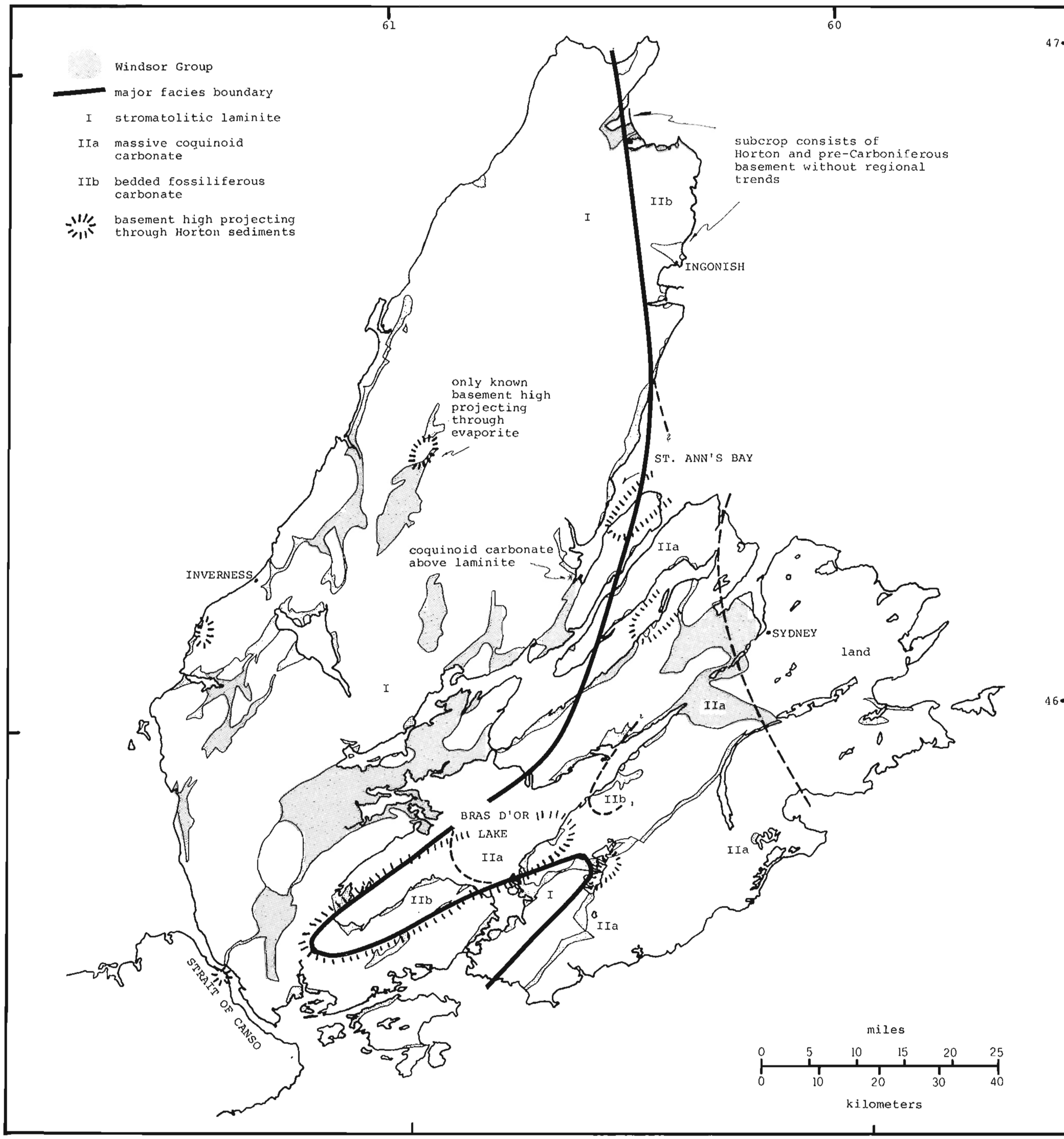

Figure 86.1. Facies distribution during Early Windsor time in Cape Breton Island, Nova Scotia.

deviations of the currents through time probably account for the slight westward spread of fossiliferous carbonates in the St. Ann's Harbour area.

The laminites of the western facies become vuggy and disrupted near the top reflecting a progressive shallowing of the subtidal conditions that prevailed throughout the major part of the laminite accumulation. Where the top of the laminite is exposed, a soft yellow marl with laminite fragments occurs. This zone generally measures only about a metre in thickness but north of Mabou Harbour it grades upsection into a thick carbonate breccia with a clayey or sandy matrix. At North Shore, along the northwestern shore of St. Ann's Bay, the contact with the overlying evaporites is exposed. Here about 30 cm of dark gray sandy clay with well rounded basement pebbles lies below the base of the evaporites and then grade downward into yellow marl with laminite fragments and finally into laminite. This facies change at the top of the laminite may be taken as evidence for a short period of emergence affecting the entire region prior to the deposition of the evaporites.

Evaporite Deposition

This brief uplift was followed by renewed subsidence which brought about a second transgression of the Windsor sea. In contrast to the earlier transgression the influx of seawater was now very restricted leading rapidly to high salinities and to the deposition of evaporites without the formation of stromatolitic laminites. The massive evaporites of Cape Breton Island consist almost entirely of calcium sulphates which probably represent marginal deposits of a much wider basin subsiding to the west where thick salt sections have been recorded, e.g. in central Nova Scotia, southern New Brunswick and in the St. Lawrence Bay region. The environmental setting of the evaporites in the Appalachian Region may be compared with the Upper Permian Zechstein of Central Europe. Richter-Bernburg (1960) was able to correlate varves in the Zechstein evaporites over considerable distances and determined the actual time span of deposition. Marginal evaporites similar to the Lower Windsor evaporites of Cape Breton Island were shown to accumulate at a rate of 0.5 to 4 mm/year. At a rate of 2 mm/year 300 m of calcium sulphates would have accumulated in 150 000 years. The occurrences of oolitic bands in the Lower Windsor evaporites at various levels indicates that the evaporitic deposition was repeatedly interrupted. However, even if the time of 150 000 years is doubled or tripled the time interval remains relatively short. For this reason, fossil assemblages above and below the evaporite body are probably not likely to vary significantly. The coquinoid carbonate below the evaporite near St. Ann's Harbour is dominated by the brachiopod *Cranaena tumida* Bell, a form typical for the subzone B above the evaporite.

Algal-dominated carbonates and the influx of fine clastics mark the end of evaporite deposition. Fossil assemblages assign these carbonates to B to E subzones of Bell. The algal character of these mostly Upper Windsor carbonates remains dominant throughout Cape Breton Island, and the earlier facies boundary disappears. A more detailed analysis of these Upper Windsor sediments must await the evaluation of further data.

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PRELIMINARY REPORT ON THE PETROLOGY OF THE LEVEL MOUNTAIN VOLCANIC CENTRE, NORTHWEST BRITISH COLUMBIA

E.M.R. Research Agreement 1135-D13-4-138/76

T.S. Hamilton[1] and C.M. Scarfe[1]
Regional and Economic Geology Division

Introduction

Level Mountain is a shield volcano which outcrops over an area of about 700 square miles in northeastern British Columbia (131°20'W; 58°31'N). The volcanic plateau, with an average elevation of about 4500 feet, has a central region of peaks and ridges rising to a maximum elevation of 7200 feet at Meszah Peak.

Reconnaissance mapping by Gabrielse and Souther (1962) indicates a sequence of lavas of Late Tertiary to Recent age. These lavas belong to the alkalic series and range in composition from ankaramite to comendite (Hamilton and Scarfe, 1976). Souther (1966) and Souther and Symons (1974) have described similar alkalic lavas on Mt. Edziza, a volcano about 60 miles southeast of Level Mountain. Late Cenozoic volcanism in this part of British Columbia may be associated with deep crustal rifting (Souther, 1970).

This report covers work completed during the 1975 and 1976 field seasons, a total of 11 weeks, together with supporting petrographic and chemical analyses.

Volcanic Stratigraphy

The geological sketch map (Fig. 87.1) shows the distribution of the major lava types on Level Mountain: alkali basalt, trachyte, comendite and pantellerite. The alkali basalts are volumetrically dominant and account for most of the plateau thickness. The gently rolling uplands and glacially dissected ridges exhibit a varied stratigraphy dominated by trachybasalt and peralkaline trachyte. Peralkaline comendites and pantellerites are restricted to the central peaks region.

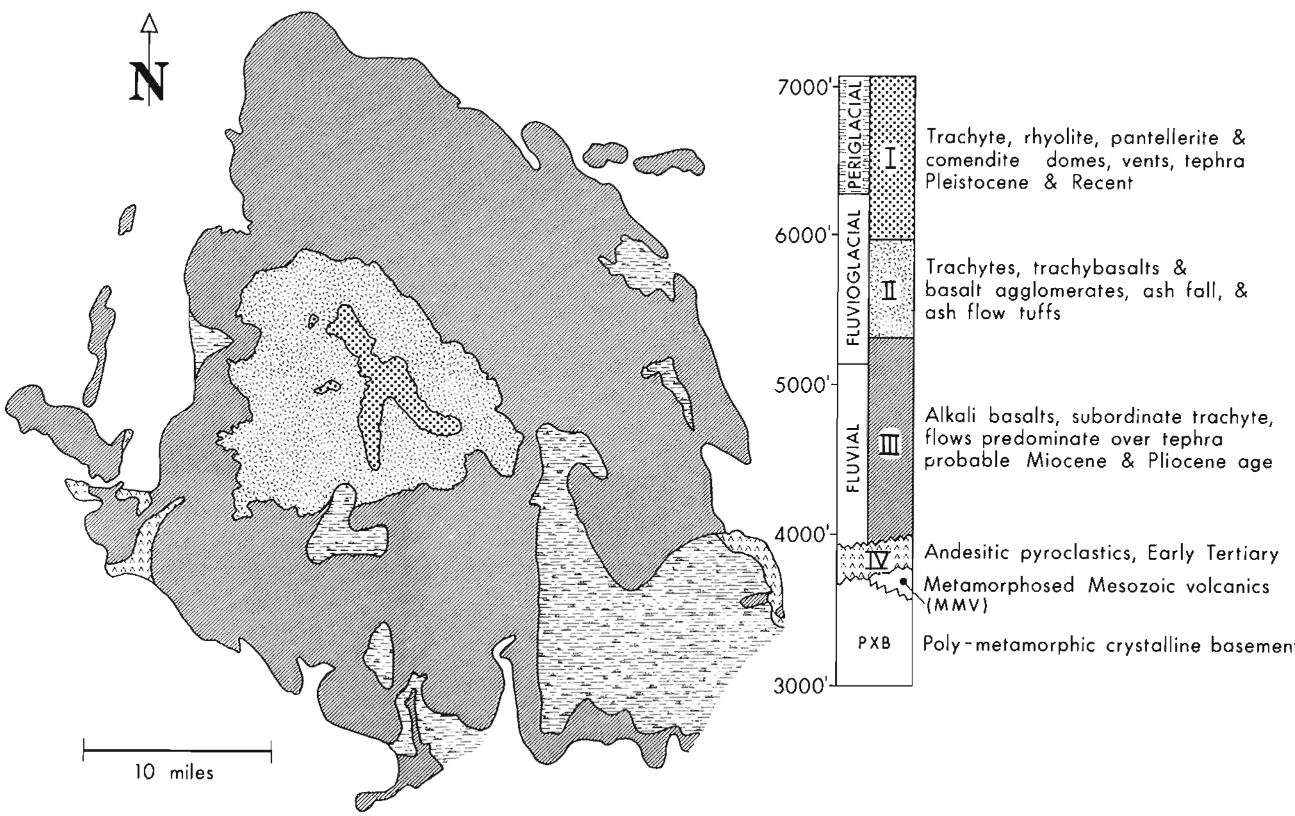

Figure 87.1. Sketch map of the geology of the Level Mountain volcanics, northwestern British Columbia. Age increases from Unit I to Unit IV.

[1]Department of Geology, University of Alberta, Edmonton, T6G 2E3

Table 87.1

Petrography and modes of selected Level Mountain rocks

Rock Type Map Unit/Sample #	Ankaramite III/h	Alkali-Basalt III/s3	Tristanite II/LMI20c	Phonolite I/LMI20i	Trachyte II/24/1hi	Pantellerite I/25/5e
Phenocrysts	Olivine Fo_{82}-5% Orthopyroxene En_{87}-3% Plagioclase An_{70}-7%	Olivine Fo_{95}-10% Orthopyroxene En_{100}-10%	Aegerine augite-10% Plagioclase zones-15% core An_{33} rim An_{37} Microcline-5% (xenocryst)	Aenigmatite-10%	Aegerine augite-3% Anorthoclase-6%	Aegerine augite-1% Aenigmatite-2% Anorthoclase+sanidine-12%
Groundmass	Titanaugite-25% Plagioclase An_{62}-40% Titanomagnetite-5% Glass-15%	Augite-5% Plagioclase An_{56}-55% Magnetite-10% Glass-5% Zeolite-5%	Aegerine augite-10% Plagioclase An_{70}-25% Sanidine-15% Magnetite-10% Calcite+zeolite-10%	Plagioclase An_{17}-40% Sanidine-35% Nepheline-5% Rutile-5% Glass-5%	Aegerine augite-10% Plagioclase-tr Sanidine-60% Magnetite+pyrite-10% Glass-10%	Aegerine augite-20% Aenigmatite-5% Sanidine-40% Magnetite+hematite-3% Quartz-15% Glass-2%
Textures	Ophitic with glomeroporphyritic phenocrysts	Vesicular Isolated phenocrysts in panidiomorphic to felty mesostasis with intersertal glass and zeolites	Poikilitic xenocrysts seriate phenocrysts protoclastic mesostasis intersertal alteration	Pilotaxitic	Glomeroporphyritic (aegerine augite + anorthoclase+pyrite) in pilotaxitic mesostasis	Glomeroporphyritic phenocrysts in felty to orthophyric mesostasis

Map Unit IV – Basement

The Level Mountain plateau basalts rest on a Tertiary erosional surface with a relief of several hundreds of feet. The pre-Tertiary basement rocks are polyphase metapelites and acid plutonic rocks in the west, and meta-argillites and greenstones in the south.

In the southwest, along Egnell Creek, approximately 300 feet of intermediate to acid flows and agglomerates may be equivalent to the early Tertiary Sloko Group described elsewhere in northern British Columbia (Souther, 1970). These rocks which have been faulted, folded and subjected to zeolite facies metamorphism, are separated from the Late Cenozoic Level Mountain lavas by an angular unconformity.

Map Unit III – Plateau

The plateau basalts are divided into four sub-horizontal units (Fig. 87.2). The lowest unit (175 feet thick) is composed of thin columnar flows of alkali olivine basalt and altered grey-green vesicular basalts. Two dark green ankaramite flows (25-30 feet thick) provide good marker horizons. The next unit (350 feet thick) has up to seven columnar cooling units (25 feet thick) of alkali olivine basalt separated by buff-weathered vesicular flows. The third plateau unit (250 feet thick) is mainly ankaramite occurring as spheroidally-weathered massive flows. The highest unit (400 feet thick) is represented by eight to ten columnar cooling units of alkali basalt, frequently exhibiting plagioclase megacrysts and plagioclase in hyalopilitic clots. Inclusions of glacial erratics, thin lacustrine silts, pillow basalts and palagonite tuff-breccia date this uppermost unit as synglacial.

Map Units II and I – Central Peaks

The stratigraphy of the glacially-dissected central region is more varied and strongly influenced by local vents and eruptive centres. Mapping indicates that the headwaters of Kakuchuya Creek were the site of a high stratocone, now largely eroded away. The north-east flank of the cone is an alternating succession of alkali olivine basalt and ankaramite with one pantellerite and two peralkaline trachyte horizons. The peralkaline flows account for up to 40 per cent of the section and their thick wedge-shaped nature is responsible for the 26° initial dips in the upper part of the succession. On Meszah Peak (Fig. 87.3) the sequence is alkali basalt followed by peralkaline trachyte and comendite and finally by alkali olivine basalt. Other sections to the east of Meszah Peak contain considerable volumes of trachybasalt, tristanite and phonolite.

Two till horizons are present in most of the central ridges. Erratics and ice-erosional features are also found at the highest levels, including the summit of Meszah Peak. Thus most of the Level Mountain volcanics are older than the last continental glaciation. However, several minor basaltic vents, scattered throughout the central region, are of postglacial age. Undisturbed fusiform bombs and lapilli litter the slopes up to 100 feet from these vents, indicating the rather feeble nature of this latest activity.

Petrography and Chemistry of the Principal Rock Types

The petrography of selected samples is summarized in Table 87.1 while the corresponding chemical analyses and norms are presented in Table 87.2. The mineral compositions in Table 87.1 are based on optical data. Whole rock chemical analyses were obtained by X-ray fluorescence (Holland and Brindle, 1966) and by microprobe energy dispersive analyses of glasses using the method of Schimann and Smith (in prep).

Ankaramites occur as dark-coloured massive flows with spheroidal weathering and frequent columnar jointing. These are the least glassy (generally <10%) of the mafic rocks. Texturally there are two distinct types with minor differences in mineralogy. Most of the ankaramites of the plateau margins have low phenocryst contents and an ophitic or subophitic texture. In sample h, titanaugites are purplish brown and zoned to higher Ti contents at crystal margins. Magnetite and titanomagnetite occur as euhedral to skeletal crystals in the glassy mesostasis. The other textural variety has a high phenocryst content with fragmented euhedral olivine, orthopyroxene, clinopyroxene and sometimes plagioclase in a randomly-oriented felty groundmass of augite, plagioclase, titanomagnetite and glass. This latter type is found in spacial association with pantellerites and peralkaline trachytes.

Alkali basalts occur as scoria, vesicular flows (Table 87.1, S3), and massive columnar jointed flows and dykes (Table 87.2, KD-1). Trachytic textures are common in many flows particularly in regions of high phenocryst content. Phenocrysts are fragmented and show reaction textures with the groundmass, particularly in grassy flows. Some types contain three generations of plagioclase which vary markedly in crystal size and chemistry. Megacrysts of calcic labradorite or bytownite, up to 2-3 inches in length, are common in dykes and in

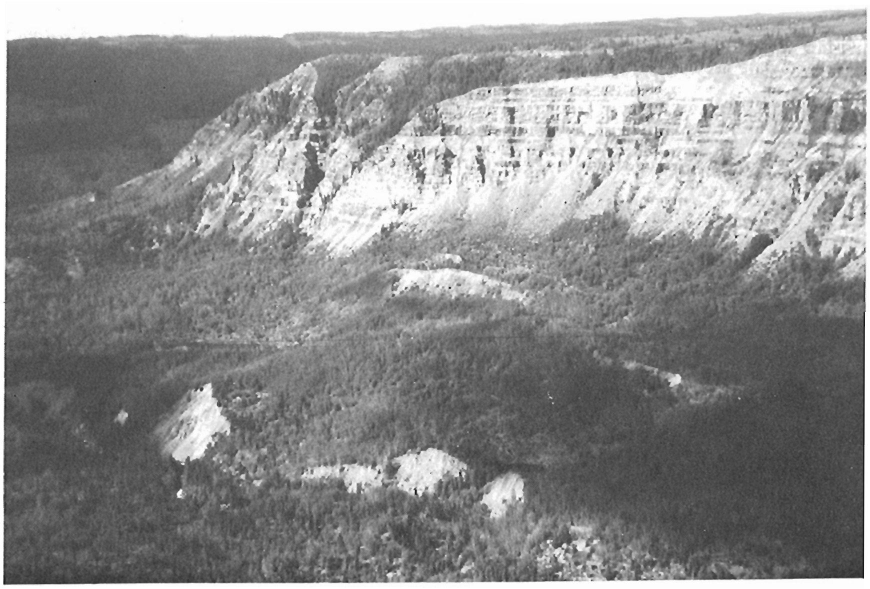

Figure 87.2.

The stratigraphy of map unit III as exposed in the canyon of the Little Tahltan River on the southern plateau margin. An 1100-foot section shows an alternating sequence of ankaramite and alkali basalt flows.

Figure 87.3.

Meszah Peak west view. Dark unit in centre foreground is alkali basalt. Trachyte flows and agglomerates are exposed as a series of benches at the base of Meszah Peak. Pantellerite, comendite pitchstones and peralkaline welded tuffs from the cliff escarpment and light coloured high slopes. Thin flows of alkali olivine basalt capping the peak are under snow cover.

Table 87.2

Chemistry and CIPW Norms for selected Level Mountain rocks

Sample #	Ankaramite h*	Alkali-basalt KD-1*	Tristanite LM120c	Phonolite LM120i	Trachyte 24/1hi*	Pantellerite 25/5e*
SiO_2	44.5	48.4	54.2	55.1	64.1	67.0
Al_2O_3	14.0	15.5	15.9	15.5	17.5	13.4
Na_2O	2.0	3.4	5.0	5.0	6.1	5.9
K_2O	0.9	0.4	3.2	4.4	5.3	4.9
FeO**	12.0	12.2	7.9	9.2	4.5	7.0
CaO	7.9	10.1	5.0	5.1	1.0	0.3
MgO	11.5	6.0	1.9	1.9	0.2	0.1
TiO_2	1.8	2.2	1.4	1.5	0.4	0.5
MnO	0.2	0.2	0.1	0.1	0.1	0.2
P_2O_5	0.4	0.3	0.4	-	0.1	-
SO_2	0.1	0.1	BaO 0.4	-	-	-
H_2O	4.3	0.7	3.9	0.2	0.2	0.3
Total	99.6	99.5	99.3	100.0	99.6	99.6
CIPW Norms						
Q	-	-	-	-	5.9	16.3
Or	5.7	2.2	19.9	26.8	31.5	27.2
Ab	17.2	29.4	44.5	37.8	52.1	43.7
An	28.0	26.1	12.1	7.2	4.4	-
Ne	-	-	-	3.0	-	-
Cpx	8.3	18.4	9.6	15.7	0.1	0.4
Opx	16.1	5.6	0.9	-	0.4	Wo 0.4
Ol	13.5	9.2	9.1	2.2	-	Ac 5.5
(Il + Mt)	10.2	8.4	2.9	7.3	5.5	6.5
AI	0.3	0.39	0.74	0.74	0.91	1.17

Microprobe EDA and H_2O (Penfield) analyses by T. S. Hamilton;

* XRF by J. G. Holland. Norm calculation and classification after Irvine and Baragar (1971).

Agpaitic index, $AI = \frac{Na + K}{Al}$ (mol)

** Total iron reported as FeO.

the uppermost flows of the plateau. Calcic andesine or labradorite crystals often occur as cumulophyric aggregates together with black glass. Microphenocryst or groundmass plagioclase may exhibit reverse zonation. Euhedral magnetite and glass are ubiquitous. In the older alkali basalts the groundmass is altered to chlorite, goethite, calcite and zeolites. With slight variation in the proportions of the ferromagnesian minerals to plagioclase these rocks are gradational to hawaiites.

Trachybasalts show the most variable mineralogy and chemistry of any rock type. Two types, fragmental flow agglomerates and phenocryst-rich flows, may be distinguished. Chemically and mineralogically these rocks are gradational with alkali basalts and tristanites. There is a silica gap from about 52-64 per cent SiO_2 between trachybasalts and trachytes. The agglomeratic type has up to 30 per cent glass and may contain phenocrysts or xenocrysts of olivine, orthopyroxene, clinopyroxene and magnetite. The phenocrysts in trachybasalt are always in reaction with the groundmass. Both agglomerates and flows contain phenocrysts of andesine and frequently have glomeroporphyritic aggregates of anorthoclase, pyrite and aegerine augite or sodic ferro-hedenbergite. In trachybasalt flows phenocrysts of aenigmatite are sometimes present. The groundmass of these trachybasalts is characteristically composed of flow-oriented laths of oligoclase, aegerine augite and sandine. The distinction between trachybasalt and tristanite flows is sometimes difficult to make without chemical analyses. Sample LMI20c has been presented in Tables 87.1 and 87.2 as a transitional type.

Phonolites of two textural varieties occur in successions dominated by trachyte. Whether the phonolites show vesicular pumiceous character (yellow, white, buff colours) or trachytic texture (generally grey) they are relatively aphyric. Sample LMI20i is representative of the pumiceous type.

Trachytes and peralkaline trachytes are a dominant rock type in the central ridges. Sample 24/1hi is a low phenocryst variety. The groundmass usually shows trachytic texture of lathlike sanidine, sodic ferrohedenbergite, magnetite and glass. Sometimes the groundmass ferromagnesian mineral is a sodic amphibole or aenigmatite; in which case magnetite is absent. Chemically these grade into pantellerites. Occasionally dense inclusions of coarsely crystalline clinopyroxene occur as nodules in trachyte flows.

Pantellerites and the more silica-rich iron-deficient comendites have a wide variety of eruptive styles, textures and mineralogies. Sample 25/5e comes from high in the succession on the southwest flank of Kakuchuya Cone. Phenocryst-bearing pantellerites generally contain anorthoclase and aegerine augite. Comendites contain only anorthoclase and sanidine phenocrysts. The groundmass invariably contains sanidine, quartz and glass with one or more sodic ferromagnesian minerals. Comendites contain riebeckite, aenigmatite or arfvedsonite. Pantellerites may contain all three plus astrophyllite. These unusual sodic minerals generally poikilitic include oriented laths of sanidine, giving the rock a eutaxitic texture. The peralkaline salic rocks occur as flows, pitchstones, welded tuffs, dykes, laccoliths and hypabyssal plugs. Comendite flows appear to have been fluid, some having lava tubes several feet in diameter. The thickness (>100 feet) and great lateral extent of some flows and their frequency in the stratigraphy of the central region attest to their importance as a magma type. Rhyolites and dacites, frequently hydrothermally or deuterically altered, are also present as flows and domes but are volumetrically insignificant compared to their peralkaline counterparts.

The strongly alkalic chemistry of the Level Mountain lavas can be seen in Table 87.2. With increasing silica content there is a corresponding increase in alkalies and decrease in calcium. All of the rocks show a high level of TiO_2 and correspondingly have titaniferous minerals in the mode (Table 87.1). The peralkaline rocks, represented by sample 25/5e, have acmite in the norm and an agpaitic index greater than 1. Acmite in the norm corresponds to the presence of alkali pyroxenes, alkali amphiboles, aenigmatite or other alkali ferromagnesian minerals in the mode. Where alkali ferromagnesian minerals are present but the agpaitic index is less than 1, the rocks are considered to have a peralkaline affinity (MacDonald, 1974).

Interpretation

In summarizing the petrochemistry of the Level Mountain volcanics the following points should be emphasized:

(1) The dominant magma type of the Level Mountain shield is alkali basalt;

(2) Ankaramite, alkali basalt, hawaiite, Na-rich trachybasalt and trachyte belong to the sodic alkali basalt series;

(3) Eruptive products include both silica-saturated and silica-deficient types;

(4) Peralkaline lavas are important in the central region, particularly in the upper part of the succession.

Firm conclusions concerning the petrogenesis would be premature at this stage; however, it appears that derivation of the suite via simple differentiation of a basaltic parent is unlikely. Any model must account for large volumes of two essentially different magma types: alkali basalt and peralkaline trachyte or pantellerite. The large volumes of alkali basalt and the local tectonic setting indicate that this magma type is mantle-derived. Peralkaline trachyte may represent a second primary magma type. The other rock types which are present in smaller volumes are probably explained by a variety of petrogenetic processes to be discussed at a later time.

Acknowledgments

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GEOLOGY AND MINERALIZATION IN THE DOLORES CREEK AREA, BONNET PLUME RANGE, YUKON

Peter Laznicka[1]
Regional and Economic Geology Division

Introduction

The Proterozoic core of the Ogilvie and northwestern Wernecke mountains contains scattered copper ± cobalt, and radioactive mineralization traceable over a distance of about 400 km in an east-west direction. Most showings are discontinuous, crosscutting veins of siderite or quartz-chalcopyrite, but conformable to semi-conformable deposits in sediments and a 'porphyry-like' mineralization in feldspathic metasomatites also occur. Their regional controls are largely unknown.

Two rocks types, however, seem to be repeatedly spatially associated with many of the occurrences particularly in the eastern portion of the belt: 1) diorite dykes and sills; 2) siliceous or ferruginous carbonate paraconglomerates to breccias. Both these rocks and abundant copper showings are present in a relatively small (20 x 10 km) area near the headwaters of Dolores Creek (a tributary of the middle reaches of Bonnet Plume river, and around Glacier Lake (Findlay 1969) (Fig. 88.1). This small area, therefore, was selected as a possible key for understanding the metallogeny of the broader region. The writer, ably assisted by Robin Edwards and Craig Chappell and sponsored by a grant from the Department of Indian Affairs and Northern Development as well as a generous logistic support received from the Geological Survey of Canada, devoted one month to field work in the area.

This paper is a preliminary report based mainly on field observations with a limited amount of thin section and analytical work. The two most fundamental problems – the nature of the various conglomerates and breccias, and the effects of metamorphism and metasomatism of rocks and mineralization in the area – are the subjects of M. Sc thesis work of Messrs. Edwards and Chappell, respectively, and are expected to be completed by mid-1978.

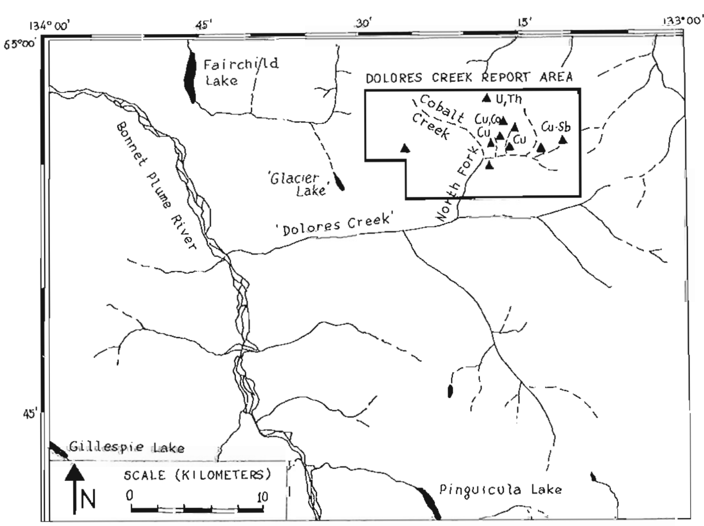

Figure 88.1. A sketch map showing the Dolores Creek report area (NTS 106 C)

Geological setting and stratigraphy

The research area is part of a broad northwest trending Proterozoic basement uplift located on both sides of Bonnet Plume River and surrounded by sediments of the Paleozoic shelf and continental margin association. The Helikian ? (Hcs; Hsc; Hc) and Hadrynian ? (HSFe; Hsq; Hc; and Hsc) map units distinguished in the broader area by Blusson (1974) are present in the Dolores Creek-Glacier Lake area.

The lowermost unit (Hcs) (Blusson, 1974) present at Dolores Creek, has a local thickness of almost 2000 m. It starts with a thick succession of fine grained quartzite (fine metaquartzarenite and quartz-siltite) of variable carbonate content, interbedded and intergraded with subordinate clastic quartz-rich carbonates (dolomites and rare limestones) and argillaceous siltites. Most of these sediments are metamorphosed to the lower to middle greenschist facies, but the metamorphic grade is not obvious in the quartzites. The argillaceous and carbonate-rich members have been altered to quartz-albite-muscovite-chlorite and quartz-albite-biotite ± epidote, tremolite and sphene schists and horfelses.

Near the top of unit Hcs (Blusson, 1974) isolated thin beds of bright red to purple ferruginous and commonly carbonate-rich chert (jasper and siliceous ankeritite) occur. The surrounding metasiltites, with characteristic purple or light pink (= faded purple) shades, generally grade into breccias and conglomerates. The fine grained rocks composed essentially of quartz that occur in and near the zone of alteration and mineralization may be genetically more complex so they are referred to by a nongenetic term silicite. Diorite dykes and copper showings are most abundant in this milieu.

A thick, rather monotonous sequence of alternating grey argillites to phyllites, quartzites and rare ferruginous dolomites (unit Hsc; Blusson, 1974) appears to overlie unit Hcs conformably, although the contact in many places is a fault. In the most intensively disturbed area around Cobalt cirque, distinct green to yellowish green 'phyllites of Cobalt cirque' appear to be hydrothermally altered (bleached) dark grey equivalents of the basal portion of unit Hsc. Low angle, small scale cross-lamination and rare mudcrack casts suggest a shallow water depositional environment.

The distinctive unit Hc (light grey, yellow and orange-weathering dolomites) conformably overlies unit Hsc and reaches a thickness of about 1000 m in the study area. Nodular limestones and argillaceous

[1]University of Manitoba, Winnipeg 19, Manitoba.

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dolomites are interbedded locally. Stromatolites, algal mats, intraformational and sharpstone conglomerates and molar tooth structures are common and indicate shallow water origin.

Unit Hc may represent the top of the Helikian sequence. In the Dolores Creek area it is overlain by Hadrynian sequence. The sub-Hadrynian erosional unconformity is marked locally (outside of the area of study) by a spectacular boulder conglomerate. The base of Hadrynian (unit HSFe; Blusson, 1974) is composed of brownish red to maroon, thin bedded, laminated argillites with rare dolomite and limestone interbeds. The thickness ranges from 0 to 150 m. Patchy, light green, reduced sections appear locally. The stratigraphically higher, thick, interbedded sequence of shales, quartzites and impure carbonates (units Hc, Hsc, Hsq) is developed largely outside the study area.

There is no clear evidence of Proterozoic volcanism synchronous with sedimentation in the Dolores Creek-Glacier Lake area, although the ferruginous cherts and carbonates near the top of unit Hcs may be the result of an exhalitive contribution. All the volcanic-looking fragmental rocks ('agglomerates', 'lappili tuffs', etc. mentioned in some unpublished mining company reports) can be accounted for by metasomatism of carbonate-rich silicites and derived breccia-conglomerates.

Altered and unaltered diorite dykes (the only Proterozoic? intrusions identified) of irregular form

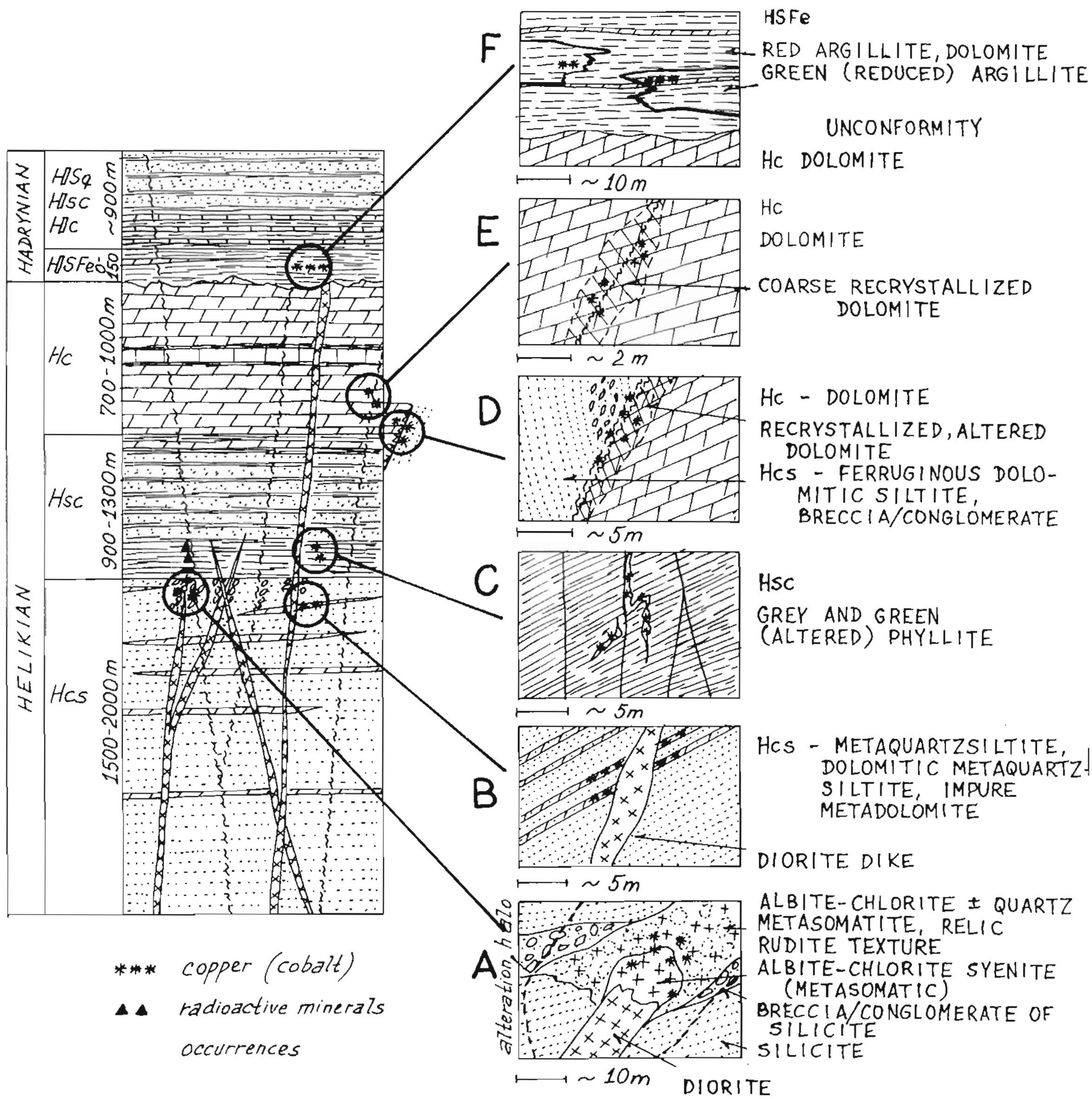

Figure 88.2. Schematic, restored stratigraphic column, Dolores Creek area, showing the setting and generalized diagrams of the main mineralization types. Types A to F described in Table 88.1.

and ranging from 1 to 20 m wide occur in the area. These dykes have not been dated; there may be several generations of them. The most likely interval of emplacement of the bulk of dykes is, according to field observations, between the time of deposition of the upper part of unit Hcs and the end of the sub-Hadrynian hiatus. Spatially associated patchy albite-chlorite metasomatites (syenites) occur with some diorites.

Structurally, a resistant core of unit Hcs metasediments is centred in its southwestern part of the small map-area. The core is flanked by younger units that generally dip outwards; i.e. to the west and northwest on the northwestern flank, and to the east and southeast on the southeastern flank. Higher metamorphic grade, spectacular vertical fracture cleavage, steep dips, and locally tight folds near the centre of this core contrast sharply with a lower degree of metamorphism, gentle dips, and generally open folds in the flanking units. The core-flank contact over most of the area is marked by steeply dipping faults.

The siliceous conglomerates and breccias near the top of unit Hcs and their metasomatism

The peculiar pink or purple siliceous and carbonate-rich conglomerates and breccias and similar rocks marked by abundant specularite have regional though patchy distribution in the Proterozoic core of central Yukon and commonly host copper mineralization (Dolores Creek, Bear River area, Dempster Highway, near Mt. Harper, etc.). For this reason considerable effort has been devoted to their study in which could be their type area.

In the Dolores Creek area these rocks are hosted by and developed from dolomitic ferruginous silicites (metaquartzsiltstones ?) in unit Hcs usually in terranes where faults and diorite dykes are abundant. Clast size, clast angularity, and clast/matrix ratio vary considerably between the two textural end members which are 1) brecciated silicite and 2) paraconglomerate consisting of well rounded silicite clasts floating in quartz-dolomite matrix. The clast composition of over 99% of studied rudites is silicite, same as in their undisturbed host. The only other type of clasts, found once, is diorite of the same type as in nearby dykes.

The rudite lithosomes may be conformable with the depositional fabric (bedding, lamination) of their sedimentary host; they may fill channels, wedges, syngenetic/diagenetic intraformational diapirs and clastic dykes cutting across the host's fabric at all angles, or they can fill clearly postdepositional, fault-bounded spaces. The shape and size of breccia/conglomerate lithosomes vary considerably. Usually they are elongated, commonly parallel to a fault. In several cases these lithosomes were observed as the strike extention of a pinched-out (dissipated ?) diorite dyke. Their regional distribution is discontinuous and individual outcrops cover tens to thousands of square metres.

Table 88.1

Brief characteristics of the six principal mineralization types shown in Figure 88.2

Type A

Erratically disseminated chalcopyrite ± specularite, pyrite in albite-chlorite ± biotite, muscovite metasomatites replacing conglomerates and breccias of ferruginous dolomitic silicite near diorite dyke contact. "Porphyry copper-like mineralization"; 8 occurrences.

Type B

Finely disseminated and veinlet chalcopyrite replaces metacarbonate horizons in the vicinity of diorite dykes or faults; 16 occurrences.

Type C

Irregular and discontinuous quartz-siderite veins contain scattered grains and masses of chalcopyrite, cobaltite, gersodorffite ?, pyrite and pyrrhotite; 4 occurrences.

Type D

Irregular, discontinuous quartz veins in fault gouge in Hcs sediments and coarse, recrystallized Hc dolomite contain erratically scattered tetrahedrite, pyrite and chalcopyrite; 4 occurrences.

Type E

Erratically distributed masses of chalcopyrite ± bornite and, exceptionally, galena and sphalerite replaces zones of coarse recrystallized and altered dolomite along faults; 8 occurrences.

Type F

Chalcopyrite and chalcocite microconcretions and trains of small grains occur parallel with lamination in light green dolomitic argillites and argillaceous dolomites, near the facies transition into oxidation (maroon) equivalents. "Copper Shale" type; 3 occurrences.

The study of sedimentary stuctures reveals that these rocks are products of long-lasting, continuous tectonic instability which had been affecting the sediments during and after lithification. The earliest 'conglomerates' consist of angular to rounded clasts resing in laminated, plastically deformed silicites. The rock strongly resembles laminated siltstones with rafted dropstones as known from Gowganda and Conception Groups in eastern Canada.

Some intraformational conglomerates/breccias in silicites resemble diamictites and give an impression of having been transported, whereas others were closely generated *in situ* along early faults affecting a semi-plastic sediment. The majority of siliceous rudites, however, are a product of fragmentation of a brittle consolidated silicite *in situ* or very near the source.

All the rudites showing soft sediment deformation are premetamorphic. The superimposed metamorphism and metasomatism may have altered both the original depositional fabric and mineralogy, depending on intensity. Because the alteration is most intense near the diorite dykes where a crude zonality can be recognized, it is suspected that hydrothermal metasomatism rather than a simple load metamorphism has been the principal agent of alteration.

The least intense alteration can be seen in rudites with the lowest carbonate content. The previously purple rocks are as a rule discoloured (pink) and the formerly dispersed hematite is accumulated into scattered bladed crystals of specularite located mostly in the matrix. The original rudaceous texture is fully preserved. Carbonate-rich rudites in the outer alteration halo (Fig. 88. 2A) contain abundant recrystallized ankerite in the matrix, and rarely 1-5 mm, large, sharply developed ankerite idioblasts (rhombs) scattered in the pink silicite clasts. The resulting rock strongly resembles a tuff with feldspar porphyry clasts.

As one approaches a diorite dyke, abundant green chlorite ± specularite and carbonate replace the matrix and form scattered prophyroblasts in clasts of carbonate-rich siliceous rudite. The innermost alteration zone (Fig. 88. 2A) contains an assemblage of chlorite-albite-quartz ± epidote, tremolite and specularite which progressively replaces 1) matrix, and 2) clasts. The original rudaceous texture is well recognizable in the outer and intermediate alteration zones, whereas in the inner zone it is fully obliterated, and end product being a pink to greenish grey albite-chlorite syenite with granitic texture. All the zones are irrugular and have gradational boundaries. A small quantity of chalcopyrite occurs disseminated in certain parts of the albite syenite and in adjaect albite-chlorite-quartz metasomatite with nebulous relics of the original rudaceous texture. The highest copper grade seems to coincide with sites of potassium enrichment marked by the presence of accessory biotite and muscovite (K-metasomatism ?). The resulting "porphyry-copper like" mineralization has been the subject of exploration in several prospects.

The feldspathic metasomatites form several irregular bodies with gradational boundaries against their bedded host. In the Dolores Creek area the host is always the silicite conglomerate/breccia, and a diorite dyke always occurs in the vicinity so the metasomatizing fluids are likely products of the same magma chamber that generated the diorite. The ascent of fluids, however, postdates the diorite crystallization because the diorite itself is considerably fractured, altered, and in places (Cobalt cirque) laces with albite-chlorite-quartz veinlets often carrying chalcopyrite. The most likely genetic interpretation of the development of rudites and metasomatites is as follows: The initial tectonic instability at the time of deposition of the upper part of unit Hcs caused the soft sediment deformation within the sedimentary basin and generated fractures that at depth were gradually being filled by diorite dykes. The deformation continued during and after the sediment lithification. Many ascending diorite dykes abruptly terminated upon reaching the porous, water-saturated patches of carbonate-rich siliceous rudites. This could trigger the generation, expansion and release of gases, possibly at explosive rates, that caused additional fragmentation of sediments in a series of small, linear diatremes and which initiated the flow of sodium and minor potassium-carrying hydrothermal solutions that altered diorites and that withdraw trace copper and possibly cobalt, uranium and thorium. The fluids were subsequently diluted and discharged into the porous rudites causing alteration and local redeposition of copper.

Mineralization

About 40 copper showings some with minor amounts of cobalt in cobaltite and gersdorffite ? and antimony with silver in tetrahedrite were recorded. The six most widespread mineralization types and their setting in the local stratigraphic column are illustrated in Figure 88. 2 and described in Table 88. 1. The mineralization is probably a product of two independent mineralization epochs: 1) the earlier ? epoch represented by epigenetic mineralization in Helikian rocks (ore types A-E, Fig. 88. 2) whose age is most likely late Helikian to early Hadrynian although some minor adjustments may have taken place as late as during the Mesozoic orogeny, and 2) the Hadrynian epoch during which stratiform diagenetic mineralization formed in green shales.

The first epoch generated a polytypic mineralization probably in several pulses. It seems that the emplacement of diorite dykes and their subsequent autometasomatism acted both as a metal source and driving force (by providing and generating heat, hydrothermal solutions and space) of mineralization. A crude zonality of ore types seems to be developed in respect to the focus (the site of diorite dyke dissipation/interaction with water saturated porous sediments), from the more proximal to the more distal: types A, B, C, D and E.

No evidence has been found in Helikian sediments of syngenetic-diagenetic mineralization or trace metal enrichment except for pyrite, sometimes framboidal and resembling "mineralized bacteria". The hairthin trains of chalcopyrite and cobaltite following the planes of bedding schistosity in the "phyllites of Cobalt cirque" (associated with type C – Fig. 88. 2) occur in transversal fractures as well, and diminish outwards from tectonically distrubed and diorite-intruded zones.

The radioactive mineralization has not been studied in detail. It could have developed by withdrawal of U and Th from accessory minerals in diorite (radioactive zircon, monazite, allanite) and subsequent redeposition in a relatively distal zone.

The second, Hadrynian mineralization epoch generated an erratic copper enrichment and mineralization confined to the light-green, reduced patches in sediments of the basal Hadrynian unit HSFe. Scattered grains and microconcretions of chalcocite and chalcopyrite surrounded by secondary malachite occur parallel with the lamination in dolomitic argillites and argillaceous dolomites and are most common near the "green"/"red" facies transition. This ore type is a mirror image of the world-wide "copper shale" model (Bogdanov *et. al.* 1973) and its presence was predicted before the actual mineralization was found. The erosional source of copper in this model is usually distal, located outside the study area, although some of the copper may well have been derived from the underlying Helikian rocks.

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PHYSICAL VOLCANOLOGY OF AN ARCHEAN VENT COMPLEX, MINNITAKI LAKE AREA, NORTHWESTERN ONTARIO

E.M.R. Research Agreement 1135-D13-4-123/76

Richard O. Page[1] and Paul M. Clifford[1]
Regional and Economic Geology Division

Introduction

During the last three field seasons, a detailed investigation has been carried out on a 6000 to 7000-metre (18 000-21 000-foot) thick volcanic pile near Sioux Lookout, Ontario (Fig. 89.1). The pile has been informally termed the "Central Volcanics" (Turner and Walker, 1973), but a more precise stratigraphic subdivision and nomenclature is currently being devised by one of us (Page). The study was undertaken to provide a more accurate picture of Archean volcanism, first by documentation of vertical and lateral facies changes within the extrusives, and secondly by formulating a generalized model of Archean basin evolution by comparison with modern analogs. Previous work of this type in the Minnitaki area has dealt with the two groups of sedimentary rocks which bound the "Central Volcanics" strata (Walker and Pettijohn, 1971; Turner and Walker, 1973).

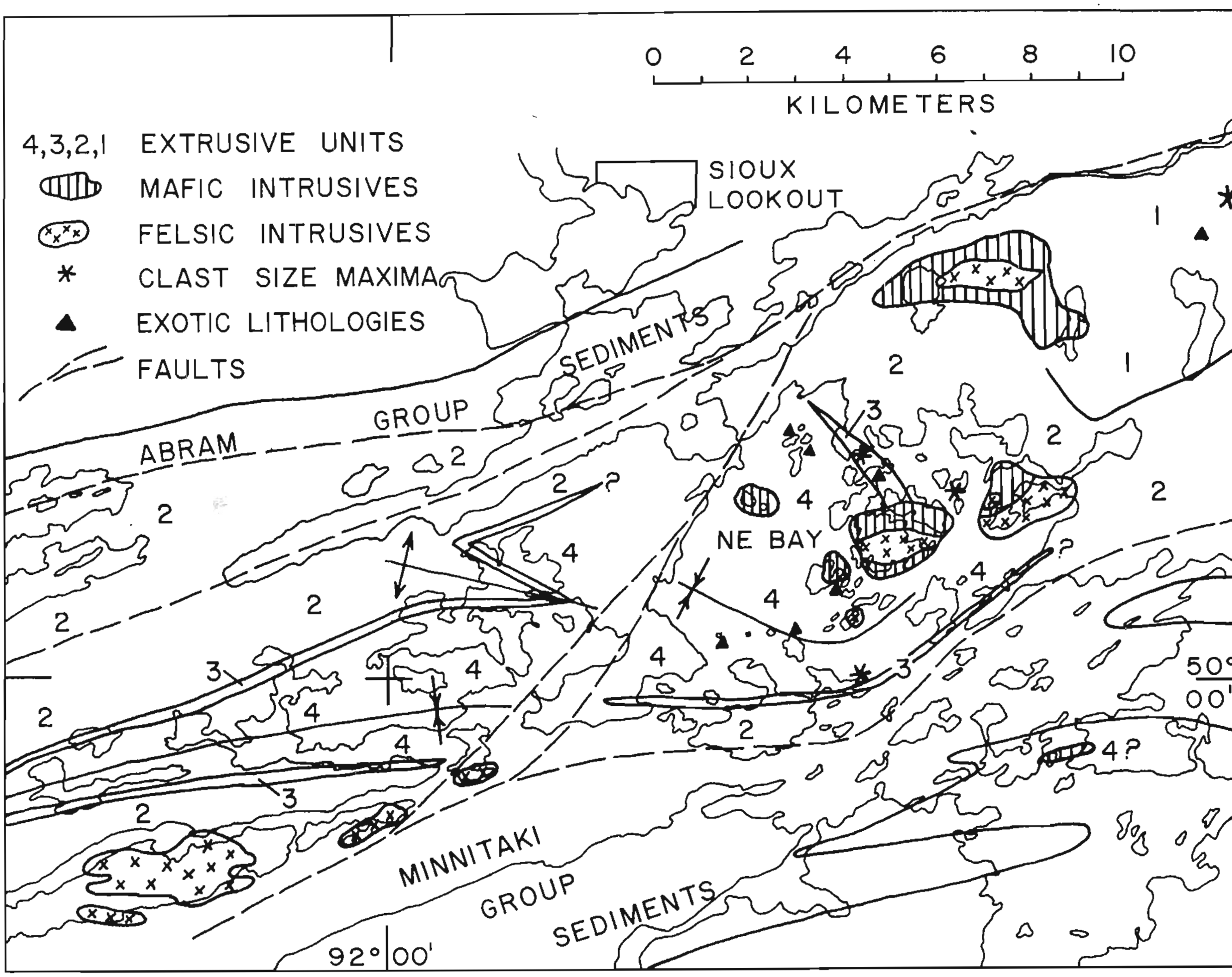

Figure 89.1. General geologic map, Minnitaki Lake Area.

[1]Department of Geology, McMaster University, Hamilton, Ontario, L8S 4M1

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Table 89.1

Major units and lithologies, "Central Volcanics", Minnitaki Lake

UNIT	THICKNESS	LITHOLOGIES: DOMINANT	LITHOLOGIES: ACCESSORY
4	1500 m (top not seen)	1. phyric andesite lavas 2. andesitic breccias and crystal tuffs 3. aphyric basalt lavas 4. basaltic pyroclastics	1. dacite-clast breccias 2. phyric basalt lavas
3	variable 100-500 m	mixed assemblage of Unit 4 + Unit 2 dominant lithologies	
2	3000 m	1. tholeiitic basalt lavas 2. felsic pyroclastics	1. feldspar-megacrystic basalt lavas or tuffs 2. mafic pyroclastics
1	ca. 2000 m (base not seen)	1. felsic lavas and pyroclastics 2. mafic lavas and pyroclastics	1. volcanic/plutonic-clast conglomerate (?) or breccia (?)

Stratigraphy

The volcanic sequence at Minnitaki Lake has been subdivided into four major units based on dominant and accessory lithologies (Table 89.1). Preliminary geochemical data indicate that the lower two units are a bimodal suite of tholeiitic basalt + dacite/rhyolite, while the upper units consist of basalt + andesite ± dacite with calc-alkaline affinities (Scime, pers. comm.).

Unit 1 is poorly exposed and has only been investigated on a reconnaissance basis. It contains notable occurrences of flow-banded rhyolite and rhyolite breccia (clast diameter $\leq$ 30 cm) which cannot be far removed from their source vents.

Unit 2 is well exposed in three areas and consists primarily of massive and pillowed tholeiitic basalt. The lower half (in the one section where it is seen) contains considerable bedded mafic pyroclastics intercalated with the effusive rocks. The upper 1000 m, seen in all three sections, contains numerous thin horizons of felsic pyroclastics, ranging from breccia to crystal tuffs and fine ash. In all three sections, feldspar-megacrystic basalt flows and/or crystal tuffs are found within the uppermost 300 m.

Unit 3 is transitional between units 2 and 4 and contains lithologies which are characteristic of both. It is well exposed in only one section but there contains an 80-m-thick unit of bedded, basaltic breccia, lapilli-tuff, and coarse lithic ash. This explosive unit is a valuable marker horizon. It is represented in other sections by exposures of 15-30 m of bedded mafic tuffs.

Unit 4 is a highly diversified sequence, with subequal proportions of effusive and fragmental rocks. It can be viewed simply as an alteration of basaltic and andesitic compositions, but this is made more complex by the occurrence of effusive and explosive units of either magma type. Thus, four major types of eruptive units comprise the unit 4 alternations:

a. plagioclase-phyric andesite flows and flow-breccias
b. andesite breccias and lapilli-crystal tuffs
c. aphyric basalt flows and flow-breccias
d. basaltic tuffs and breccias (agglomerates in part).

These alternations, involving both style of eruption and magma chemistry, bear many similarities to the relationships seen in Holocene sequences on Mount Pelée, Martinique (Roobol and Smith, 1976) and Mount Misery, St. Kitts (Baker and Holland, 1973). Geochemical studies now in progress on the pyroclastics will allow closer comparisons with the Holocene rocks.

Lateral Facies Variation

A vent complex, or more properly, a composite cone facies, has been outlined within the volcanics from two independent lines of evidence:

1. clast size variations within fragmental units
2. presence of exotic lithologies.

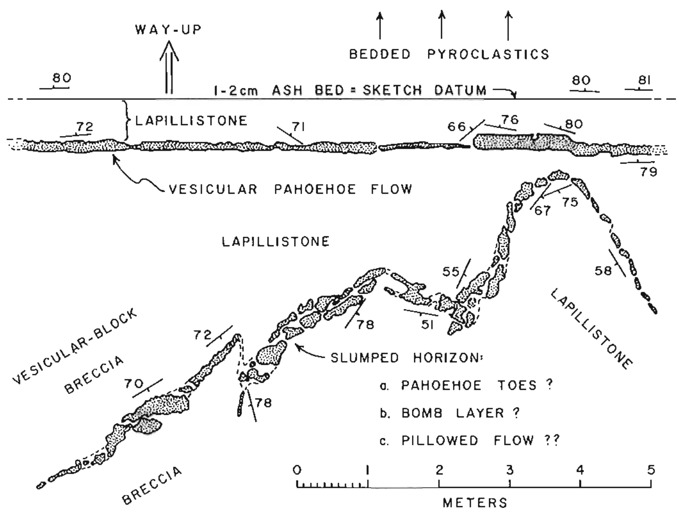

Figure 89.2. Outcrop detail, large-scale asymmetric slump fold.

Maximum clast sizes within units 2, 3 and 4 are clearly related to the style of eruption and chemistry of material produced. The largest clasts found in units 2 and 3 (dacite/rhyolite and basalt, respectively) are only 20 to 40 cm, while the unit 4 breccia clasts (andesite) are commonly 1.0 m and reach maximum of 3.0 by 1.0 m. In all three units, however, the largest clasts are found in the central or south-central portion of Northeast Bay, Minnitaki Lake. Distances between these maxima are < 5 km. The rhyolite lavas and felsic breccias of unit 1 are 5 and 10 km farther to the northeast, but are only some 5 km from the northern stock of the Northeast Bay Plutonic Complex.

Exotic lithologies are interpreted to have a proximal relationship to a source vent. A partial list of such lithologies and structures found only in the Northeast Bay area include:

1. vesiculated dykes
2. air-fall? bombs (crenulated margins and/or tear-drop forms)
3. cored bombs (blocks of earlier lavas enclosed in later lava, then ejected from a vent)
4. asymmetric, large-scale (10-200 cm) slump folds within bedded pyroclasts (Fig. 89.2).

Clast size variations and the presence of exotic lithologies point to a cone facies centred in the area of the Northeast Bay Plutonic Complex. This complex consists of composite stocks of gabbro, diorite, quartz diorite and quartz-feldspar porphyry. Border phases of the diorite are plagioclase-phyric, fine grained mafic rocks which are extremely difficult to distinguish from the unit 4 lavas. Collectively, there is a strong suggestion that the plutonic complex represents the frozen magma chambers which produced the Minnitaki Lake extrusives.

Within the cone facies and at some distance from it, well preserved sedimentary structures (graded bedding, loads, ripple cross-lamination, and scours) are found in bedded tuff units. These structures are of great importance to the reconstruction of the paleogeography. Bedded pyroclastics of unit 4 closely resemble the Miocene Tokiwa Formation, Japan (Fiske and Matsuda, 1964) and the Eocene Ohanapecosh Formation of Washington (Fiske, 1963). The Japanese examples have been assigned an open-sea water depth of 150-500 m, primarily on fossil evidence, but also on the basis of their unreworked aspect. Minnitaki Lake pyroclastics, by contrast show ample evidence of reworking and were certainly deposited in water above wave base (less than 150 m?). Slump structures and erosional surfaces, along with wood fragments, are common in the more "distal" Ohanapecosh lithologies, suggesting that land may have been emergent at the vent. This comparison provides a reference point for modelling the Archean volcanism at Minnitaki Lake.

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Project 750063

W. Blake, Jr.
Terrain Sciences Division

Introduction

An investigation of the geomorphology and glacial history of the Carey Øer (Carey Islands), Greenland was carried out between July 18th and August 12th, 1976. The easternmost island in the group, Björlings Ø, is located approximately 100 km west-northwest of Dundas (Thule Air Base) and 45 km southwest of the nearest point on the coast of Greenland proper (Fig. 90.1). These isolated islands, because of their position between Greenland and Ellesmere Island, offered the promise of providing unique data bearing on the glacial history of northern Baffin Bay, and a one-day visit in 1974 had revealed the presence of deposits and features worthy of further investigation (Blake, 1975).

Travel to the Carey Islands was by a Greenlandair Charter (GLACE) Bell 204B helicopter under the administration of the Royal Greenland Trade Department at Dundas. This aircraft also was used to visit two of the outlying islands, Björlings Ø and Bordø (Table Island). Additional helicopter support to visit Fireø (Fourth Island) and to revisit Bordø was provided by the *U.S.C.G.C. Westwind.* The base camp utilized was that established by the "North Water Project" in 1972 (Müller *et al.*, 1973). Now this base is administered by Kommissionen for Videnskablige Undersøgelser i

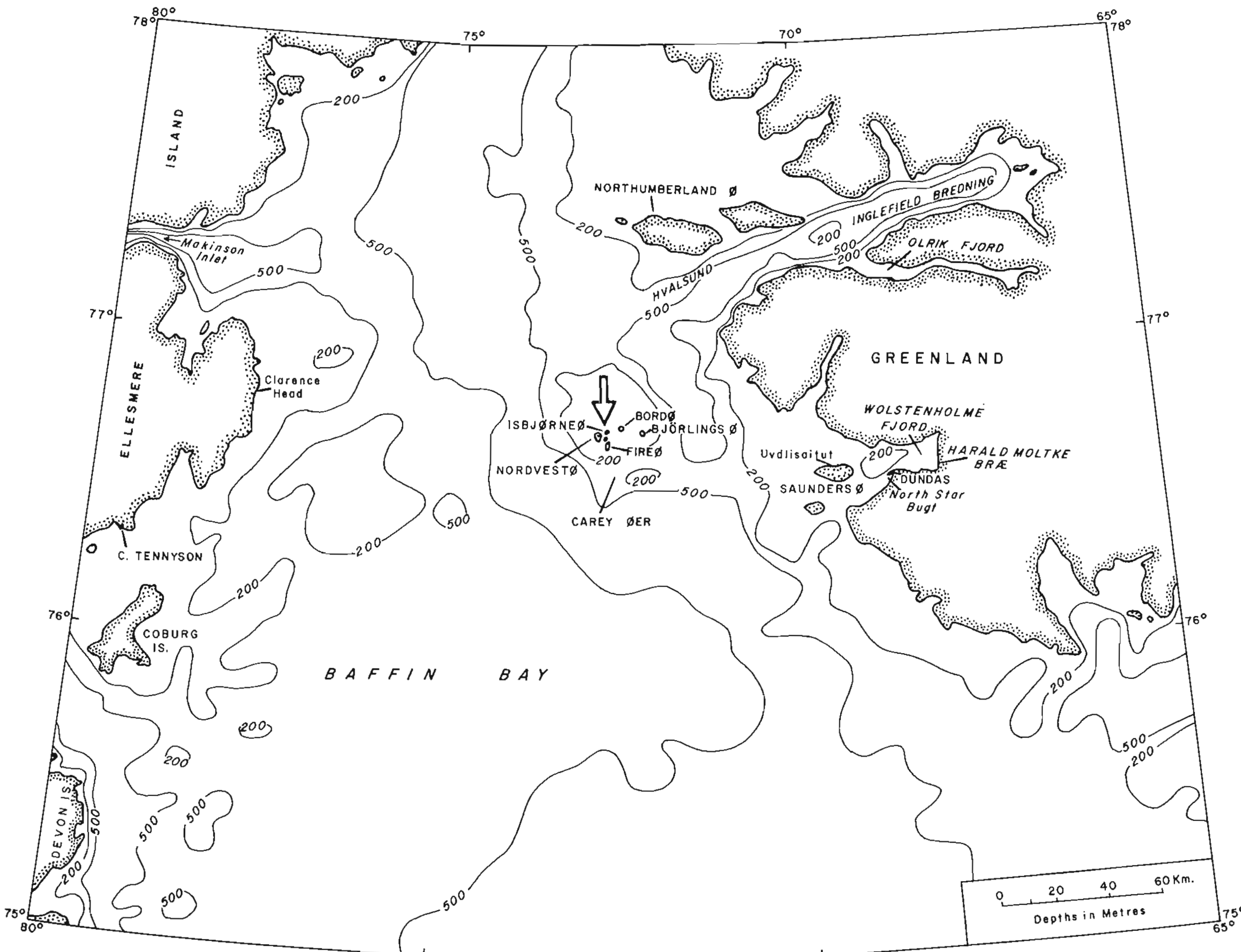

Figure 90.1. Location map, northern Baffin Bay. Adapted from Chart 896, "Arctic Bathymetry north of 72°, 0° to 90° West", Canadian Hydrographic Service, 1967. The arrow shows the general direction in which glacier ice once flowed across the Carey Islands.

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Table 90.1

Radiocarbon Age Determinations

Location	Approx. sample elev. (m) [1]	Material [2]	Field sample no.	Laboratory dating no.	Uncorrected age (conventional ^{14}C years before 1950) [3]	$\delta^{13}C$ ‰	Corrected age (conventional ^{14}C years before 1950)	Sample weight (g)	Counter (litres)	Pressure (atm)	No. of days counted	Remarks [4]
SE corner Nordvestø 76°42.8' N, 73°07' W	1.5	marine shells, *Balanus balanus* (L.)	BS-74-54	GSC-2102	3440 ± 130	+1.5	3460 ± 130	26.9	2	2	2	Shell fragments (41) in sand and gravel approx. 1.0 m below surface of cobble beaches. Calcite.
SW coast Isbjørneø 76°43.5' N, 73°04' W	19.5-20.0	marine shells, *Hiatella arctica* (L.)	BS-76-102	GSC-2372	7870 ± 70	+2.2	7900 ± 70	48.3	5	1	3	Intact shells (19 right and 8 left valves). Many paired valves in same deposit. Aragonite.
	8.0	marine shells, *Balanus balanus* (L.)	BS-76-61	GSC-2374	>38 000	+1.8	>38 000	48.2	5	1	3	Shell fragments (43) from intact individuals in living position on a single boulder. Calcite.
	2.5-3.5	marine shells, *Chlamys islandicus* (Müller)	BS-76-48	GSC-2367	38 200 ± 1100	+1.7	38 300 ± 1100	47.0	5	1	4	Shell fragments (22), but many intact and paired individuals in same deposit. Calcite plus <5% rhodochrosite.
Northern Nordvestø 76°44.2' N, 73°13' W	~140	moss peat, *Aplodon wormskioldii* (Hornem.) R.Br.	BS-76-128 (253-258 cm)	GSC-2368	6280 ± 80	-23.5	6300 ± 80	9.2	2	2	3	Peat consists entirely of this moss species, interspersed with ice lenses.

1 Elevations (rounded off to the nearest half-metre) were determined by Wild NK-10 level, using the approximate position of high tide as datum. The elevation of the peat on northern Nordvestø is only an approximate value determined by Paulin altimeter; it is uncorrected for temperature and pressure changes.

2 The barnacle fragments comprising GSC-2102 were identified by Dr. E.L. Bousfield, National Museum of Natural Sciences, Ottawa; the other pelecypod and cirriped samples were identified by the writer. The moss in GSC-2368 was determined by Dr. G.R. Brassard, Dept. of Biology, Memorial University of Newfoundland, St. John's.

3 All age determinations from the Radiocarbon Dating Laboratory, Geological Survey of Canada, are based on a ^{14}C half-life of 5568 ± 30 years and 0.95 of the activity of the NBS oxalic acid standard. Ages are quoted in conventional radiocarbon years before present (B.P.) where present is taken to be 1950. All finite age determinations from this laboratory are based on the 2σ criterion; i.e., there is a 95.5% probability that the correct age in conventional radiocarbon years lies within the stated limits of error. All "greater than" ages are based on the 4σ criterion (99.9% probability).

4 Mineral identifications were carried out by A. Roberts of the X-ray Diffraction Laboratory.

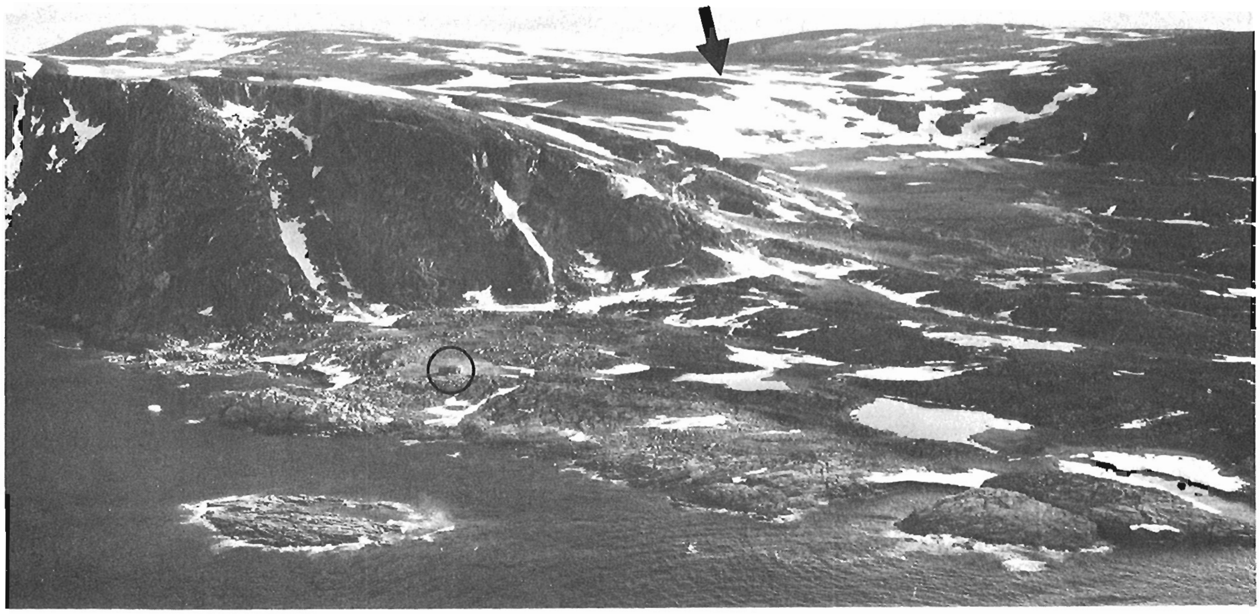

Figure 90. 2. Aerial view southwest at the strandflat along the north coast of Nordvestø. Note the rounded shape of the north-facing gneissic ridges above the base camp (circle). Striated and rounded rock surfaces abound on the plateau; the arrow indicates the position of the diabase ridge where Figure 90. 4 was taken, and one area of shelly till was directly beyond (southwest of) this ridge. July 23, 1976 (GSC-203107).

Figure 90. 3. Aerial view west along the northern coast of Björlings Ø. Note the rounded form of the main east-west ridge which forms the backbone of the island. The direction in which ice flowed across the plateau is indicated by the arrow. July 23, 1976 (GSC-203107A).

Grønland (the Commission for Scientific Investigations in Greenland), and it has been the site of both Swiss and Danish automatic weather stations, of which the latter is still in operation. The hut is situated on the strandflat at the northeast corner of Nordvestø (Northwest Island), and nearby a flight of raised beaches provides easy access to the plateau which makes up most of the island's area (Fig. 90. 2). The nearest islands, such as Isbjørneø, were reached by means of a 4. 2 m-long Canova rubber boat equipped with a 9. 5 h. p. outboard motor. The field party was picked up by *C. S. S. Hudson* on August 12th and landed at Dundas the following day.

Field work was devoted to four main topics: (1) studying the stratigraphy of unconsolidated deposits exposed in coastal cliffs and collecting marine molluscs and cirripeds for absolute age determinations; (2) searching for evidence of glaciation, such as striated and polished rock surfaces, as well as erratic pebbles and boulders; (3) coring some of the numerous peat deposits on the islands; and (4) echo sounding to obtain data on the depth and configuration of the inter-island channels. The present report will deal chiefly with the radiocarbon age determinations (Table 90. 1), four of which are on samples obtained in 1976.

Geomorphology and the Pattern of Ice Flow

The Carey Islands comprise six major islands (Fig. 90. 1) and a considerable number of tiny islets. The bedrock is dominantly gneiss, although a number of diabase intrusives are present also (Wordie, 1938; Munck, 1941; Bendix-Almgreen *et al.*, 1967). The summit plateaus of five of the main islands lie between 125 and 170 m a. s. l., but elevations over 200 m occur only on Nordvestø, the largest of the group.

In addition to the authors listed above, mention of features related to glaciation and emergence on Björlings Ø (Fig. 90. 3) was made as early as the last century by Chamberlin (1895a, 1895b), on the basis of his observations during the Peary Auxiliary Expedition of 1894 to northwestern Greenland. Later visits to the islands were made by Koch in 1916 (Koch, 1928) and by the Godthaab Expedition 1928 (Riis-Carstensen, 1931), but none of these expeditions was in the vicinity of the Carey Islands for more than a few days.

Observations made in 1974 and 1976 revealed that erratics are abundant on the summit plateaus of all the major islands shown in Figure 90. 1. In some places the ground surface takes on a varicoloured aspect, so numerous are the pebbles and cobbles of quartzite, sandstone, dolomite, conglomerate, granite, etc., whereas in other localities erratics are much more scattered. The rocks other than gneiss and diabase are all foreign to the Carey Islands, and the most logical source areas are to the northeast and north; cf. Dawes (1971) for a general summary of bedrock in North Greenland. Also, because of the sharp contacts between the diabase intrusives and the gneissic country rock on the plateau of Nordvestø, it was possible to see how boulders of one rock type have been plucked up and displaced southward onto the other rock type. Another type of erratic consists of fragments of marine mollusc shells. Bendix-Almgreen *et al.* (1967) were correct in suggesting that shell-rich till *should* occur, but they did not succeed in finding any. More time was available for searching in 1976, and as a result till containing a variety of shell fragments was discovered on both Nordvestø and Bordø. Both sites, at elevations >100 m a. s. l. (above the limit of Holocene boulder beaches), are situated in saddles on the plateaus, and the flow of glacier ice across the islands would have been concentrated in just such topographic depressions. As is shown in Figure 90. 2, the shelly till on Nordvestø is south of an extensive area of raised beaches which occupies a re-entrant in the bastions of gneiss making up the north coast of Nordvestø. This re-entrant provided the easiest route by which ice could reach the interior of Nordvestø, and the older marine sediments, which can be presumed to underlie the Holocene boulder beaches, would have furnished a ready source of mollusc shells.

In addition to the evidence provided by the erratics, the presence of striated and rounded rock outcrops, accompanied by plucking on the lee (south) side and by numerous chattermarks and crescentic fractures, attests to the flow of ice from north to south (Figs. 90. 4 and 90. 5). Striated surfaces were found on the summit plateaus on Björlings Ø, on Fireø, and in many localities on Nordvestø, even on the highest ridge at elevations above 200 m. When these observations are coupled with the gross morphology of the islands – i. e. cliffs are better developed at the southern ends of all the major islands, whereas the north-facing exposures are only steep slopes at many sites (cf. Figs. 90. 2 and 90. 3) – there can be little doubt that these islands have been overridden by ice moving southward (Fig. 90. 1). Presumably a major contribution to this southward flow came from ice which derived from Inglefield Bredning (Gulf) and Hvalsund (Whale Sound).

Marine Fauna and Radiocarbon Age Determinations

In the course of the brief visit to the Carey Islands in 1974, collections of marine fauna were made on the eastern side of Isbjørneø and near the southeastern tip of Nordvestø (Blake, 1975). During the flight to the Carey Islands in 1976, however, a site which appeared more promising was noticed at the southwestern corner of Isbjørneø (Fig. 90. 6). Later in the summer this pocket of marine deposits was visited several times by boat, resulting in some extraordinarily rich and interesting collections of marine fauna. Comments in this preliminary report will be limited to the species actually utilized for dating (Table 90. 1).

As is the case throughout the Carey Island, because of the virtually unlimited fetch available and the presence of open water for more of the year than is common at these high latitudes (Nutt, 1969), the surface of the marine deposits on Isbjørneø is made up of a veneer of cobble and boulder beaches (cf. Fig. 90. 7). Beneath the surface boulders are finer grained deposits (mainly sand), although boulders are present throughout the

Figure 90. 4. Striated diabase outcrop above an extensive beach area on the north coast of Nordvestø (cf. Fig. 90. 2). The direction of ice flow was north to south, from left to right in the photograph. The hammer handle is 32. 5 cm long. July 30, 1976 (GSC-203107B).

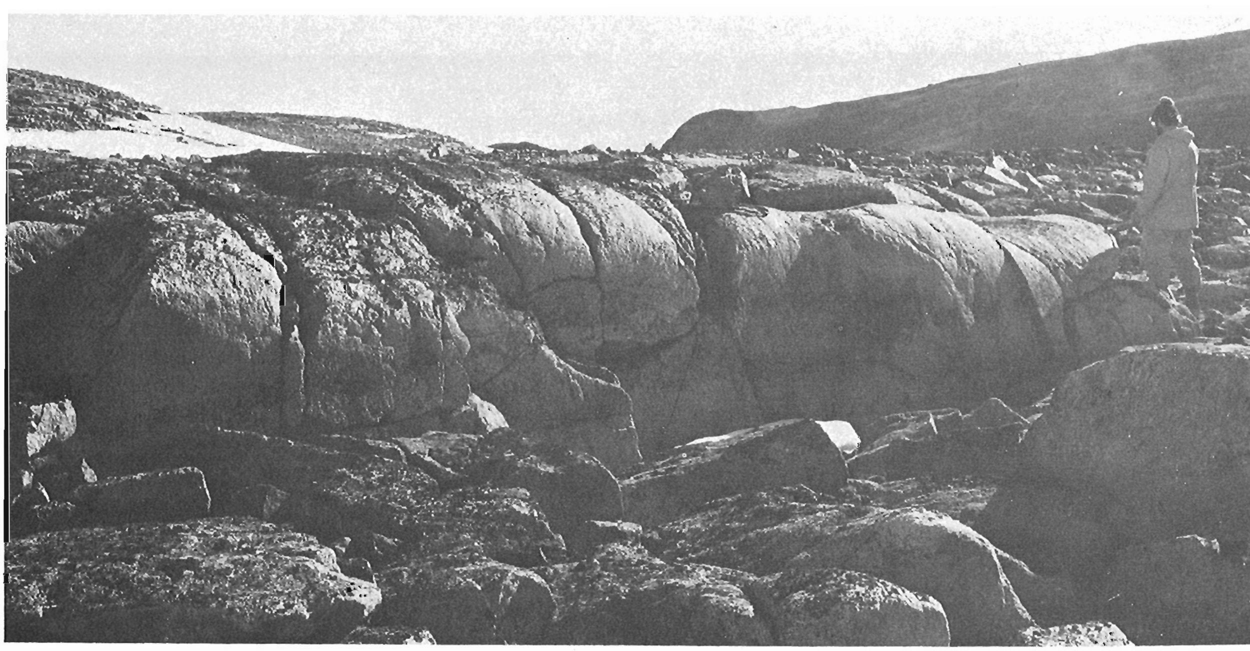

Figure 90. 5. View south-southwest at a rounded outcrop of gneiss in central Nordvestø. The direction of ice flow was toward the deep valley whose west wall is visible on the skyline. August 11, 1976 (GSC-203107C).

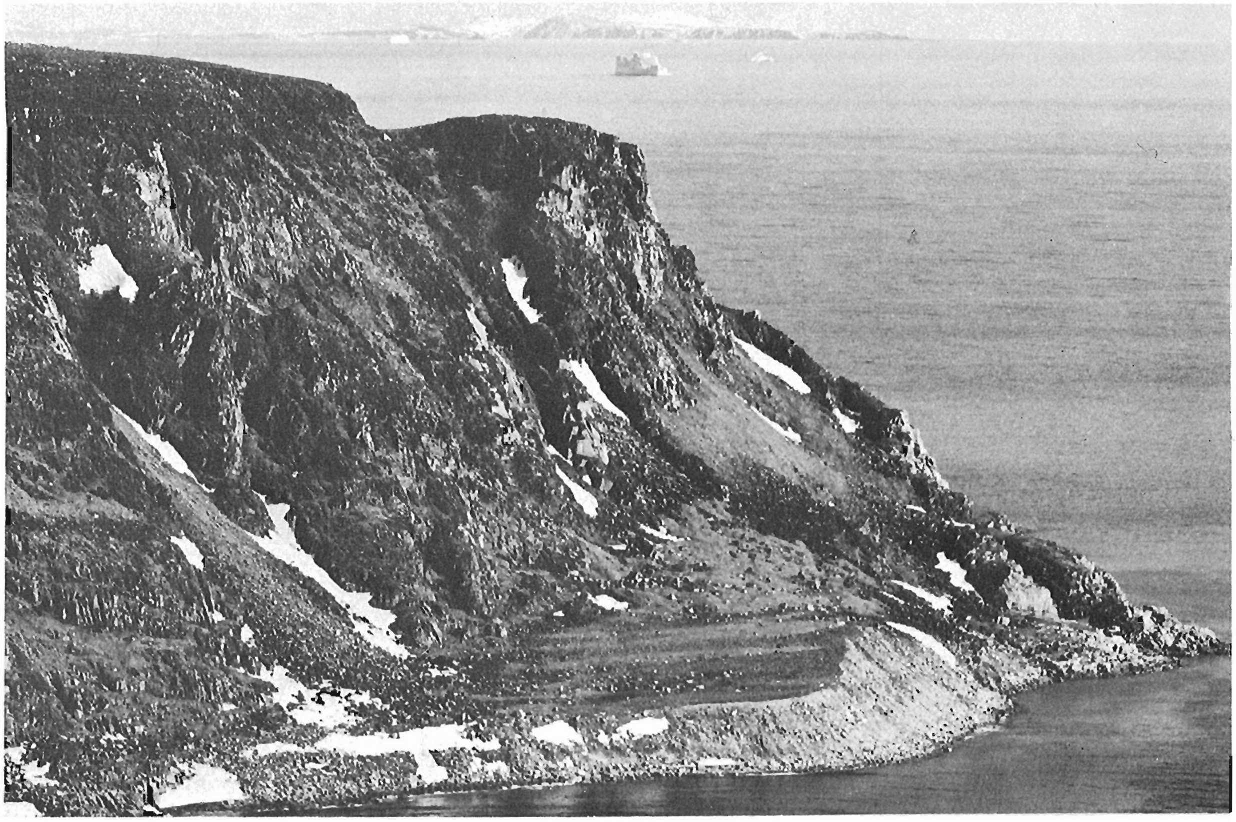

Figure 90.6. Telephoto view southeast from Nordvestø at the southwestern corner of Isbjørneø. The 25 m-high section of unconsolidated sediments here has a veneer of Holocene cobble-boulder beaches which are underlain by marine deposits more than 38 000 years old (GSC-2374; Table 90.1). July 19, 1976 (GSC-203107D).

Figure 90.7. Boulder beaches in the major valley extending inland from the south coast of Nordvestø. Spruce (*Picea* sp.) driftwood was found where the two people are standing, at an elevation of some 20 to 25 m. Many of the boulders comprising these beaches are over 30 cm in diameter. July 30, 1976 (GSC-203107E).

25 m-high section. Most surprising, perhaps, was the discovery of many intact, commonly paired, valves of the Iceland Scallop, *Chlamys islandicus,* in sand and in voids between boulders near the base of the section. Such large and well preserved individuals are rare in deposits which the writer has investigated elsewhere around northern Baffin Bay; in fact, because of its fragile nature this species usually occurs only as fragments. The largest intact individual, with a height of 9. 8 cm and a length of 9. 3 cm, is larger than any of the living individuals reported by Ockelmann (1958) from East Greenland and Jan Mayen, or than those from emerged marine deposits in the northern part of West Greenland (Laursen, 1944). Because the finite radiocarbon age determined for these shells (38 300 ± 1100 years, GSC-2367; Table 90. 1) is largely a function of statistics, it seems wisest to regard this value as a minimum age.

Higher in this section, in a zone characterized by an abundance of massive boulders (some over 1 m in diameter), numerous individuals of the Northern Ridged Barnacle, *Balanus balanus,* still were attached to rock surfaces. These shells, also extremely well preserved, were expected to be of Holocene age, but they, too, are beyond the limit of radiocarbon dating (>38 000 years, GSC-2374; Table 90. 1).

Near the top of the section, just beneath the veneer of boulder beaches at approximately the 20 m level, the fauna is dominated by the Arctic Saxicave, *Hiatella arctica,* including many intact and paired valves. These pelecypod shells, with an age of 7900 ± 70 years (GSC-2372; Table 90. 1), represent the oldest Holocene material dated so far from the Carey Islands.

None of the shell samples discussed above can be related to a specific position of the shoreline, nor can the barnacle fragments from an elevation of 1. 5 m in southern Nordvestø (3460 ± 130 years, GSC-2102; Table 90. 1). The highest beaches at the Isbjørneø locality shown in Figure 90. 6 are about 30 m a. s. l. ; above that elevation there is too much turf and scree masking the surface for beaches to be discernible. On Nordvestø at the locality shown in Figure 90. 7, however, Bendix-Almgreen *et al.* (1967) reported that the boulder beaches extended to approximately 265 feet (80 m), and altimetry in 1976 confirmed that their result is of the right order of magnitude. All that can be stated at present is that when the 7900 year-old pelecypods were living, relative sea level was certainly above 23. 5 m (the elevation of the boulder beaches immediately above the collection site), and it could well have been above 30 m.

As to the "old" marine fauna, both of the species dated live at varying depths. Pilsbry (1916) states that *Balanus balanus* occurs as deep as 165 m, and Bousfield (1960) gives a value of 180 m. In the vicinity of Saunders Ø and Wolstenholme Fjord *Balanus balanus* has been dredged from several sites between 30 and 40 m depth (Vibe, 1950). For *Chlamys islandicus* the range is given as up to 150 m by Odhner (1915) and up to 356 m by Ockelmann (1958). In view of the severity of the storm waves which pound the coasts of the Carey Islands at present, it seems reasonable to assume that the level of the sea could easily have been 10 m or several tens of metres higher, relative to the land, when the delicate scallops and the barnacles were the dominant elements in the marine fauna.

During the collecting on Isbjørneø no significant break, such as might be represented by a till or by a unit devoid of marine fauna, was observed in the section, although extensive digging could not be undertaken because of the danger of boulders rolling down from above. In terms of the general grouping of radiocarbon age determinations (i. e. , Holocene and "old"), however, the section is similar to those studied earlier on Saunders Ø, Coburg Island, and at Cape Storm on Ellesmere Island (Blake, 1976). Hence, it is perhaps not unreasonable to postulate that the *Chlamys islandicus* unit on Nordvestø corresponds to a gravel unit on Saunders Ø in which fragments of *Mytilus edulis* are >40 000 years old (GSC-2143; Blake, 1976) and in which fragments of *Chlamys islandicus* are also present. Likewise, on Coburg Island, fragments of *C. islandicus* occur with *Hiatella arctica* shells >40 000 years old (GSC-1062), beneath a *Mytilus*-bearing unit for which an age determination of >38 000 years (GSC-1425; Blake, 1973) was obtained.

Peat Deposits

Following their visit to Nordvestø in 1965, Bendix-Almgreen *et al.* , (1976) reported that "very thick (4 feet or more) peat deposits of the palsa type were found at several places near the edge of the summit". Thick deposits of peat were observed by the writer on top of Dark Head, Isbjørneø in 1974 (Blake, 1975), and peat mounds elsewhere on that island were discovered in 1976.

Coring with a SIPRE-type auger (equipped with tungsten carbide cutters; core diameter, 7. 6 cm) was carried out at two localities on Nordvestø in 1976. One site, in the central part of the island, is located in the upper reaches of the same valley shown in Figure 90. 7; the other site is at an elevation of approximately 140 m on the edge of the plateau, overlooking the north coast of the island (Figs. 90. 8 and 90. 9). At the north coast site, one mound was cored to 258 cm below the surface, at which depth an object (presumably either bedrock or a boulder) was encountered of sufficient hardness that the cutters started to chip. Core recovery was 100 per cent, and the increment at 253 to 258 cm depth gave an age of 6300 ± 80 years (GSC 2368; Table 90. 1). Dating of the top of the frozen peat (only the top 15 cm was thawed) is planned as part of a more detailed treatment of this interesting site. In addition to its unusual location, the moss species constituting the basal dated peat, as well as the rest of the core, is *Aplodon wormskioldii;* this is a species commonly restricted to growing on carcasses or excrement (G. R. Brassard, pers. comm. , 1976; cf. also Holmen, 1960; Brassard, 1971; Kuc, 1973).

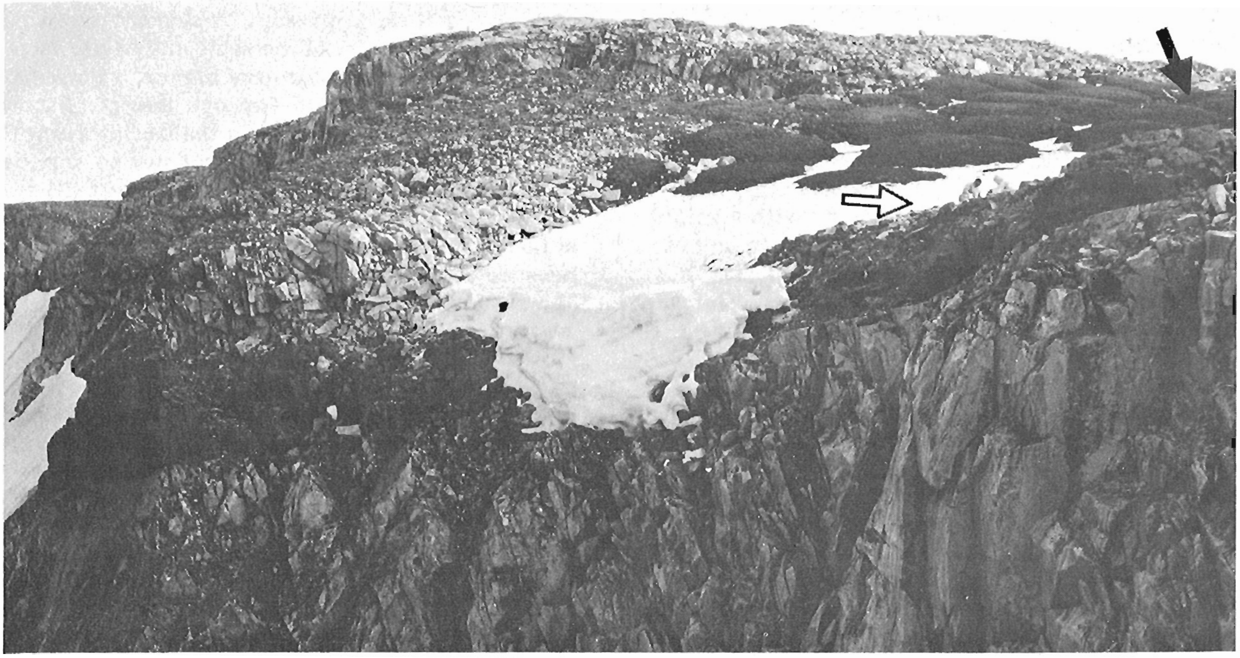

Figure 90. 8. View southeast at peat mounds and coring site (black arrow) on the plateau rim along the north coast of Nordvestø. Erratics plus striated boulders of locally-derived diabase were common on the edge of the plateau at the position indicated by the open arrow. August 9, 1976 (GSC-203107F).

Figure 90. 9. Detail of the coring site shown in Figure 8 (view northwest). The basal peat here, at a depth of 253 to 258 cm, is 6300 ± 80 years old (GSC-2368; Table 90. 1). August 7, 1976 (GSC-203107G).

Acknowledgments

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91. ALTITUDINAL WEATHERING ZONES AND GLACIAL LIMITS IN WESTERN NEWFOUNDLAND, WITH PARTICULAR REFERENCE TO GROS MORNE NATIONAL PARK

Project 740072

Douglas R. Grant
Terrain Sciences Division

Introduction

A major new approach has been taken to understanding and differentiating the limits of successive glaciations and to developing a Pleistocene stratigraphy for western Newfoundland. The basis is the recognition over wide areas of concordant limits of three distinct degrees of weathering of glaciated terrains. That is to say, one can distinguish a lower, freshly glaciated zone, a middle higher zone of weathered glacial bedrock landforms and an upper zone of broken bedrock and tors devoid of glacial features except rotted erratics.

Previous Work

The present ability to delimit and appreciate these differences has evolved over several decades as a result of early observations in Newfoundland, numerous papers on the subject in Arctic Canada, and the author's ongoing task of mapping the surficial geology of Newfoundland and other areas in the Atlantic Provinces. Coleman (1926) for example cited numerous localities in Newfoundland where undegraded glaciated areas abutted against weathered terrain and concluded that the last major glacial expansion was restricted to lower areas within an older drift zone. Fernald (1925) provided corroborative botanical evidence in the form of disjunctive populations of arctic-alpine plants that required the existence of ice-free biological refugia in western Newfoundland during at least the last glaciation. The hypothesis fell into disrepute when Wynne-Edwards (1937) disputed this interpretation, contending that the species required coastal refuges, not summits, and when Flint (1943, *et seq.*) disallowed any possibility of less than complete glacier cover.

More recently, mapping by Grant (1969a) drew attention to smooth, rounded, felsenmeer-covered summits rising above the roche moutonnées surface of the Long Range Mountains at the northern end of Northern Peninsula. These were regarded as nunataks projecting through the local ice cap during the latest glacial expansion ca. 11 000 years B.P. Related papers (Grant, 1969b, 1972) expanded the scope by treating the remainder of the Northern Peninsula where several dozen more summits of felsenmeer delimited by ice-marginal features were depicted as having remained above "a late Pleistocene readvance of piedmont glaciers" from the Long Range ice cap that deployed between coastal highlands onto the lowlands as contiguous expanded-foot glaciers. A series of detailed surficial geology maps (Grant, 1973) differentiated three bedrock zones according to the degree of preservation of glacial erosion, viz., an exposed fracture-lineated, scoured, and polished bedrock terrain, abutted by an intermediate zone of glacial rock basins and knobs that were distinctly subdued by surface weathering, surmounted by summit areas with soliflucting colluvial slopes and felsenmeer tracts nearly devoid of any depressional or constructional features except a few scattered erratic stones. The significance of these contrasts in terms of glacier chronology was not realized until mapping of the Burin Peninsula on the south coast of Newfoundland (Grant, 1975a) revealed evidence for three major glacial pulses separated by nonglacial periods of weathering.

Similarly, mapping in southwestern Nova Scotia (Grant, 1976a) and in northern Cape Breton Island, Nova Scotia (Grant, 1976b) produced evidence that the last (late Wisconsinan) ice failed to re-expand over certain coastal and/or high areas remote from the apparent sources of ice dispersal. The evidence was marine deposits ca. 38 600 years (GSC-1440) old not covered by late Wisconsinan tills, an intertidal rock platform of last interglacial age bearing no sign of glacial overriding, and a peat bed more than 38 000 years old (GSC-283) overlain not by till but by colluvium and alluvium. These observations and earlier interpretations for New Brunswick and Gaspé (Chalmers, 1886), for the Magdalen Islands (e.g., Coleman, 1919), and for southwest Newfoundland (Brookes, 1975) were the basis for a tentative portrayal of the late Wisconsinan ice margin in the Gulf of St. Lawrence region (Grant, 1976b) showing in essence only a moderate expansion of upland ice caps beyond the present coast.

Review of earlier accounts of similar weathering zones in northern Labrador and Baffin Island (summarized in Ives, 1975), and discussions with A.S. Dyke, who has had experience in both Newfoundland and Baffin Island, prompted a more exhaustive examination of airphotos of western Newfoundland. Two samples of aerial photographs exhibiting terrain contrasts attributable to varying periods of weathering are shown in Figures 91.1 and 91.2. Figure 91.3 summarizes the airphoto interpretation. Dyke attested to the similarity of the three Arctic weathering zones to those seen in Newfoundland and emphasized the conclusion that each subsequent glacial advance was of lesser extent. If the interpretation placed on the Arctic weathering zones is considered to be valid in the Newfoundland situation, then the areas of felsenmeer and degraded roches moutonnées must be of considerable antiquity. The lack of chronologic control on these phases, except for the two indirectly related Nova Scotian dates cited above, precludes any but the most speculative correlation with published stratigraphic subdivisions of the late Pleistocene.

On two occasions in September and October 1976 an opportunity to improve knowledge of the surface aspect of these zones, which had been interpreted from

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Geol. Surv. Can., Paper 77-1A (1977).

airphotos was provided during a period of several days spent in Gros Morne National Park, serving as geological consultant to the National Film Board in the preparation of a film for Parks Canada depicting the geological evolution of the area. Helicopter traverses along fjords and landings on the intervening tablelands (Fig. 91.4) enabled direct observation of glacial features known to mark the weathering breaks. Of special interest was the exact degree of surface degradation, specifically the grade of felsenmeer, the size of tors, the thickness of oxidation rinds on both erratic and local boulders, and the relative loss of glacial markings on outcrops.

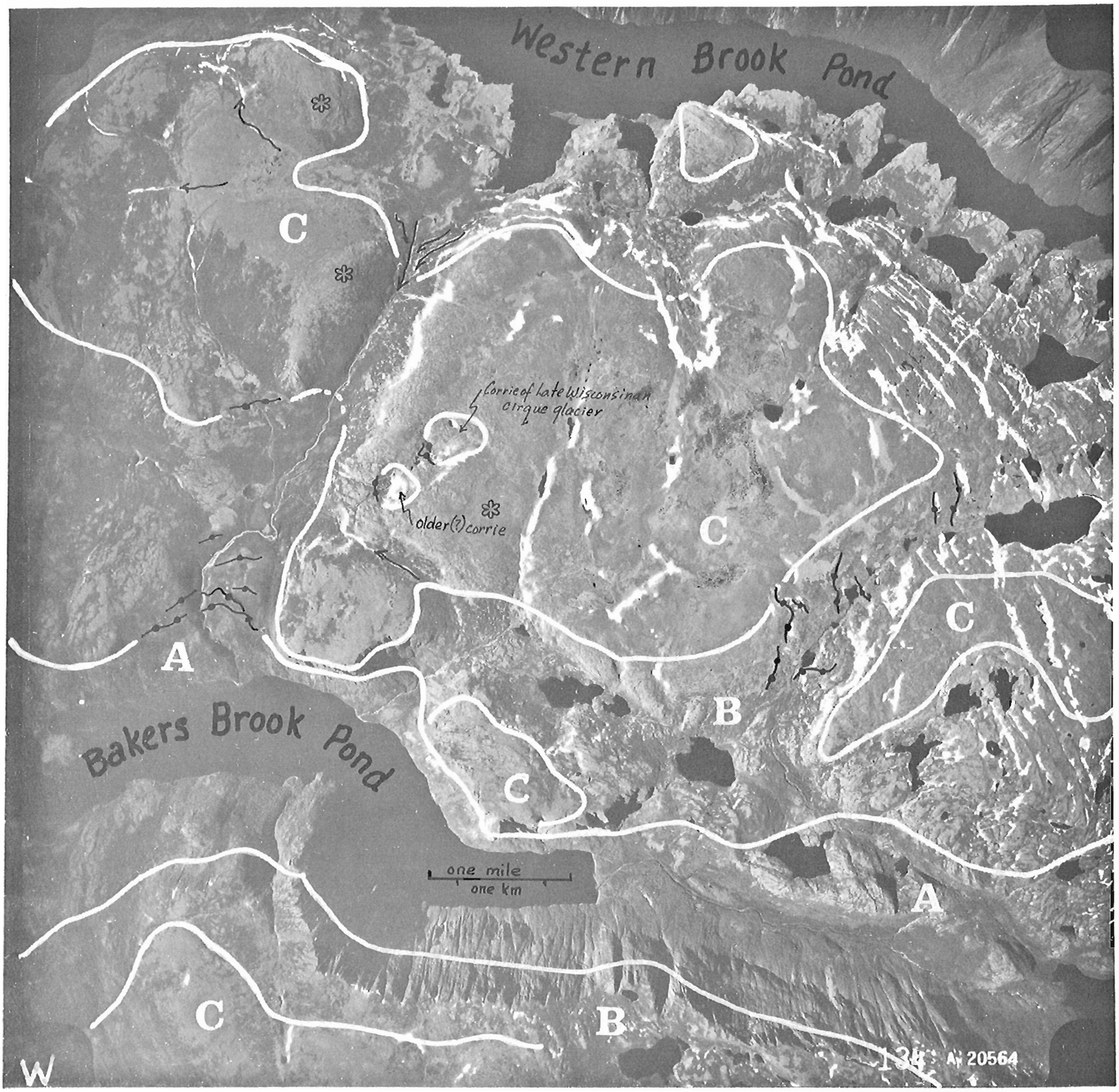

Figure 91.1. Plateau between Bakers Brook Pond and Western Brook Pond showing terrain contrasts produced by weathering subsequent to three major glacial phases. A, B, and C denote major weathering zones; beaded lines are moraines; arrows mark meltwater channels; asterisks are tors (National Airphoto Library Canada, A20564-134).

Description of Weathering Zones

Zone A

Weathering zones are designated A, B, and C from youngest to oldest to provide for possible additional older zones; the older zones are recognized and defined in terms of their contrast with the most recently developed zone. Zone A (Fig. 91. 5) on Long Range is characterized by barren rock with a freshly striated, grooved surface still with glacial polish, on which equally unaltered erratics are perched. The upper boundary of this zone on the plateau near the heads of the fiords is at 600 m and its descent down the fiords can be traced as a polished surface on the lower part of the sheer 600 m-high cliffs. This limit blends concordantly with end moraines at 120 to 150 m on the coastal plain. In places adjacent ice lobes appear to have coalesced and interlobate

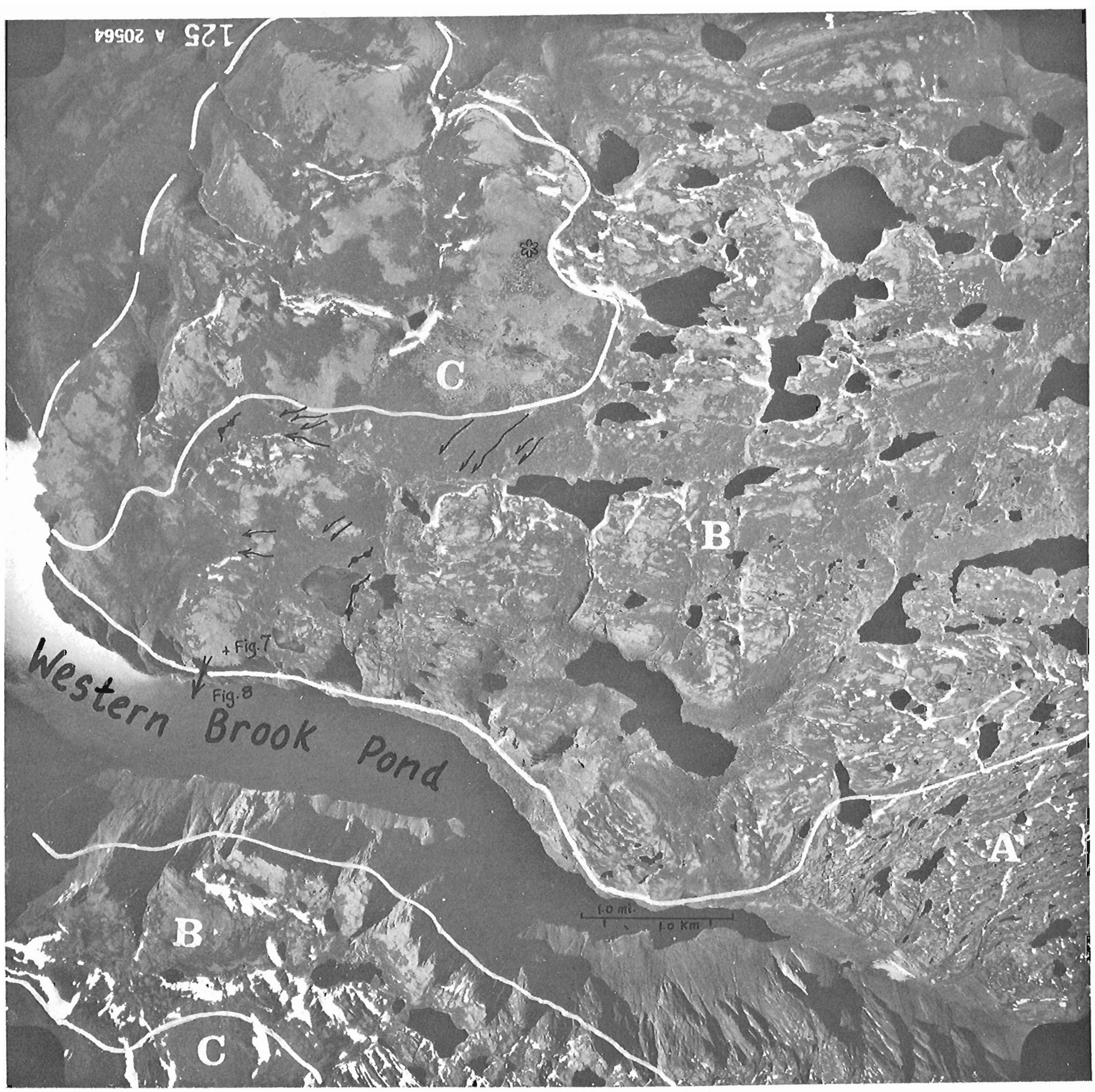

Figure 91. 2. Airphoto of the plateau north of Western Brook Pond showing sharply etched bedrock relief of Zone A, basined but grusified nature of Zone B, and smooth felsenmeer of Zone C. Same legend as Figure 92. 2. (National Airphoto Library Canada, A20564-125).

A, B, and C designate weathering zones:

A (unpatterned) = "freshly" glaciated terrain covered by late Wisconsinan ice;

B (stippled pattern) = grusified bedrock probably not glaciated since early or middle Wisconsinan time;

C (solid black) = felsenmeer area possibly not glaciated since the last interglacial;

dashed lines mark = late Wisconsinan end moraines;

dotted lines mark = older moraines;

diagonal ruling marks = late Wisconsinan marine overlap of former ice-free area;

question marks = hybrid areas of uncertain affiliation;

numerals = approximate elevations of boundaries in feet.

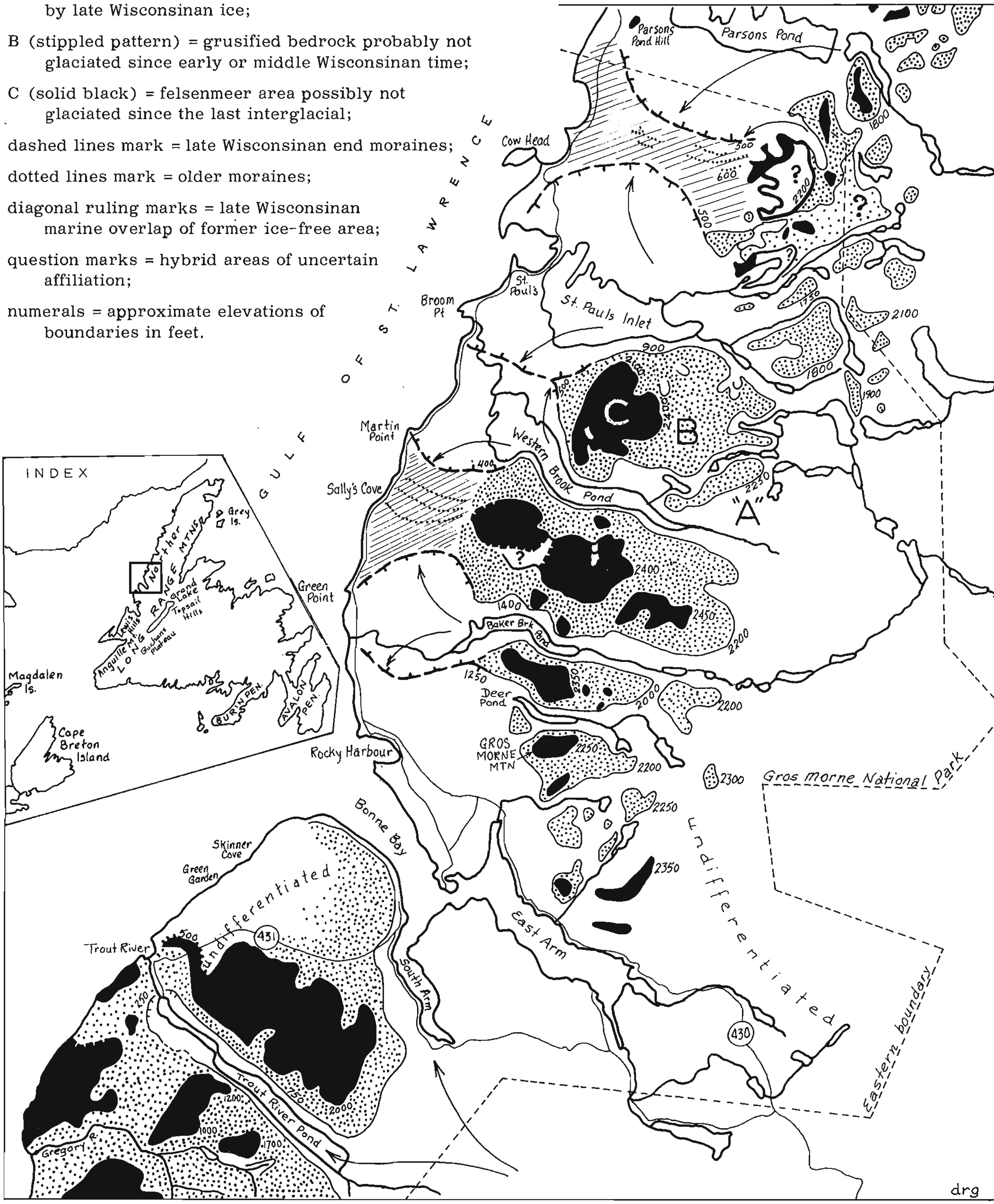

Figure 91.3. Weathering zones delimiting late Pleistocene glacial boundaries and former nunataks in Gros Morne National Park.

moraines were built between Deer Pond and Baker Brook Pond and between St. Pauls Inlet and Western Brook Pond. But in two cases the piedmont glaciers did not coalesce, so that ice-free enclaves existed on the coastal lowlands, as inland from Sally's Cove and Cow Head. The glacier that occupied the trough of Trout River Pond terminated a mile from the present coast and built a marine terrace of outwash into the sea at 45 m a.s.l. The exact limits of the ice cap that fed these glaciers from a source on Northern Peninsula is not known, neither is the relationship with ice reaching Bonne Bay and Trout River from a separate inland ice cap located farther south over Grand Lake basin.

During this glacial phase some of the many corries fretting the margins of the tablelands were occupied by ice, with two or three distinct looped moraines being built at their lips.

The exact time of culmination of this last major pulse is unknown, but the series of end moraines in this area and other corresponding moraines associated with troughs farther north were built prior to 12 600 years ago, judging by the age of shells in marine deposits laid down over them after recession (Grant, 1972).

Zone B

This zone retains the glaciated rock-basin character of Zone A, but the gneissic rock structure that is etched out so sharply in Zone A is much more subdued and the average minimum size of discernible bedrock prominences is ten times larger (Fig. 91.6). The distinguishing characteristic of this zone is the lack of glacial erosional features up to the dimension of deep grooves and the presence instead of a layer of disintegrated rock (grus). The rock surface, where visible, is degraded; minerals have weathered out; the surface is etched and more resistant veins stand out in relief of several inches. Erratic blocks, while numerous and scattered everywhere, are similarly etched and crumbling (Fig. 91.7).

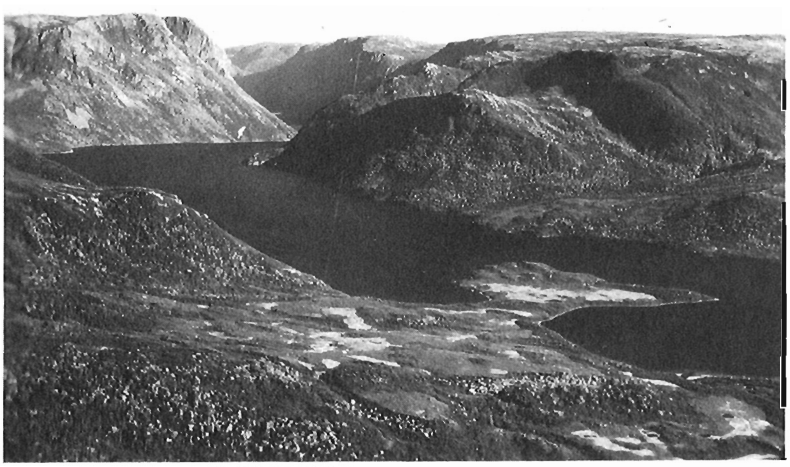

Figure 91.4.

Aerial view eastward into Bakers Brook Pond glacial trough, with adjacent felsenmeer-covered tablelands.

Figure 91.5.

Aerial view of Zone A showing craggy scoured bedrock and perched erratics.

Figure 91. 6.

Aerial view of Zone B showing rock basins and hills of subdued relief mantled with grus and mudboils in till or colluvium.

The upper limit of this zone is usually found at elevation of 680 to 750 m where it is commonly marked by small moraines, sidehill meltwater channels, or a zone of stripped bedrock; proglacial overflow channels are found in cols where ice tongues can be inferred to have lodged in re-entrants. In a few cases valley fillings are recognized in tributary valleys where an ice tongue had blocked the drainage. This limit, though not precisely definable within the confines of the fiords, is represented locally by a few subdued moraines banked into fiord tributary valleys. The extent of glaciers on the lowlands during this period might be represented by moraines near Sally's Cove and Cow Head.

At this point a comment about the implications of successive but incomplete glaciations for sea level chronology might be interjected. If the amount and extent of marine submergence following the last major glaciation can be established, and if the last ice expansion did not cover certain parts of the coast such as at Trout River, then the possibility exists of finding shorelines belonging to the marine submergence related to the previous glacial event. That glaciation was more extensive judging by the fact that it reached some 100 m higher around the nunataks, consequently the two marine submergences should be separable on the basis of maximum elevation, as well as in terms of any differences in the weathering of their respective sediments. The likelihood that the numerous raised shorelines in the region belong to more than one unrelated glacial episode limits the value of previous syntheses of shoreline deformation and calls for a completely new evaluation of these features throughout Newfoundland.

To conclude, the age of the glaciation after which the weathering of Zone B proceeded is unknown. It probably will never be possible to derive the time interval solely from the degree of rock surface degradation compared to the "fresh" surfaces of Zone A. A more fruitful approach would be to search for dateable materials in related deposits. Material for dating might come from the associated marine deposits, which theoretically could be present in the areas not overrun by the later ice, or from lake sediments in the rock basins within this zone.

Zone C

This, the most strikingly obvious terrain unit because of its dissimilarity to the usual character of Newfoundland, has been described by most reports on

Figure 91. 7. Closeup of nature of granitic erratics in Zone B showing etched and basined surface *see* Fig. 91. 3 for location).

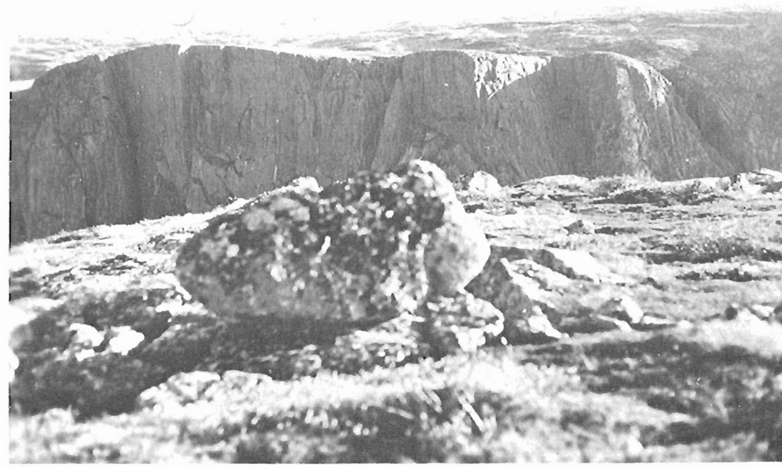

Figure 91.8.

View across Western Brook Pond from Zone B, with flat skyline across Zone C illustrating absence of glacial relief (*see* Fig. 91.3 for location).

the area. The surface is flat to rolling (Fig. 91.8), without any trace of constructional glacial features or rock basins except where corries have been excavated along its margins by cirque glaciers dating from younger ice ages. The local relief is only about one metre - attributable to large stone polygons and sorted circles (Fig. 91.9) or mud boils. The entire surface is mantled with a patterned felsenmeer. Rarely does bedrock emerge intact, and then only as heaps of slightly disrupted blocks or embryonic tors. Joint features are deeply etched to impart a mammillated surface. Any grus produced has been lost in the deeper parts of the felsenmeer mantle or incorporated in mud boils, which are circulating material including stones that come as erratics from distant inland sources. Such materials appear relatively fresh because of the rapid mechanical disintegration and the short residence time on the surface. On the other hand, on the tors, within the felsenmeer, and along the stone gutters of the sorted circles, erratic blocks several feet across, which are presumed to have remained on the surface for a considerable period, are oxidized to depth of several centimetres. In general, the weathering rinds on such stones are 10 times as thick as they are in Zone A.

An interesting parameter, bearing on the development of weathering products on terrains of successively longer subaerial exposure, is the degree of dissection on the slopes of the former nunataks. This is particularly well displayed on the massif separating Western Brook Pond from St. Pauls Inlet. There, each stream flows through three valley segments. The lower course in Zone A, representing erosion during postglacial time, is no more than a gully 5 to 10 m deep; the intermediate segment in Zone B might be described as a deep ravine, about 30 to 50 m deep, whereas the upper reach in Zone C is a large gorge 100 to 200 m deep. Such differential entrenchment may suggest a measure of the real time intervals involved.

The age of Zone C, that is to say the time since it was affected by ice carrying erratics, is tentatively believed to be more than 100 000 years, or in other words it is considered not to have been ice covered since before the last interglacial. This estimate is made because of the similarity of surface aspects to those described in Baffin Island, and because of what appears to be at least 10 times more surface degradation and slope dissection than has occurred in 10 000 years of postglacial time.

Discussion and Summary

Weathering Zones as Stratigraphic Units

The significance of these weathering zones lies in the fact that they manifest the time intervals that have elapsed since they were glaciated. In essence the weathering products and alteration effects constitute a stage of soil development, so that insofar as bedrock surfaces are concerned they represent soil-stratigraphic time units. Theoretically, there should be places where tills of each successive glacial stade are superimposed, each with its own characteristic depth of alteration. In the same way, tills could overlie and preserve ancient weathered bedrock surfaces. The point is that, although these zones are recognized mainly by their areal juxtaposition and altitudinal separation, they form a sound basic for a stratigraphic subdivision of the Newfoundland Pleistocene. Following the preparation of maps showing in detail their development throughout the western highlands, the relationships and status should be sufficiently well established to warrant formal designation. In this connection, it might be noted that their distribution is not restricted to towering summits of the western escarpment. Terrain mapping of the Red Indian Lake map-area (NTS 12 A) southeast of Grand Lake has revealed an equivalent contrast between young, fresh tills in the lowlands and weathered summits and areas of distinct till lineation on the Buchans Plateau and Topsail Hills (Grant, 1975b, 1976c). The existence of ice-free areas in the interior is impossible if the island supported a single independent late Wisconsinan age ice cap (Brookes, 1970). A.S. Dyke, (pers. comm., 1976) can reconcile these conflicting

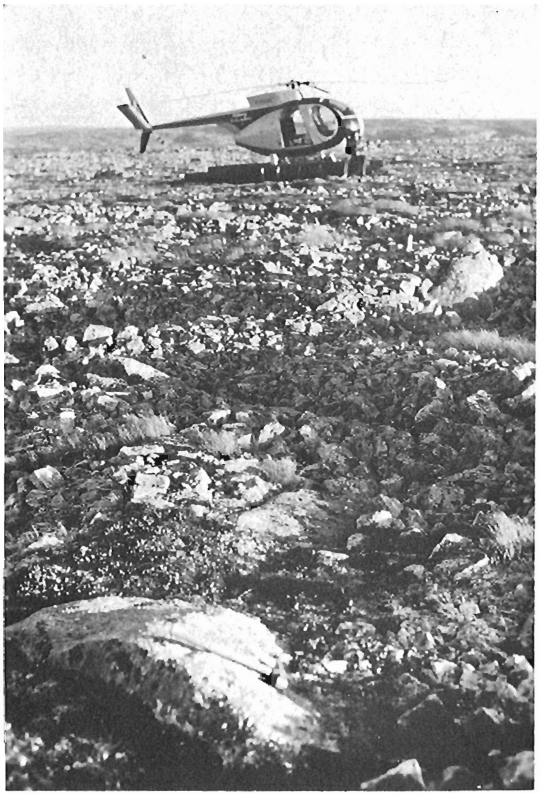

Figure 91.9. Closeup of felsenmeer of Zone C on Gros Morne at elevation 750 m (2500 feet) showing stone circles composed of quartzite, and large gneissic erratics.

facts by hypothesizing that a large early Wisconsinan ice cap complex thinned, shrank, and separated during interstadials into several remnant ice domes located at lower elevations in interior basins, away from high areas too small to sustain glacierization. These residuals, figured by Grant (1974) only for the late Wisconsinan, became the nuclei for expansion during ensuing glacial periods which were simply not intensive enough to reglaciate high areas like the Buchans Plateau.

The Dilemma of Wet Sliding Ice vs Dry Fixed Ice

The close association of freshly scoured fluted till areas and felsenmeer plains is not unique to Newfoundland. The Cairngorm Mountains of Scotland, also a maritime region affected by local ice caps, exhibit the same contrasts which Sugden (1968, 1976) explains as due to the transition, beneath an overriding ice sheet, from a "warm-based" zone at the pressure melting point capable of intense scour and erosion, to a "cold-based" zone essentially frozen to the bed with shear movement occurring only at higher levels. The relatively sharp elevational limits of the resulting bedform (or lack of them) he attributed to the fact that topography, and hence thickness, determines the thermal regime. By this hypothesis, troughs get scoured because ice is thick and therefore sliding, whereas intervening flat summit areas remain unmodified and preglacial surface characteristics are preserved because the ice is thin and fixed at the base. In this way delicate tors and loose felsenmeer can abut roches mountonnées without implying ice margins or weathering intervals. The weaknesses of this model are that it requires a very sharp transition of thermal regime to account for the abrupt changes of sruface aspect, and it cannot explain the occurrence of moraines and meltwater features at zone boundaries. In conclusion, for Newfoundland at least, these weathering differences are taken to mark the limits of successivvely weaker glacial expansions and to manifest the duration of subaerial exposure.

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ESTIMATION OF BULK DENSITY AND WATER CONTENT OF BEAUFORT SEA SEDIMENT CORES USING X-RADIOGRAPHS

Project 700092

R. V. Wahlgren and C. F. M. Lewis
Terrain Sciences Division

Fresh sediment cores are commonly X-radiographed while still sealed in plastic core liner tubes. Such radiographs, although crude because of parallax distortion induced by the core thickness, conveniently provide a useful and permanent film record for description and analysis of sedimentary and organic structures and for selection of subsampling zones. It has been noted previously (Rukavina, 1967; Edmondson and Allison, 1970) that radiographic film density variations are strongly correlated with sediment water content and porosity. It is also known that cellulose acetate butyrate (CAB) plastic core tubing is somewhat pervious to water vapour. Cores have been observed to 'dry out' significantly over long storage periods, yet it is often desirable to speculate on the original physical properties of the cored sediment when interpreting variations of strength, acoustic reflectivity, or the consolidation history of the sediment column, etc.

The principal purpose of this paper is to examine the feasibility of rapid estimation of water content and bulk density values from whole core radiographs by comparing optical film density with gravimetrically determined water contents and bulk densities on sediment cores from Beaufort Sea. A procedure is described for standardizing core radiographs that facilitates the recovery of physical property estimates when a few calibration values are available from the sediments of interest.

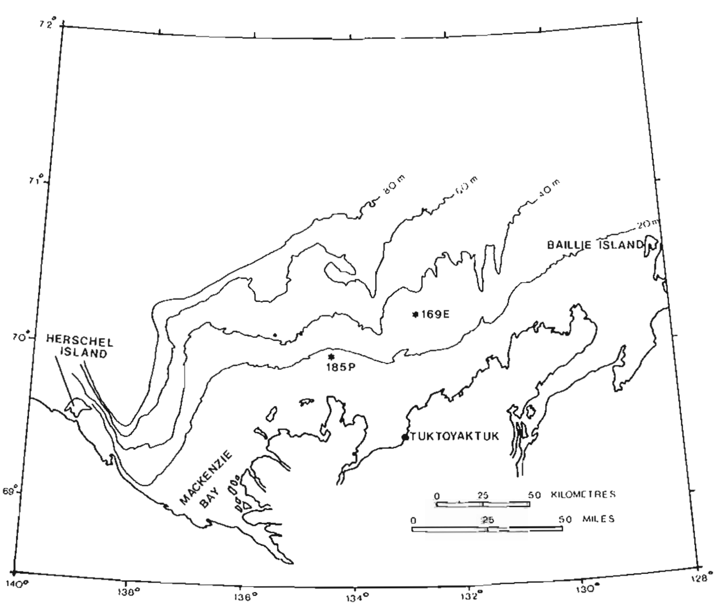

Figure 92. 1. Map of Beaufort Sea showing locations of piston cores 169E and 185P.

From: Report of Activities, Part A;
Geol. Surv. Can., Paper 77-1A (1977)

Coring

In August and September of 1975, several oceanographic cores were obtained in the southern Beaufort Sea from the *M. V. Theta* for studies related to ice scour, sediment dispersal, and Holocene history of the Beaufort Shelf. Three sizes of cores were obtained on the *Theta* '75 cruise. Plastic (CAB) core tubing of 3. 8, 6. 4 and 7. 3 cm outside diameter (OD) was used in a 45 kg Alpine gravity corer, a 550 kg Alpine piston cover, and a 60 kg Benthos gravity corer, respectively.

The unextruded cores usually were cut into 80 cm long segments and shipped south via air freight and rail. They were individually wrapped in plastic film and laid horizontally in waxed cardboard boxes. Ideally, however, cores should be stored and transported upright, handled gently, kept at a temperature of 4°C and at a humidity of 100 percent (Dzwilewski and Richards, 1974). Logistically this was impossible to achieve. Some minor water loss occurred from the cores in the period between collection and radiography[1].

The unextruded cores were X-rayed within two months of their arrival in Ottawa so as to be able to make a fairly rapid preliminary analysis of structures to be found within the first one to two metres of unconsolidated sediment lying on the Beaufort Shelf. As a part of this analysis, water content and bulk density relationships to radiograph film density were determined on two of the 6. 4 cm OD piston cores (Fig. 92. 1).

Radiography

Previous work in X-radiography of soft sediment cores is discussed in Rukavina (1967), Stanley and Blanchard (1967), Bouma and Boerma (1968), Baker and Friedman (1969), Edmondson and Allison (1970), Krinitzsky (1970), and Axelsson and Händel (1972).

Our cores were X-rayed using a Picker Industrial Gemini 160/320 machine with 2. 5 mm focal spot and maximum kilovoltage of 150. The cores were laid horizontally on a table below the X-ray source and were arranged on top of the film cassette. Identification of cores on the film was facilitated by the use of lead letters and numbers. Overlap marks were made on the cores if a core segment did not fit within the width of the 35. 6 x 43. 2 cm film. Welding rods of about 1. 6 mm diameter were laid across core markings to make the overlap locations and centre-line of the radiograph visible on the negative. An arbitrary azimuth was assigned to each core section and maintained in all exposures.

For the 6. 4 and 7. 3 cm outside diameter cores, a satisfactory image was obtained on Kodak AA industrial

[1]In a two-month experiment carried out by the authors it was found that only 1% of the water in a 200 cm^3 section of water-filled CAB tubing (unwrapped) was lost at room temperature.

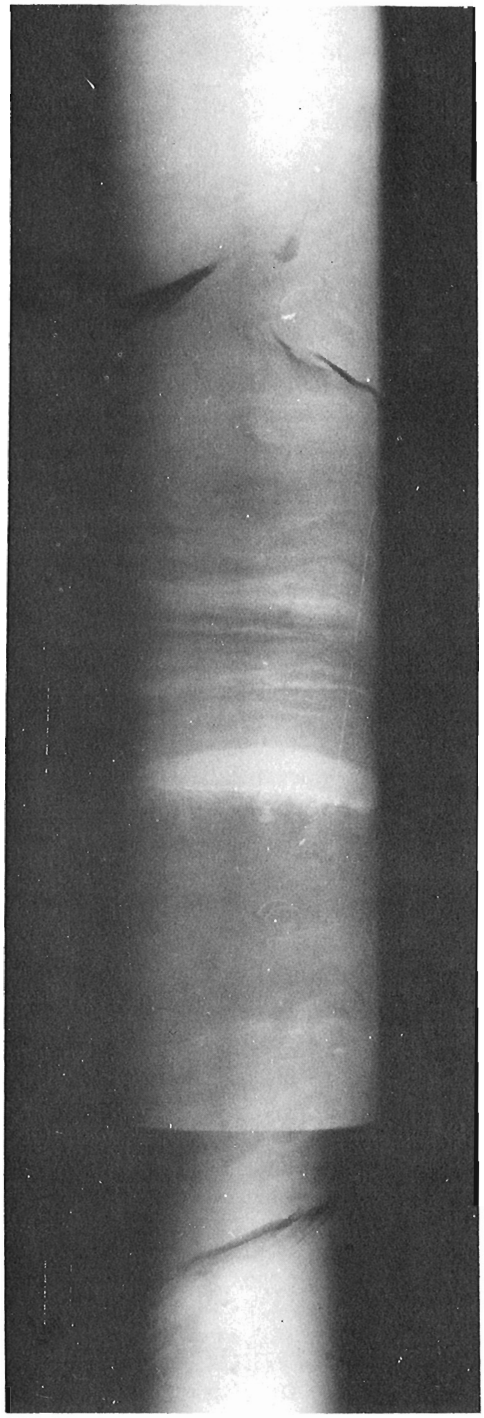

Figure 92. 2. Core X-radiograph showing the uniform exposure across the core width achieved by using an aluminum filter.

X-ray film using a 35 second exposure at 10 milliamperes (ma) and 90 kilovolts (KV). The focal distance was 122 cm. The 3. 8 cm outside diameter cores were exposed for 25 seconds at 10 ma and 60 KV using the same focal distance. Optimum exposure values are best determined experimentally for a given sediment type,

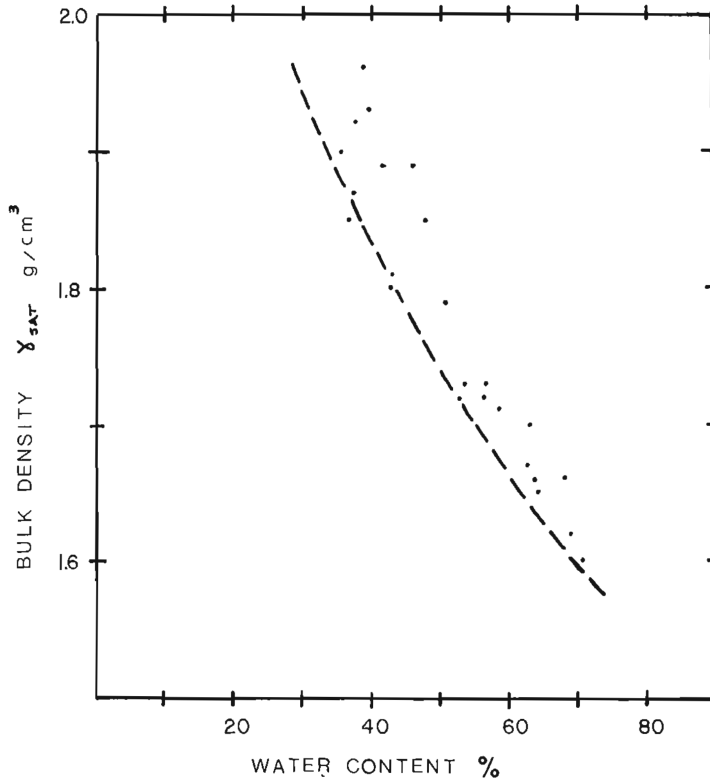

Figure 92. 3. Saturated bulk density versus water content (salt included) for Beaufort Sea cores 169E and 185P. The regression curve is for 1680 marine silt and clay samples from Atlantic and Pacific Oceans (Richards *et al.*, 1974).

core diameter, and radiographic apparatus. Development of exposed radiographs was carried out manually for 5 minutes.

So as to be able to develop useful film density to water content and bulk density relationships, close control had to be kept on other factors that could affect film density. The X-ray machine parameters were kept constant, but these may have been affected slightly by line voltage fluctuations at the time of exposure. The manual development technique left some room for density differences to occur.

Because perfect uniformity in exposures and development could not be achieved, it was decided to include a series of aluminum blocks (step wedge) of varying thickness in each exposure. A radiograph was judged satisfactory if the optical density of the thinnest aluminum block (3. 175 cm thick) was 1. 8±0. 2 optical density units[1]. This range was chosen on the basis that it was representative of the range of densities exhibited on the core images. Densities were measured with a repeatability of ±0. 01 density units using a Macbeth Quantilog densitometer.

[1] optical density, $D = \log \frac{I_O}{I_T}$
where I_O is incident light intensity
I_T is transmitted light intensity.

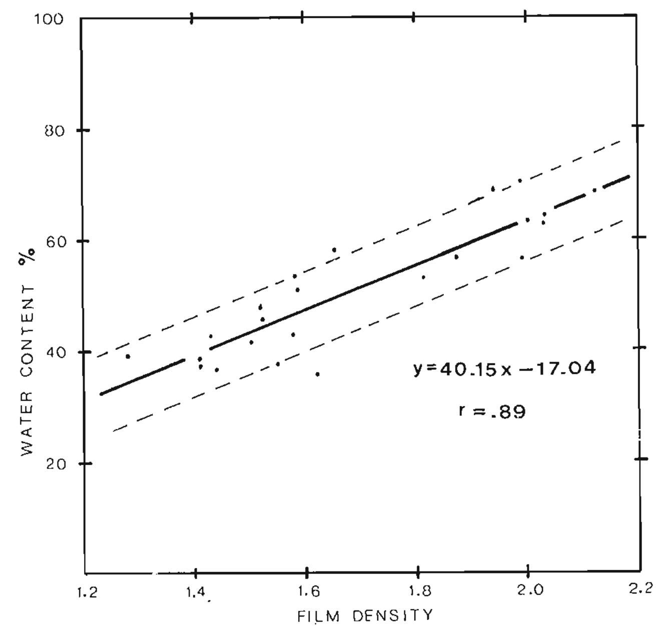

Figure 92.4. Water content (% dry weight) versus X-radiographic film density for Beaufort Sea silt and clay (cores 169E, 185P in CAB tubing of 6.4 cm OD). The standard error about the Y-axis intercept is shown.

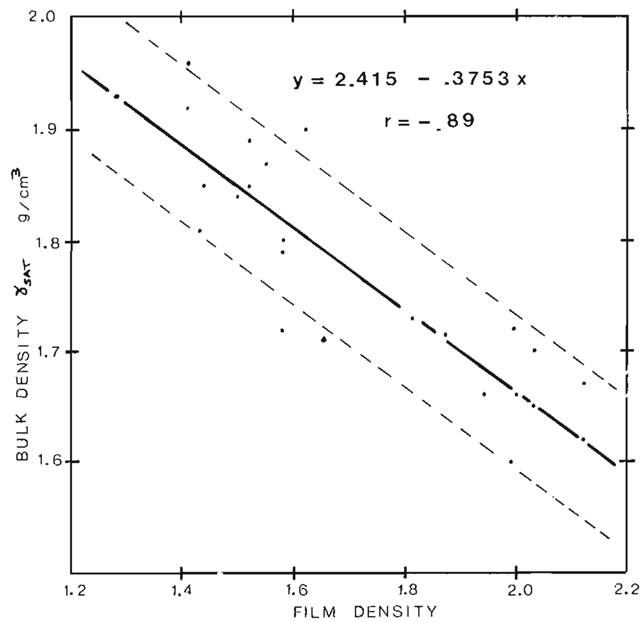

Figure 92.5. Bulk density (saturated) versus X-radiographic film density under the same conditions as in Figure 92.4.

Correcting Radiograph Densities to a Standard Density

If quantitative interpretations are to be made from film density measurements on several different radiographs, it is necessary to standardize the densities. Only rarely will any two radiographs be taken and developed under exactly the same conditions even though uniformity of exposures is being strived for. As long as exactly the same object (e.g., step wedge) appears in each radiograph, however, a correction factor can be calculated.

The absorption of X-rays passing through an aluminum step wedge is given by (Krinitzsky, 1970, p. 145);

$$I_T = I_0 e^{-ux}$$

where I_T is intensity of transmitted radiation
I_0 is intensity of incident radiation
u is the absorption coefficient for aluminum
x is the thickness of the aluminum block

The intensity of incident radiation, I_0, is related to the X-ray tube voltage and current during the exposure. Since the voltage and current values may differ slightly from exposure to exposure, I_0 also must differ. Variations in the film densities between radiographs from this and all other sources were standardized by multiplying the measured film densities on a given radiograph, i, by the correction factor $\frac{Ds}{Di}$. Ds is the standard optical film density (1.8 density units) of the chosen aluminum wedge step, and Di is the measured optical film density of the same wedge step on the given radiograph, i.

Use of X-ray Filter

A common problem encountered in radiographing cores contained in tubes is that the image becomes progressively darker toward the edges. This effect occurs because the X-rays are passing through less sediment along the edges than along the centre of the core tube. A simple remedy is to lay a semi-circular aluminum filter on top of the core to be radiographed (Fig. 92.2). Full details on construction and application of such a filter may be found in Baker and Friedman (1969).

The filter is not absolutely necessary if one just wishes to obtain film densities. These can be satisfactorily measured along the central axis of the core image.

Sediment Properties as Derived From Radiographs

Useful guides to the radiographic interpretation of structures found in unextruded soft sediment cores are to be found in Stanley and Blanchard (1967) and Krinitzsky (1970). In addition to showing structure, an X-ray image will indicate the location of shells and pebbles within the core.

Much work has been done on the problem of nondestructively determining sediment properties, particularly water content and bulk density, within an unopened core tube (Preiss, 1968; Corey and Hayes, 1970; Whitmarsh, 1971; Richards *et al.*, 1974; Meyers *et al.*, 1974; and Chough and Richards, 1974). These workers have used various methods of nuclear transmission densitometry to obtain the required information. The radiographic method is examined here as a possible alternate or complementary source of geotechnical information to standard methods such as nuclear transmission densitometry. The film density method, if sufficiently developed, should have unique benefits since both structural and geotechnical property information are recorded in the same place, that is, on one sheet of radiographic film. These two types of information therefore can be more readily related.

Film Density Relation to Water Content and Bulk Density

After radiographing two Beaufort Sea piston cores (169E and 185P, each 6.4 cm OD), the optical film density was measured along the central axis of the radiographic core images. The cores were then split to measure water content and bulk density conventionally (Gardner, 1965; Blake, 1965) with the estimated analytical precisions being ±0.1% and ±3.5%, respectively. A graduated syringe-body was used to recover known volumes of sediment (1 to 3 cm^3) from the freshly opened cores. Wet sediment weights were measured immediately; dry sediment weights were measured after heating to constant weight at 110°C.

In this report water content is defined as:

$$W(\%) = \frac{\text{weight of water in sediment x 100}}{\text{dry weight of solid particles (including salt)}}$$

and bulk density for a saturated sediment is defined as:

$$\gamma_{SAT}(g/cm^3) = \frac{\text{weight of wet sediment}}{\text{volume of wet sediment}}$$

The Beaufort physical parameters are compared to 1680 Atlantic and Pacific silt and clay samples (Richards *et al.*, 1974) in Figure 92.3. The Beaufort sediments clearly define a distinct relationship between bulk density and water content, as expected, but one that parallels and is slightly offset from the data based on southern ocean samples. The offset may be due to a biased sample, to changes induced in the sediment during transportation and storage, or the Beaufort sediments may have a significantly greater effective grain density thereby giving them greater bulk density for a given water content. Other measurements (not shown) based on sandy sediments and bottom sediments packed into core tubes extended the range of the core sample points but deviated widely from the indicated relationship. Since the principal interest is the interpretation of Beaufort silt and clay core radiographs, the comparison with film density was based on the 24 samples taken from piston cores 169E and 185P. The following analysis is based on these two 6.4 cm OD cores.

The calibration curves (Figs. 92.4 and 92.5) show linear regression lines for the relationships of water content (percent dry weight) and bulk density of a saturated sediment to optical film density. The relationships show considerable scatter for both variables. Water content is predicted from film density with errors

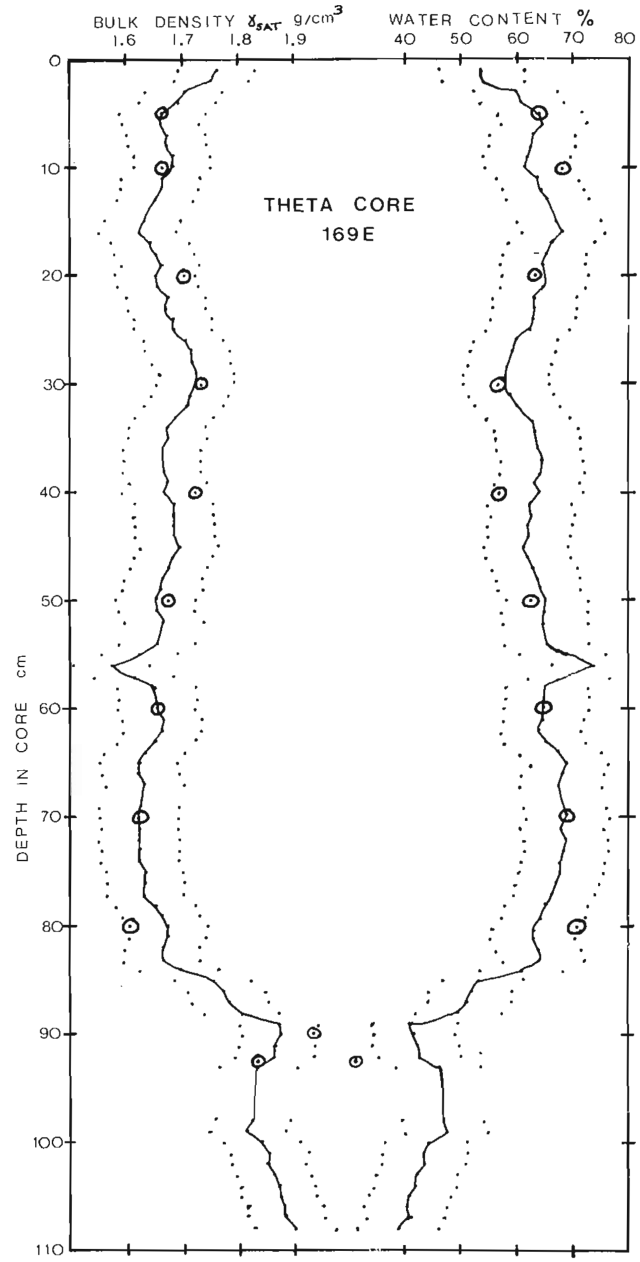

Figure 92.6. Saturated bulk density and water content profiles for Beaufort Sea core 169E as estimated from X-radiographic film densities. The pairs of dotted lines approximate one standard deviation about the estimates. Values obtained by conventional gravimetric methods are indicated by circles. The upper 5 cm are finely laminated sands hence the decreased water content and increased bulk density. Similar changes below 85 cm are also due to sand strata. The peak at 56 cm reflects a cavity due to worm tubes and slight separation of core segments.

of ±7.4% (σ) or ±15% (2σ). Bulk density similarly is predicted with errors of ±0.07 g/cm^3 (σ) or 0.14 g/cm^3 (2σ). Both sets of errors are based on the standard deviations of the y-axis intercepts of the pertinent regression lines (Figs. 92.4 and 92.5). This scatter is likely due to beam divergence variations in X-ray intensity across the film sheet and to uncontrolled changes in core thickness, structure, or other sediment attributes such as particle size, shape, mineralogy, etc., that may influence X-ray attenuation and ultimately radiographic film density. The scatter also includes uncertainties of the conventional analyses. In spite of these uncertainties, the precision of the radiographic method is still adequate for detecting major relative changes in water content and bulk density, particularly in cases where other more precise methods are unavailable or cannot be applied because of changes in the cores during storage. This assumes that cores were radiographed shortly after being collected.

These calibration curves can be used with any sediment from the southern Beaufort Sea falling within the silt to clay range of particle size. The relationships appear to break down when a sediment contains a high percentage of sand.

The results were tested further by comparing the W and γ_{SAT} values obtained conventionally versus those values obtained indirectly using film densities. It was statistically shown that both sets of values are the same at the 97% confidence level for the two methods. In these tests the two sets of values were regressed, and the slope of the regression line was shown to be not significantly different from unity at the 97% confidence level.

A major advantage of the film density method is illustrated in Figure 92.6. The γ_{SAT} and W curves shown are derived from film densities measured 1 cm apart along the image of core 169E. Such close spacing gives a smooth curve for resolving relative γ_{SAT} and W changes within the sediment. Continuous curves could be generated easily with the use of automatic recording densitometers.

Improvements in Procedure

Radiography of cores aboard ship as soon as they are collected is the ideal procedure. Storage, transportation, and water loss problems are then not so important. It would be even better to split some cores in the field so that calibration curves for W and γ_{SAT} determinations from film density could be constructed immediately. To avoid distortions in the core images caused by longitudinal parallax effects, the use of moving-silt radiography is recommended (Parker and Meleskie, 1970). A better indication of structure within the core can be obtained if it is radiographed in two orientations at 90 degrees to each other.

Logging of core thickness and development of a suitable correction factor to X-ray attenuation and observed film density for thickness variations may improve the precision of core property estimates.

During the calibration procedure the densitometer used should read a relatively large area on the core image (perhaps a circle 1 cm in diameter) to better represent each region in the core from which a sample is withdrawn for conventional analysis. During this study the densitometer light source aperture was 2 mm. This could have caused some of the scatter in the γ_{SAT} and W relationships to film density. During actual application of the method a narrow aperture densitometer with continuous automatic recording holds great potential for resolving fine detail in sedimentary sequences.

Summary

Radiograph film density measurements of Beaufort silt and clay cores have been useful in developing a procedure for rapid, quantitative analysis of core sediment bulk density and water content. Shipboard radiography is desirable but not essential. Provided radiographs and a limited number of calibration samples are obtained on fresh sediment, radiographic film density analysis will allow the subsequent recovery of continuous water content and bulk density profiles.

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Project 750074

R.A. Klassen and W.W. Shilts
Terrain Sciences Division

Introduction

Over the past two years, research on drift prospecting in the District of Keewatin has been directed towards developing reconnaissance and detailed sampling techniques that can indicate uranium mineralization as well as base metal mineralization. Shilts and Klassen (1976) have presented some preliminary results of the uranium phase for Boothia Peninsula and central Keewatin. The purposes of this report are: (1) to discuss briefly the techniques of sampling and processing samples; (2) to update the uranium dispersal map of the Kaminak Lake area; (3) to show the results of detailed sampling around a site of known uranium mineralization, which also was indicated by reconnaissance anomalies; and (4) to show the results of reconnaissance uranium drift sampling in the Baker Lake area. Figure 93.1 is an index map of the Kaminak and Baker Lake areas.

Techniques

The mudboil sampling technique developed by Shilts (e.g. Shilts, 1973; Ridler and Shilts, 1974) was employed for both regional reconnaissance and detailed sampling. One to two-kilogram samples were collected from the mineral sediment at the centres of mudboils (sorted, nonsorted circles, frostboils) from the bottoms of holes dug by hand to depths of 30 to 60 cm. Organic inclusions and any till with obvious iron or manganese oxide staining were avoided, but a few samples (< 5%) were unavoidably collected from such material or from postglacial marine muds or silty sands on which mudboils also can form.

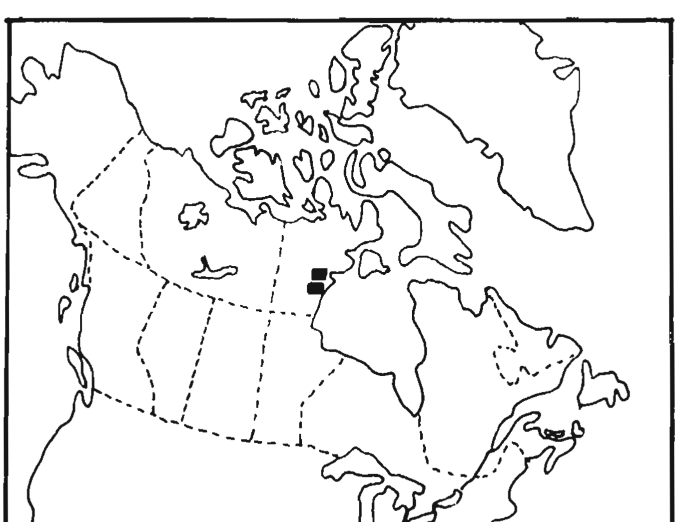

Figure 93.1. Location of Kaminak Lake and Baker Lake reconnaissance grids.

From: Report of Activities, Part A;
Geol. Surv. Can., Paper 77-1A (1977).

Samples from the Kaminak Lake reconnaissance grid were processed either by using a centrifuge to separate the < 2μ fraction from a dried, < 250-mesh separate from till ("dry clay") or by separating the < 2μ fraction from a slurry derived by agitating undried till in a milk-shake machine with a plastic agitator ("wet clay"). In the detailed grid three separates were analyzed for each sample: (1) "dry clay"; (2) "wet clay"; and (3) < 250-mesh (< 64μ) separate derived by sieving of dried, disaggregated till. In the Baker Lake grid, all results are from analysis of < 2μ material separated from the slurry derived from undried till.

Figure 93.2 illustrates the comparative results of these techniques in histogram and scatter-plot form. From the data presented, it appears that the "wet clay" technique (derived from a slurry with no drying of the sample) gives the best background to anomalous value contrasts. In the Kaminak Lake area the distribution of uranium values for "wet clay" samples indicates a principal mode that tends to be higher than that for "dry clay". Analysis of both "wet" and "dry" clays provides a range of values and an areal clustering of high values that can be meaningfully contoured. Both types of clay analysis are evidently superior to those of < 250-mesh fractions which have a much more restricted range of concentrations that are generally close to the analytical detection limit. Although not discussed here, other till fractions, such as rock fragments in the 0.5 to 4 mm size range and sand sized methylene iodide heavy mineral separates, were analyzed for some of the samples discussed above. Too few methylene iodide separates have been done to draw any conclusions other than that their uranium values seem to be slightly higher than clay values in the Kaminak Lake area and slighly lower than clay values in the Baker Lake area. The analysis of the crushed rock fragment fraction gives even less contrast among samples than < 250-mesh separates in these areas. One till sample collected a few metres from uranium (U) mineralization was analyzed at 1280 ppm U for wet clay, 57 ppm U for methylene iodide heavies, and 11 ppm U for rock fragments.

A selected suite of clay separates also was analyzed by neutron activation techniques[1] for comparison with the fluorescence analyses commonly used in this project. These analyses were close to the results obtained commercially[2] by fluorescence analysis, and the authors could not see any outstanding advantage of one method over the other for these types of samples.

Results: Kaminak Lake Area

Figure 93.3 shows the results of reconnaissance scale sampling in the Kaminak Lake area (the sampling was originally designed to cover areas of high base metal potential). None of the results have been

[1]Analyses by Atomic Energy of Canada Limited

[2]Analyses by Bondar-Clegg and Co., Ltd.

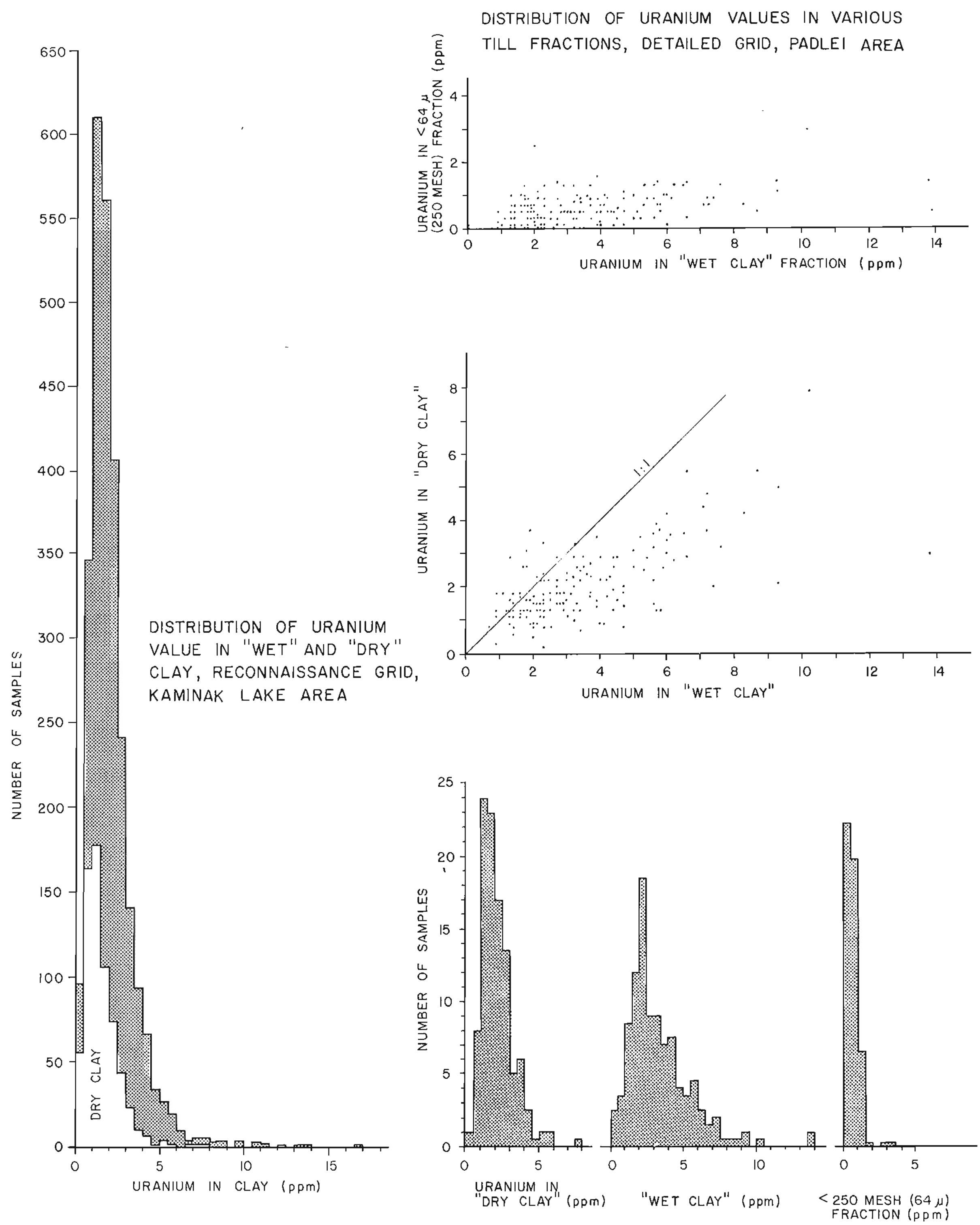

Figure 93.2

Histograms and scatter plots of uranium values from various fractions of till. Shaded part of the reconnaissance grid histogram represents samples done by the "wet clay" technique.

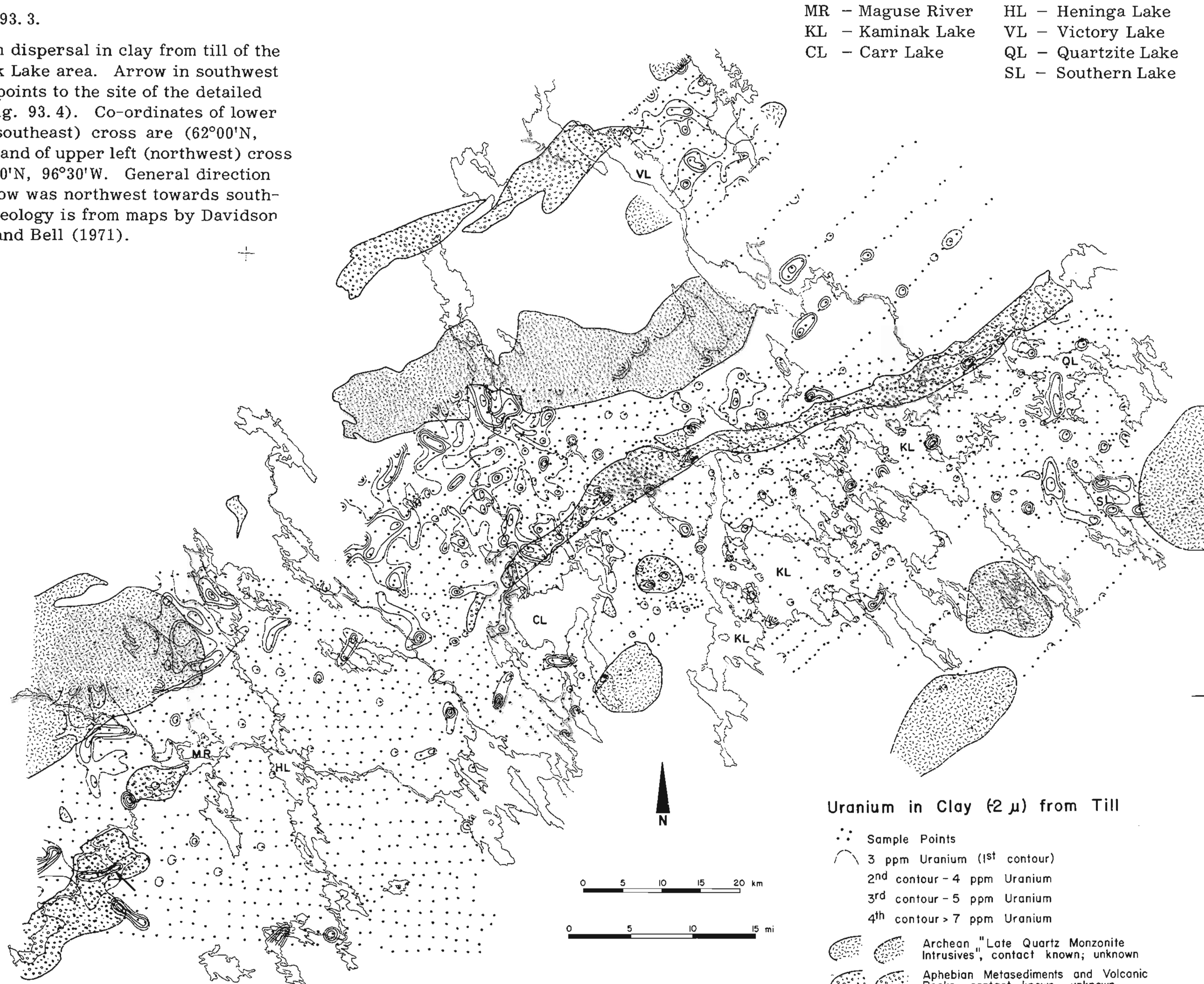

Figure 93.3.

Uranium dispersal in clay from till of the Kaminak Lake area. Arrow in southwest corner points to the site of the detailed grid (Fig. 93.4). Co-ordinates of lower right (southeast) cross are (62°00'N, 94°08'W and of upper left (northwest) cross are 62°30'N, 96°30'W. General direction of ice flow was northwest towards southeast. Geology is from maps by Davidson (1970) and Bell (1971).

"smoothed" statistically; all samples within a contour are within the concentration ranges described for the contour. Most areas of abnormal uranium content can be related to two principal rock types: (1) Aphebian sediments — Hurwitz, Montgomery Lake, and Mackenzie lake groups (Bell, 1971; Davidson, 1970) and (2) plutons comprising porphyritic acid intrusive complexes mapped by both Davidson and Bell as having been emplaced during Late Archean plutonic events (map-unit 3).

Uranium mineralization has long been known to be associated with the Aphebian basal conglomerates and sandstones. The anomalies at the southwestern corner of the map are closely related to several occurrences of uranium mineralization that were described by Bell (1970) in Montgomery Lake metasediments of the Padlei area. Other fairly coherent anomalies occur over the Hurwitz Group immediately north of Carr Lake and on a till plain eastward along the strike of Mackenzie Lake metasediments at the north end of Victory Lake. It should be noted that only some portions of the areas underlain by Aphebian rocks are characterized by till having elevated uranium content; well over 75 per cent of the Aphebian rocks are associated with till having low to background uranium contents. This may mean that uranium mineralization potential in Aphebian rocks is relatively greater in areas with tills of higher uranium content.

An unexpected and as yet unexplained series of large, coherent zones of uranium enriched till are associated with some Archean porphyritic plutons, mainly those that occur in Davidson's (1970) bedrock region B, a zone of amphibolite-grade metamorphism where the porphyritic bodies are intruded into Archean metasediments rather than Archean volcanics. At the northwest corner of the sampled area, strong, coherent anomalies are associated with the Maguse Batholith. Two samples from a hole drilled by Polargas just west of the grid on the batholith yielded uranium concentrations of 27 and 32 ppm, while a sample from a nearby hole was over 6 ppm. The western end of the next major batholith northeast of the Maguse Batholith seems to have been the source for prominent, southeastward trending uranium dispersal trains, particularly down-ice from a large, northwesterly trending lake system that occupies a depression formed by concentrated glacial erosion. A band of uranium-rich till seems to be associated with a porphyritic pluton that strikes northeastward across the south end of Victory Lake and beneath the Victory Lake till plain.

Other lithologically similar batholiths near Carr, Kaminak, and Southern lakes show no comparable development of uranium anomalies. Strong anomalies near Southern and Quartzite lakes are in areas mapped as basic volcanics or basic intrusives, but these anomalies are close to areas known to be cut by swarms of fluorite-bearing pegmatite dykes, apparently associated with the batholith just east of Southern Lake (Davidson, 1970, p. 10).

Detailed Sampling Grid

Figure 93.4 depicts uranium concentrations for the clay fractions of till collected from a grid laid out over eastward striking sandstones and conglomerates of the Montgomery Lake and Hurwitz groups. Surface zones of mineralization could be detected by scintillometer by following the map of Bell (1970). The glacial dispersal appears to be only on the order of a few hundred metres or less, probably because of the small area of mineralization. The principal trends evident are several enriched bands parallel to the strike. Analyses of these samples for other elements indicate that there is a positive correlation between copper and uranium dispersal in these tills, a relationship that is not evident between these elements on the regional grid discussed above. It can be concluded from the map that, although "smearing" of the anomalies is apparently slight or indeterminate, the smearing may be partially caused by solifluction or hydromorphic processes. On the other hand, most displacement observed is in the southeastward glacial direction, whereas the slopes in the grid are facing predominantly to the southwest and, north of the principal uranium occurrence, are facing northward.

Results: Baker Lake Area

A preliminary, contoured map of the uranium content in the clay sized ($< 2\mu$) portion of till in the Baker Lake area is shown in Figure 93.5. This map is based on the analysis of samples collected during July and August 1975. The area sampled during 1976 also is outlined on the map. The history of ice movement in the area is complex as the Keewatin Ice Divide lies immediately northwest, and the direction of ice movement can vary widely. The direction and sense of movement shown on the map is the one most commonly observed. Generally the tills within the eastern portion are brown and have a sandy silt matrix; those of the western portion are red with a silty clay matrix reflecting their derivation from Dubawnt red beds.

The map indicates that there are four clusters of samples with elevated uranium levels. There is neither an obvious nor a consistent relationship between the uranium levels and either specific bedrock types or sites of known uranium mineralization.

The reasons for the lack of correspondence of uranium levels to mineralization and the generally low uranium background levels in the Dubawnt Group rocks are puzzling but probably explicable in terms of bedrock geochemistry. Mineralization may comprise secondary deposits isolated in fractures or beds in sediments and volcanic rocks that are otherwise impoverished in all metals, including uranium. If this is a general case, the source areas would be exceptionally small with respect to sample spacing, and even the weak tails (Shilts, 1976) of the dispersal curves would be difficult to detect.

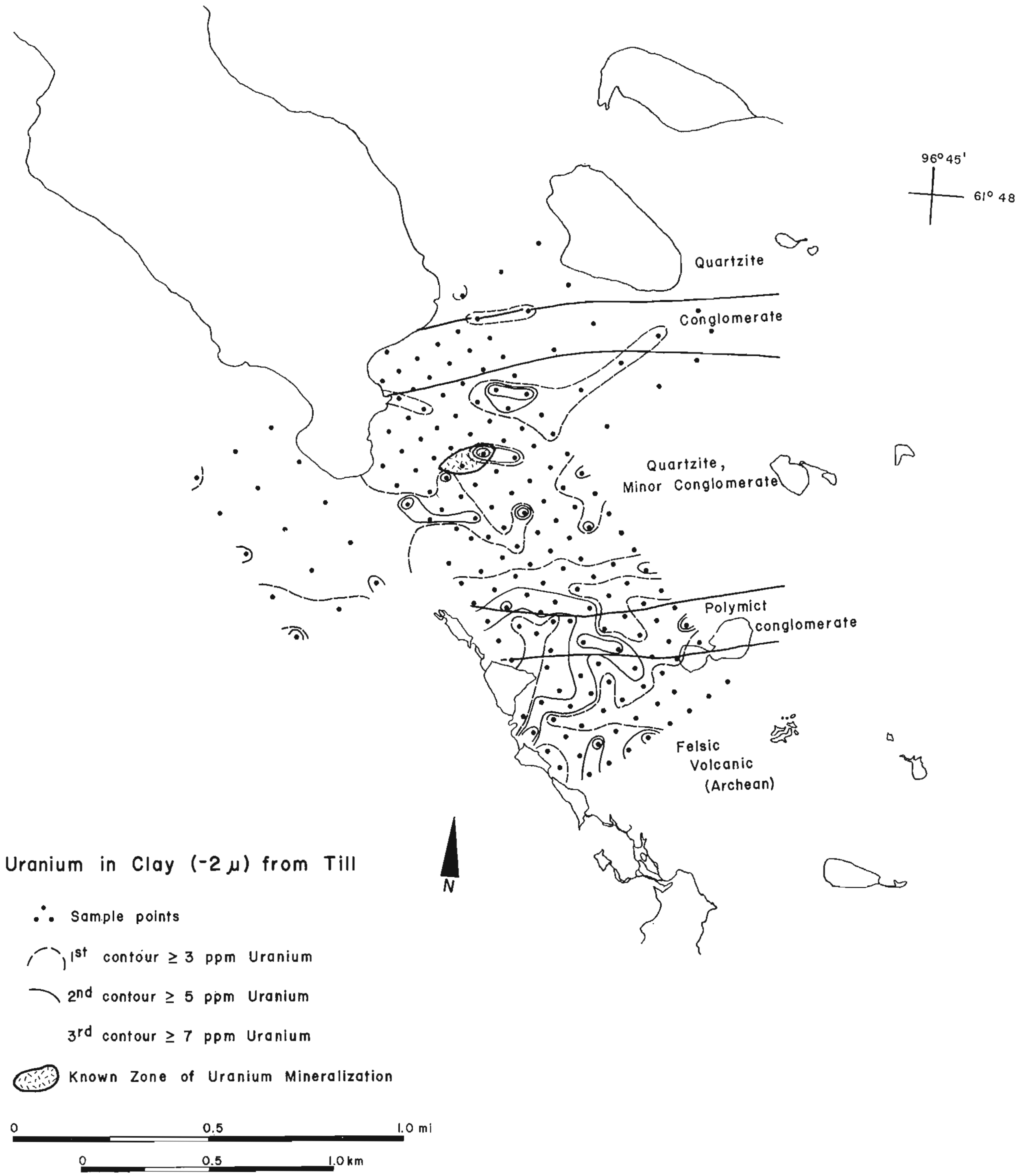

Figure 93.4. Uranium in clay from detailed grid (*see* Fig. 93.3 for location). Geology after Bell (1970).

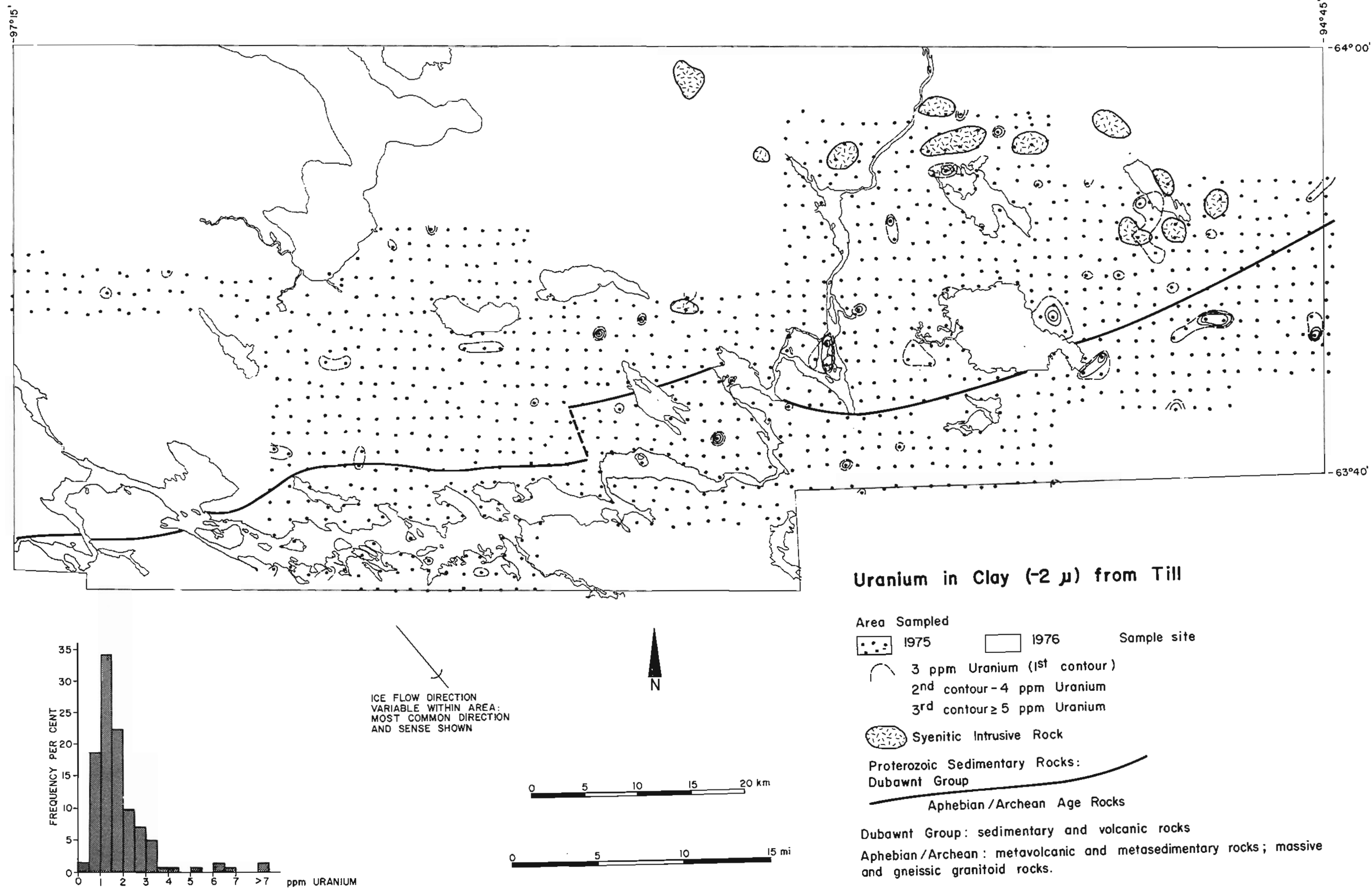

Figure 93. 5. Uranium in clay from till in the Baker Lake area. Geology after Donaldson (1965).

The contrast of the uranium concentration patterns for till in the Baker Lake area with those of the Kaminak Lake area is striking. The mineralization in the Kaminak area may be dispersed at low levels through a large volume of bedrock rather than consisting of small, discrete zones of uranium-enriched bedrock as in the Dubawnt Group. Alternatively many small, enriched zones in the bedrock may be closely clustered, acting like a larger source of dispersed mineralization. It is presently an imperfectly explained paradox that the area of the Keewatin Shield that is presently one of the most attractive for uranium exploration, the Dubawnt Group rocks of the Baker Lake area, gives a weak drift geochemical response for uranium.

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DETAILED DRIFT PROSPECTING IN THE SOUTHERN DISTRICT OF KEEWATIN

Project 700014

R.N.W. DiLabio and W.W. Shilts
Terrain Sciences Division

Introduction

In 1976 till was collected from mudboils on grids with closely spaced sample points near four occurrences of base metal mineralization and one occurrence of uranium mineralization. The purpose of the sampling was to map glacial dispersal patterns from known sources and to confirm and define by detailed sampling anomalies that were found by reconnaissance sampling on one mile grids (Shilts, 1974, 1975; Shilts and Klassen, 1976). Sampling density was between 5 and 100 samples per square kilometre. Preliminary results for samples collected near the former nickel-copper mine at Rankin Inlet are presented below.

The bedrock geology of the Rankin Inlet area (Fig. 94.1) and the nickel-copper deposit of North Rankin Nickel Mines Ltd. have been described by Drybrough (1931), Pelzer (1950), Bannatyne (1958), and Laporte (1973). The ore deposit constituted a basal layer in the southerly dipping serpentinite. In order of abundance, the metallic minerals in the ore were: pyrrhotite, pentlandite, chalcopyrite, magnetite, pyrite, violarite, marcasite, and gersdorffite (?) (Pelzer, 1950). Thin lenses of massive sulphides at the base of the serpentinite were overlain by thicker zones of disseminated sulphides. The mineralized zone was about 600 m long by up to 30 m wide.

In the north part of the sampled area outcrops are rare; only gabbro is well exposed. In the south part of the area serpentinite and mafic volcanic rocks are well exposed. Till and colluvium derived from till are the main surficial sediments; fossiliferous marine beach gravels and marine silts lie on steep slopes and in valley bottoms, respectively. All surficial sediments except beach gravels have mudboils developed on them. The direction of the last glacial flow was towards the southeast. These observations are in agreement with those made by James (1972).

Results

Abundances were determined for eight elements in the $< 2\mu$ (clay) fraction of 88 till samples. Nickel, copper, chromium, cobalt, and iron are significantly concentrated down-ice from the serpentinite outcrop (Figs. 94.1 to 94.4 and Table 94.1). The high Ni, Co, and Cr contents of the till at Rankin Inlet are comparable to those found in the $< 2\mu$ fraction of till in other areas of ultrabasic bedrock, such as the Appalachians of Quebec (Shilts, 1976), and in some areas of presumed basic volcanic bedrock in the southern District of Keewatin.

Although the area is one of high background for most of the metals, the mapped pattern of high and low abundances is orderly and consistent (Figs. 94.1 to 94.4). Where the till overlies tuff, quartzite, or gabbro, the metal contents are relatively low. On and down-glacier from serpentinite and mafic volcanic rocks, the metal contents are relatively high. The contact between areas of low and high abundances is sharp, and the contrast between the areas is high. This is reflected in the frequency distributions of Ni and Cu, which are polymodal (Figs. 94.1 and 94.2).

Discussion

The fact that the areal distribution of high and low abundances of each of the metals is similar (Figs. 94.1 to 94.4) means that the distribution of the metals is controlled by similar causes. The polymodality of the frequency distributions (Fiss. 94.1 and 94.2) is inferred to mean the presence of more than one population. The population of till samples with relatively high metal concentrations is interpreted to contain a significant amount of material eroded from the serpentinite and mafic volcanic rocks whereas the population of till samples with relatively low metal concentrations is interpreted to reflect average values for debris eroded from quartzite, tuff, and gabbro.

Other Areas of Cu-Ni Enrichment

Near Padlei, approximately 150 miles southwest of Rankin Inlet, within the Rankin Inlet-Ennadai greenstone belt, Shilts (1974) has reported a significant area of till that was found by reconnaissance sampling to be enriched in copper and nickel. An area within the zone of enrichment was sampled in detail in 1975. The bedrock under the detailed site consists of basic volcanic rocks intruded by a pluton composed of rocks of acidic to intermediate composition. The detailed sampling confirmed the presence of Cu-Ni-rich till (Fig. 94.5) and seemed to outline several areas of particularly high Cu and Ni contents in the till (Fig. 94.6 and 94.7). The high ratio of Cu-Ni to other base metals is uncommon in the part of the Rankin-Ennadai belt that has been sampled, and the relative trace element contents of till from this area are similar to those of the Rankin Inlet ultrabasic train. Relatively high Mg, Co, Cr, and Fe contents, all indicative of till derived from very basic or ultrabasic rocks, also are associated with till samples with high Cu and Ni contents. Brief reconnaissance of the bedrock in the vicinity of the anomalies has turned up many small gossan zones, but detailed mapping by Bell (1970) and by Barrett and Leggett (1976) has failed to reveal any ultrabasic sources of sulphide zones of any appreciable extent.

The nickel values in this detailed grid are very high for the Kaminak-Heninga Lake area (*see* threshold values indicated in Fig. 94.5) but are considerably lower than those noted near the ore body at Rankin Inlet. The copper values, however, are generally as

From: Report of Activities, Part A;
Geol. Surv. Can., Paper 77-1A

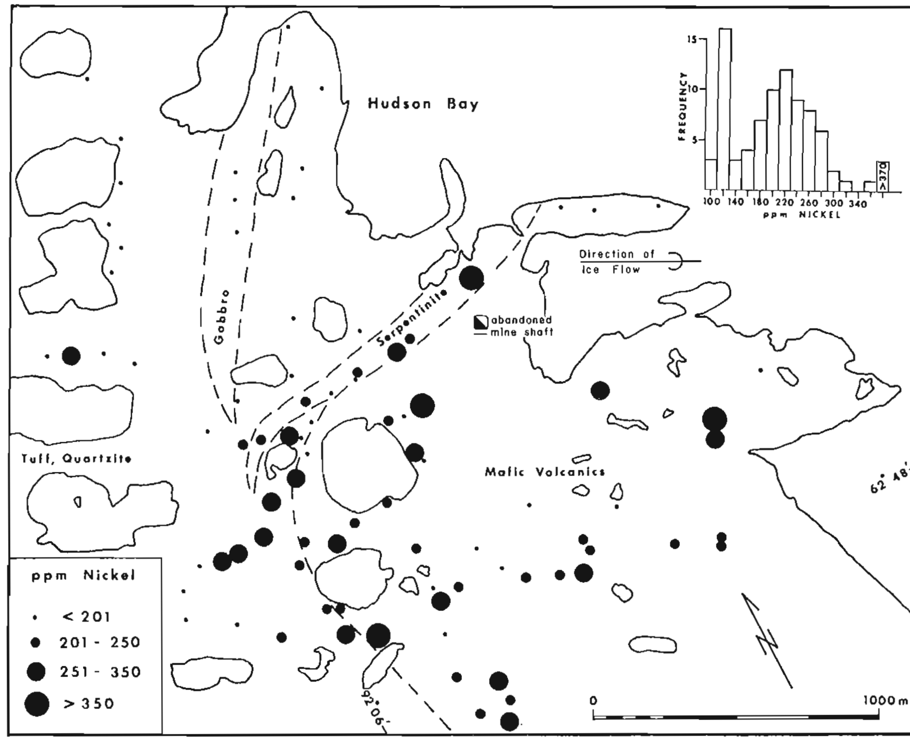

Figure 94.1.

Nickel in the <2μ fraction of till samples from mudboils in the Rankin Inlet area.

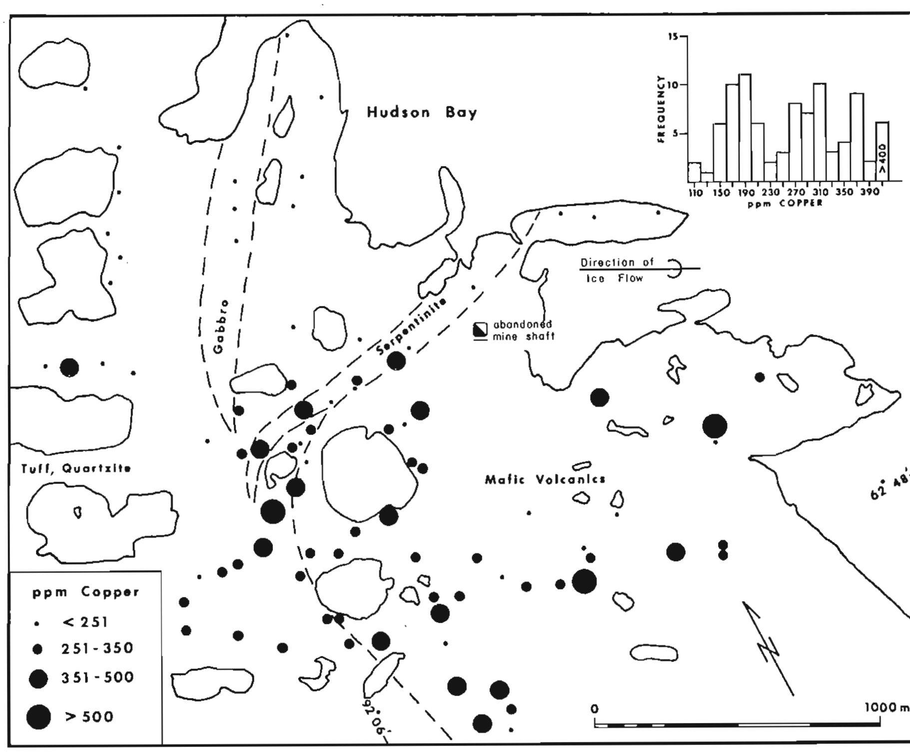

Figure 94.2.

Copper in the <2μ fraction of till samples from mudboils in the Rankin Inlet area.

high or higher than those in the till of Rankin Inlet. It should be noted here also that comparably high Cu-Ni values were found in several till samples collected during terrain mapping projects around the southeast side of Meliadine and Peter lakes, approximately 20 miles northwest of Rankin Inlet, in an area of unknown bedrock.

These data are presented here because of their similarity to data from Rankin Inlet, and because the area they represent is associated with bedrock that is geologically similar to that of Rankin Inlet. Although the bedrock in the area has been mapped in detail, large portions (more than 50 per cent) of it are covered by drift, which may obscure the true source(s) of the Cu-Ni enrichment.

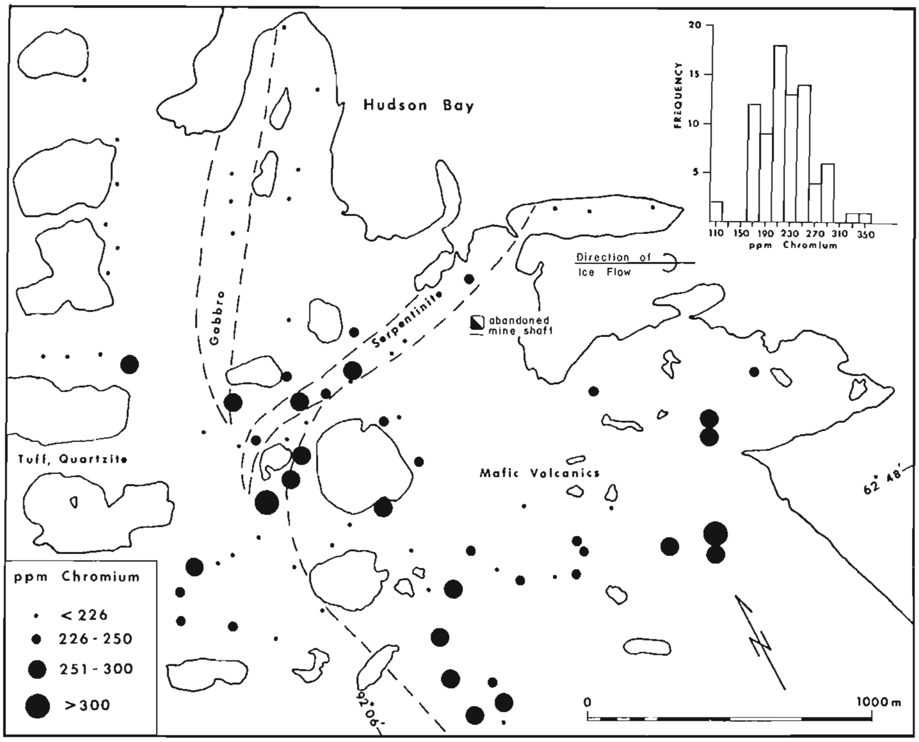

Figure 94. 3.

Chromium in the <2μ fraction of till samples from mudboils in the Rankin Inlet area.

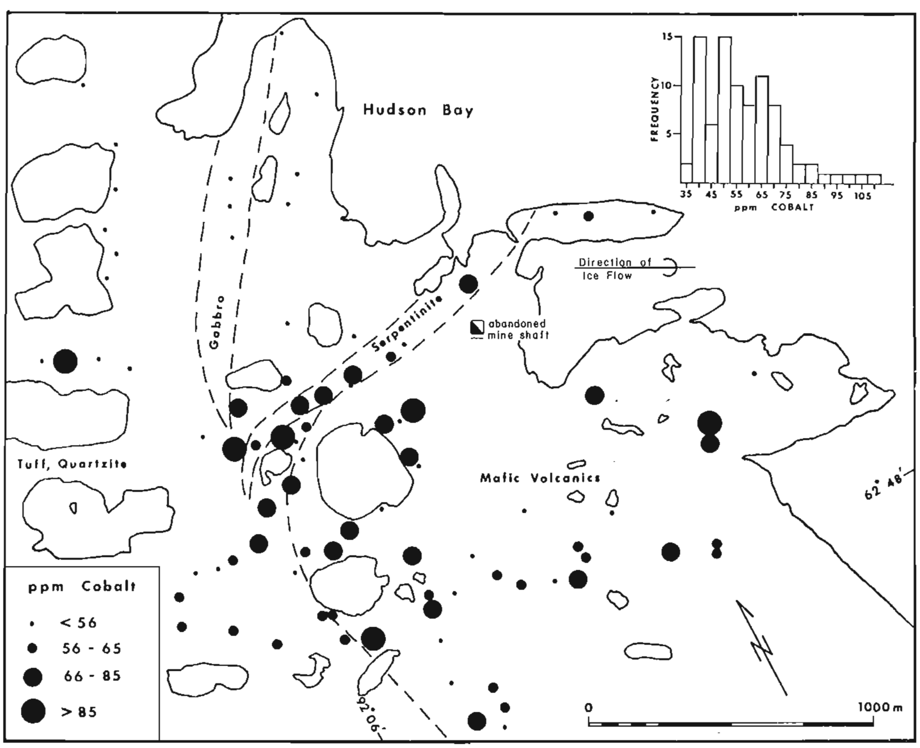

Figure 94. 4.

Cobalt in the <2μ fraction of till samples from mudboils in the Rankin Inlet area.

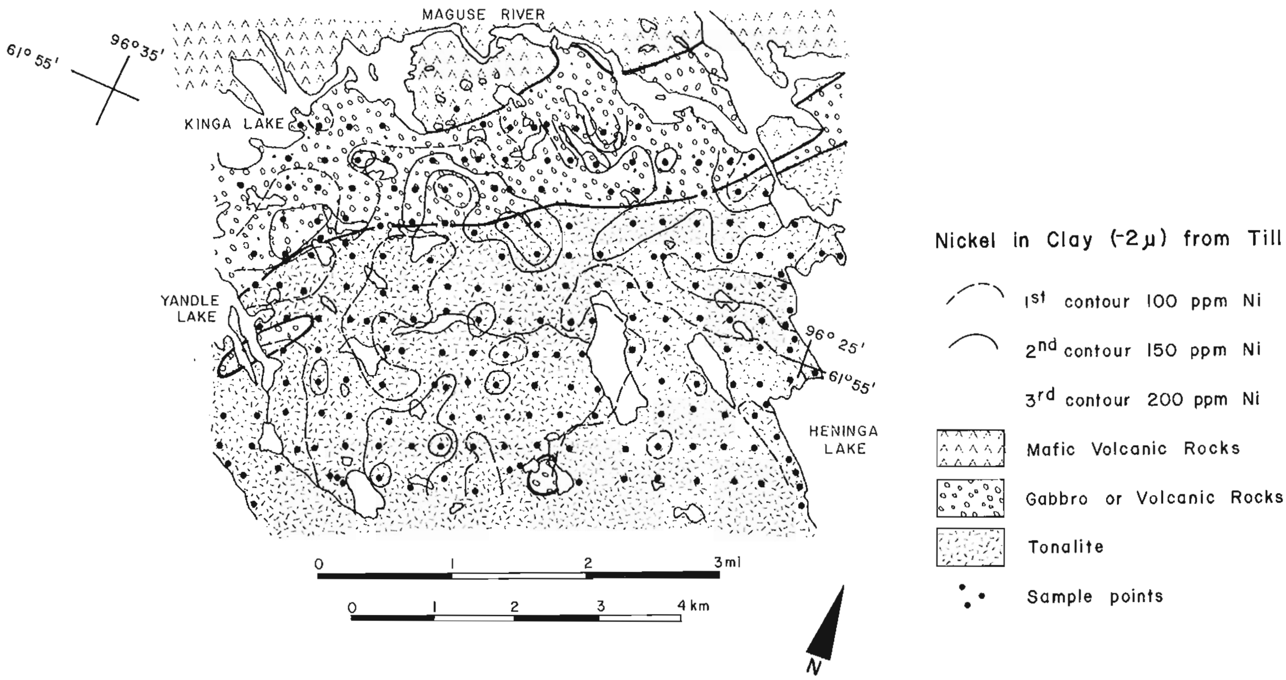

Figure 94. 5. Nickel concentrations in clay from till, Padlei area. The basic volcanic rocks are presumed to be Archean in age and apparently are overlain by Aphebian quartzites less than 2 km west of this grid. Geology modified from Barrett and Leggett (1976) and Bell (1970).

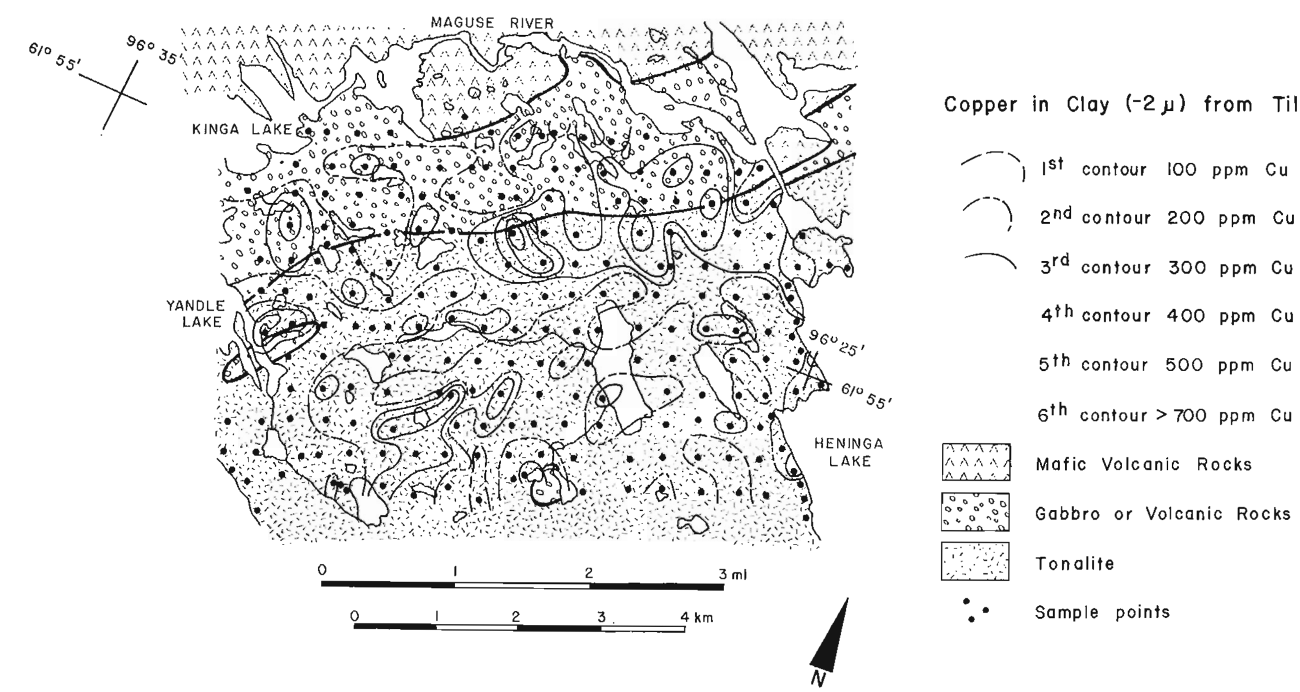

Figure 94. 6. Copper concentrations in clay from till, Padlei area. Outcrops with small amounts of sulphides and malachite staining were reported near the strongest anomaly, but rock chip samples were lost in shipment.

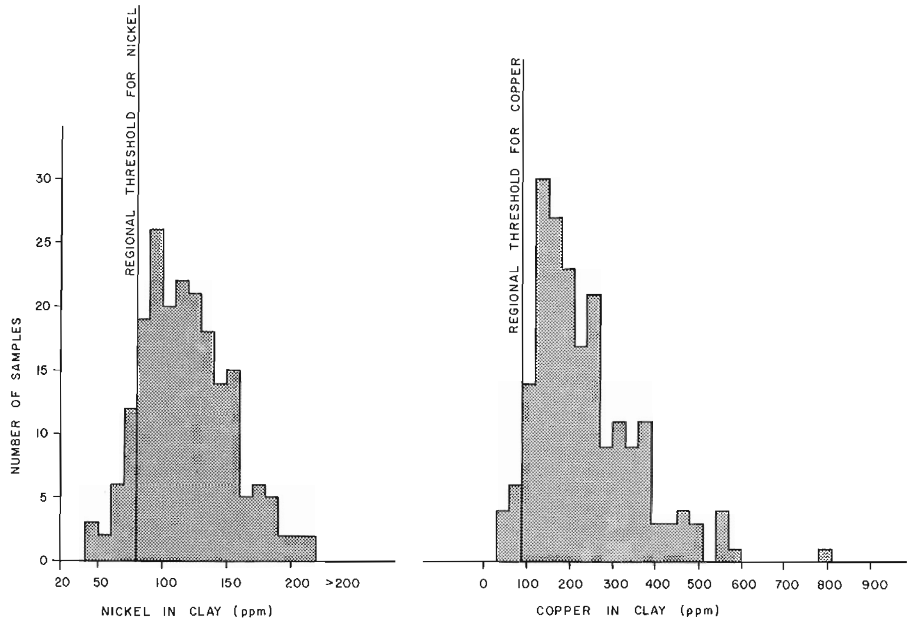

Figure 94.7.

Histograms of Nickel and Copper concentrations in clay, Padlei area. Note that most concentrations are above regional background.

Table 94.1

Range and median abundances for eight elements in the < 2µ fraction of till samples from mudboils in the Rankin Inlet area

Element	Range (ppm)	Median (ppm)
Cu	108 - 580	278
Pb	119 - 50	29
Zn	96 - 186	143
Co	33 - 112	56
Ni	94 - 1840	203
Cr	108 - 351	217
Mn	440 - 1120	655
Fe (in per cent)	3.01 - 7.28	5.96

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A PRELIMINARY ACCOUNT OF SURFICIAL MATERIALS, GEOMORPHOLOGICAL PROCESSES, TERRAIN SENSITIVITY, AND QUATERNARY HISTORY OF KING CHRISTIAN AND SOUTHERN ELLEF RINGNES ISLANDS, DISTRICT OF FRANKLIN

Project 760010

D. A. Hodgson
Terrain Sciences Division

Introduction

This report outlines some of the basic data necessary for land management in the study area. The impetus has been provided by discoveries of natural gas at a number of sites on King Christian Island and both onshore and offshore of southwest Ellef Ringnes Island.

Dominant material-genetic units and landforms (Hodgson, 1975) are described and are displayed in Figure 95.1. The small scale of presentation requires much generalization of units, particularly in the case of Holocene coastal plain sediments. The descriptive units, together with observations on active geomorphological processes, form a basis for a discussion on the susceptibility of the terrain to man-induced disturbance. Finally, some observations on the Quaternary history are reported.

Field work for 1976 was intended to be based on traverses up to 50 km long, to be run by Honda A.T.C. 90's towing trailers carrying a shallow drill and fly camp equipment. Extensive snow cover until mid-July and subsequent slow drying-out of the active layer under generally overcast and cool conditions, however, restricted ground traverses to within 10 km of three camp locations (Fig. 95.1). Thirty hours of helicopter time used in the field area partially overcame transport difficulties. Field observations were made of: (a) landforms, materials, processes, and vegetation at surface sites, with pits being dug to the frost table; (b) stratigraphy, and particularly ice content of cores obtained by drilling to a maximum 2.5 m depth with a CRREL-type auger; and (c) natural stream cuts — of which a surprising number were present in coastal plain sediments.

Material-Genetic Units

A division of the landscape into areas exposed to Quaternary marine processes and areas at higher elevations that were not affected is evident. The boundary roughly follows the upper limit of Holocene marine deposition. This limit subsequently has been uplifted to ca. 40 to 50 m above present sea level on King Christian Island and the Meterologist Peninsula of Ellef Ringnes Island, declining northwards to ca. 30 m at Dome Bay.

In this section two items should be noted: 1) colluvium is not identified because slope processes do not significantly change the material-genetic units and 2) morainal material (including till) was not identified anywhere in the study area.

Above the Marine Limit

The landscape is controlled by the bedrock — a succession of Mesozoic and Cenozoic clastic strata, generally deformed into northwest trending broad folds, with major anticlines commonly cored by diapirs. The surface is formed of scarplands, rolling plains, and low plateaus, slightly to moderately dissected by drainage lines which are essentially normal to the coasts. Slopes are less than 5 degrees other than on scarps, immediately adjacent to drainage lines, or where diapirs are present. Surficial material is composed of residual weathered bedrock; rare outcrops of resistant strata occur.

The following map-units are based on rock-stratigraphic units, which in general have a fairly uniform lithology. Boundaries follow the stratigraphic unit boundaries established by Stott (1969) for Ellef Ringnes, with the exception of the Hassel-Christopher contact which follows the revision suggested by Balkwill (1973), and by Balkwill (1974) for King Christian Island. A further description of landforms and processes on Ellef Ringnes Island is given by St-Onge (1965).

Diapirs (Co): independent, commonly dome shaped massifs, visually prominent because of relief of 50 to 200 m and a high albedo. The cores are formed by gypsum, anhydrite, and (at depth) halite, with the Carboniferous Otto Fiord Formation as the probable source (Davies, 1975). Also present are minor intercalated limestones and gabbro and basalt intrusions. The strata of peripheral formations are steeply inclined on the margins of the dome.

Where diapiric rocks are exposed, the surface materials are chiefly large crystals to grains of gypsum, solution-pitted outcrop, and minor limestone and intrusive rock rubble. The surface is commonly highly dissected by fluvial processes and solution, with slopes greater than 10 degrees. Peripheral formations are included in the diapir unit where slopes are steep and partially overrun by talus of diapiric rocks. Hoodoo Dome is exceptional in that it still has a cover of Isachsen sandstone.

Deer Bay Formation (JKd): a dark grey to black papery silty shale, which contains increasing amounts of fine grained sand towards the top of the unit. It is only present at the surface on King Christian Island, and there it is largely reworked or covered by marine

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Danish Strait
Dome Bay
KING CHRISTIAN ISLAND
NOICE PENINSULA
METEOROLOGIST PENINSULA
ELLEF RINGNES ISLAND
Jackson Bay
Cape Thorstein
Thor Island
Campbell Pt
Cape Allison
Elve Pt
Macklin Point
Sutherland Point
Hoodoo Dome
Isachsen Dome
Kilometres
N

LEGEND

Fd	Deltaic-marine sediments	R	Rock, residual weathered bedrock
Fp	Fluvial sediments	RW	Marine-washed bedrock
W	Marine sediments, undifferentiated		Superscripts:
Wb	Beach sediments	Ke_1	Eureka Sound Formation – upper member
W/R	Marine sediments, thin or discontinuous over rock	Ke_2	Eureka Sound Formation – lower member
		Kk	Kanguk Formation
TQ	Late Tertiary or Quaternary gravels	Kh	Hassel Formation
	TQ_1 or * Flat lying or knobby	Kc	Christopher Formation
	xxxx Ridged	Ki	Isachsen Formation
		JKd	Deer Bay Formation
⊗	Camp site	Co	Diapir (source in Otto Fiord Formation)

Where the areas of two or more map-units are too small to be delineated separately at the map scale, compound units are used. Components of compound units are listed in order of decreasing area.

Figure 95.1. Surficial material-genetic map-units, King Christian and southern Ellef Ringnes islands.

sediments. The surficial material above the marine limit is an impermeable clayey silt, which is commonly poorly drained, with fine sandstone beds near the contact with the Isachsen Formation.

Isachsen Formation (Ki): mainly fining upwards cycles of poorly to noncemented fine to coarse grained sandstone, with minor siltstone, shale, and coal. The lower part of the formation is massive quartzose sandstone, locally well cemented. Surficial material is dominantly loose sand with strike-aligned bands of silty sand, silt, and clayey silt where cyclic successions of inclined sediments are exposed. Although generally noncemented (weathered?) to at least a depth of 2 m, a discontinuous cover of cemented sandstone fragments, probably of lag origin, is present.

The landscape is rolling, fluvially dissected, with a succession of minor scarps. Scarps and steep slopes in well cemented sandstone are locally 50 m high. Mesas, buttes, and hoodoos are common where flat lying to gently inclined sandstone is exposed on valley sides or coastal slopes. Stream channels are broad and are composed chiefly of sand. The unit is generally well drained.

Christopher Formation (Kc): chiefly shale, divided by Balkwill (1974) into two informal members, the lower being more silty than the upper, separated by an interval of glauconitic sandstone. Surficial material is silty clay, which has a fine granular to blocky structure to a 50 cm to 2 m depth, and is underlain by platy shale fragments. There are small areas of sandstone and mudstone lag cover, and local concentrations of shattered mudstone and ironstone nodules occur in stream courses.

The landscape is rolling, with the slightly more resistant intervals coring strike-aligned ridges; it is more rounded, though not topographically lower, than adjacent sandstone formations. Drainage is generally poor, with some wet sedge/moss areas. The thixotropic clay forming the active layer over much of the unit did not dry out in the summer of 1976.

Hassel Formation (Kh): a succession of generally poorly cemented, brightly coloured, fine to coarse grained sandstones, with minor siltstone, shale, and coal beds. The surficial material is chiefly sand, over poorly or noncemented bedrock; there are minor outcrops of well cemented sandstone or siltstone and narrow strike-aligned bands of finer material. The landscape is composed of gentle slopes and a succession of minor scarps, with some local dissection. It is more subdued than the Isachsen Formation as there are proportionately fewer resistant beds. Drainage is generally good.

Kanguk Formation (Kk): black silty shale, with minor beds of light weathering bentonite near the base of the formation. Surface material is partially weathered bedrock composed of silt, clay, and platy fragments of shale. A discontinuous lag cover of shattered brown ironstone nodules and siltstone and sandstone fragments is conspicuous on level areas. The shale is highly acidic, and the one core for which pH values are available provided a reading of 3.6 at both 20 cm and 150 cm depths, i.e. above and below the 25 cm-deep frost table (*see* also Balkwill and Hopkins, 1976).

The dominant relief feature is an almost continuous escarpment, controlled by siltstone beds which divide the formation into upper and lower members. The scarp cliff is commonly 20 m and locally 50 m high and is broken only by water gaps. A second scarp developed near the base of the formation is discontinuous, but where present it is just as prominent as the first. Most of the unit varies from overall gentle slopes to extensive level areas where beds are flat lying. Fluvial dissection is locally significant, with steep sided gullies advancing headwards along the rectilinear pattern ice-wedge troughs which are so distinctive on this formation. The active layer is moderately well drained except on the bentonite (cf. Christopher shale).

Eureka Sound Formation (Ke):

Ke_1 informal lower member: sandstone, well to noncemented, and shaly mudstone. Surficial material is dominantly sand, silt, and minor clay, with bands of outcrop or lag cover of sandstone and mudstone. The map-unit is chiefly one of long gentle slopes and some low scarps. Areas of moist, well vegetated fines have a 'horse tail' drainage pattern. The distinctive ice-wedge trough pattern, subdued landforms, and finer surficial material differentiate the lower member from the upper.

Ke_2 informal upper member: sandstone, poorly to noncemented, with minor beds of gravel, lignite, and carbonized wood. Surficial material is composed of unconsolidated fine to medium grained sand, with a discontinuous lag gravel cover, and local gravel deposits. The surface is rolling, with a succession of minor scarps and some steep slopes where drainage is incised. The unit is well drained.

Late Tertiary or Quaternary Gravel (TQ): Scattered unconsolidated gravelly deposits unconformably overlie Mesozoic sediments and commonly are capping rocks. Two units are recognized – but only on the basis of morphology as no good exposures were observed. Test pits or cores are needed to assess the composition, structure, and volume of granular materials. It is difficult to assign the smaller deposits to either unit on the basis of airphoto interpretation, therefore they are identified on Figure 95.1 by the same symbol (*). The ridges which form unit TQ_2 are shown by the symbol xxxx. Precise limits of deposits are difficult to define from superficial inspection or airphoto analysis as mass movement processes transport gravel downslope over underlying bedrock units. St-Onge (1965) also has described these deposits.

TQ_1: flat lying, locally knobby, gravelly sediments up to 2 km^2 in area, though generally much smaller. The gravel is round to subangular, granule to boulder size material, predominantly quartzose sandstone, but includes limestone, intrusive rocks, granite, and rare noncarbonized wood. The matrix of fine sand to clay normally makes up more than half of the deposit; thickness is possibly greater than 5 km where knobs occur, otherwise it is much thinner. Deposits occur on regional topographic highs, preferentially on the primary divides of the Meteorologist Peninsula, and in the centre of King Christian Island. The largest deposit is at 230 m and overlies and protects Kanguk shale, which in adjacent areas is 100 m lower in elevation. Drainage is good at the margins of deposits (note little fluvial dissection) but may be poor on level areas.

Deposits are possibly residuals from extensive fluvially deposited sediments of Quaternary or older age. Similar though far more extensive deposits in western Ellesmere and eastern Axel Heiberg islands are possibly coeval. Stott (1969) noted gravels overlying and thus postdating the Beaufort Formation in northern Ellef Ringnes.

TQ_2: linear, in places winding, ridges of gravelly material up to 15 km in length, though commonly broken by several water gaps in this distance. Orientation is between west and southwest. Where observed, surface material is granular to boulder size, round to angular quartzites, siltstone, mudstone, gabbro, rare limestone, and granite. The matrix of silty sand comprises less than half the deposit near the surface but is possibly the dominant material below the frost table. No sense of direction of sediment transport was determined. Gravel covered portions of ridges appear to be 50 to 200 m wide and 5 to 20 m high. This could be a gross overestimate of cross-sectional area if, as is likely, considerable erosion has taken place subsequent to emplacement of the deposit and gravel has slumped downslope.

Deposits occur on local topographic highs and minor divides down to, but not below, the marine limit. It is unlikely that deposition occurred preferentially on the highs, and thus a measure of subsequent erosion is available. For example, the gravel ridge northwest of Hoodoo Dome intersects the main scarp of the Kanguk shale. The scarp, which is 50 m high, has retreated at least 1 km (by lateral river erosion) since gravel deposition.

Below the Marine Limit: Sedimentary Environments of the Coastal Plain

A coastal plain, commonly 5 to 15 km wide with a typically flat to gently concave profile of less than one degree, lies between the present shoreline and the marine limit. Development required one or more lengthy marine inundations to plane the variety of underlying bedrock lithologies. Closely spaced subparallel drainage lines have been extended seawards as the plain emerged during the Holocene, and channels are now commonly incised 2 to 15 m.

In addition to washed bedrock, there are fluvial, deltaic, beach, and undifferentiated marine sediments. Each of these units is described, however they are commonly not mapped separately in Figure 95.1 both for reasons of scale and because of the difficulty of defining boundaries between the sedimentary environments.

Sediments vary in thickness from discontinuous veneers to deltaic beds more than 15 m thick. Composition is greatly influenced by underlying bedrock lithologies, and for fluvial and deltaic sediments by the materials in the drainage basin. The dominant surface material below the marine limit is fine sand or silt.

Marine-Washed Bedrock (RW): a morphologically subdued form of the bedrock units previously described. KTe_2 sediments are locally reworked into sand and gravel beach ridges at elevations close to the marine limit.

Marine Sediments - Undifferentiated (W, $\frac{W}{R}$): generally featureless sediments, excluding mappable deltaic, fluvial, or beach landforms, and including nearshore sediments, marine reworked underlying material (chiefly bedrock), possibly thin beach deposits left by the regressing shoreline, some deltaic sediments from minor drainage lines, and windblown and ice-rafted sediments.

Littoral currents appear to be weak or nonexistent. There is no evidence of longshore drift at the modern shoreline, and where bedrock contacts under the coastal plain make an acute angle with the shoreline and are not covered by deltaic sediments, the lithological boundary is commonly reflected by overlying marine sediments. This is exemplified by the sharp Ki/JKd contacts on Thor Island, the adjacent area of Ellef Ringnes, and the south shore of King Christian Island. Where contacts parallel the shore (e.g., west side of Meteorologist Peninsula), the underlying lithologies may not be apparent from the composition of overlying sediments as contacts have been blurred by the retreating shoreline.

Sediment composition varies from medium sand to silty clay in massive to finely laminated deposits. Sediments are generally more than 1 m thick, with a transitional contact to underlying bedrock, and feather out towards or at the upper marine limit.

Drainage varies from good to poor, depending on slope and materials. Some wet sedge/moss areas are present, and ponding may occur in ice-wedge troughs.

Beach Sediments (Wb): Ridge and swale development is limited by short fetches in ice-infested waters and by deflation of the available material, which is commonly coarse silt and sand size. Areas of closely spaced ridges can be identified on airphotos to ca. 10 m above present sea level and to 3 km inland; but ground inspection shows that the ridges, developed in sandy material, have only a slight morphological expression.

The modern shoreline zone is narrow — 5 to 15 m, other than on modern deltas. This is a function of the small (ca. 25 cm) mean tidal range. Surface material is dominantly sand for all coasts, although where underlying material is fine grained, the beach sand may be only a veneer 15 cm thick. A few ice-push ridges were found on all coasts; ridges were up to 50 cm high and extended to several metres inland from the high-water mark.

There are exceptions to the above observations. The modern beach at the foot of Malloch Dome is gravel, with ice-push ridges to 1 m high, and flights of gravel beaches rise to 30 m above sea level. Gravel beach ridges also are developed on hills underlain by the upper member of the Eureka Sound Formation, close to the marine limit, inland from Jackson Bay, Ellef Ringnes Island. At Cape Abernethy, King Christian Island, a gravel ridge (spit?) extends 3 km inland.

Deltaic-Marine Sediments (Fd): Deltas of larger rivers have prograded as much as 10 km, each within a 1 to 2 km-wide zone, in the course of Holocene uplift. Modern arcuate deltas thrust up to 2 km beyond the adjacent coastline, indicating little wave or current erosion or lateral deflection of the channels. The generally planar raised delta surfaces show that these quiet conditions have existed throughout much of the Holocene.

A topographic profile of the coastal plain parallel to the shoreline shows the delta surfaces rising above adjacent marine sediments. The rise is commonly less than the thickness of the deltaic sediments, indicating that some rivers occupy valleys cut in bedrock.

The modern channel is commonly incised 5 to 15 m into older deltaic sediments underlain by bedrock. Channel widths of 10 m are common on minor tributary streams, and channels on the largest rivers are up to 1 km wide. The inclination of channel banks varies from 3 to 90 degrees.

Channel bank exposures typically show stratified sand and silty sand, sometimes with minor interbedded silt, clay, or organic material. This sequence is underlain by massive deltaic-marine clay or clay-silt. The sandy topset beds vary from a thin veneer over thick basal clay, to a thick unit overlying thick or thin basal clay. Successions of clay over sand, thick interbedded sand and clay, or sediments composed entirely of shaly fragments also have been noted. Source materials within the drainage basin determine the deltaic sediment composition and the fine/coarse sediment ratio. Where overlain by deltas, the upper metre of Eureka Sound Formation poorly or noncemented sandstone on the west side of the Meteorologist Peninsula appears to have been reworked and incorporated into a basal sandy bed prior to being overlain by the clay deltaic-marine sediments.

In general, raised delta surfaces are well drained because of their elevation above adjacent sediments and their relative coarse composition. Ponding, however, may occur in ice-wedge troughs on extensive level areas.

Fluvial Sediments (Fp): Active channel zone fluvial sediments cover 10 per cent of the coastal plain. Peak stream discharge, which occurs during snowmelt, is contained within a single well defined channel, which as previously noted may be up to 1 km wide. At lower water stages flow is restricted to one or more much narrower channels, 0.5 to 1 m deep. As with deltaic and undifferentiated marine sediments, sediment composition is controlled by underlying and upstream materials.

The only area of fluvial sediment large enough to be identified as a simple unit in Figure 95.1 is east of Cape Allison, Ellef Ringnes Island. At this locality unconfined and braided channel flow occurs over ca. 20 km^2 of sand.

Permafrost

Ground ice in the upper 1 to 2 m of the permafrost was examined in cores from the 63 holes drilled. A preliminary inspection of core logs shows no clear relationship between ground-ice content and materials or vegetation. Similar results have been obtained from programs conducted elsewhere in the Arctic Islands. Excess ice content can be highly variable within a single core of the same material. Values range from 0% excess ice (pore ice, or nonfrozen water which is not uncommon) to bands of ice 50 cm thick. No massive ground ice was encountered, other than in ice wedges.

Ice wedges to 5 m wide and 10 m deep are assumed to lie under all polygonal and rectilinear pattern troughs. No relationship has been established between the dimensions of a trough and the size of the underlying wedge. Wedges also may have no surface minifestation if for example the active layer is greatly disturbed by mass movement or deflation. Trough networks are present over much of the map-area.

An active layer develops between snowmelt in late June-early July and freezeup in late August. Maximum depth of the frost table ranges from 20 to 45 cm for most units. In sand or gravel, which is either well drained or saturated by subsurface water flow in a stream channel zone, the active layer may be 1 m thick.

Vegetation and Wildlife

Vascular plants were collected, and percentage cover was estimated at a number of sites on most of the material-genetic units. Both above and below the marine limit, much of the sandstone and the Kanguk shale were nearly devoid of vegetation. Elsewhere the cover was generally sparse. Certain localized areas have a moderate cover of vascular plants and a continuous vegetation cover including mosses and lichens.

Two areas in particular are botanically diverse and also support the only *Salix arctica* observed. These are the southern and western coastal margins of Malloch Dome and the deltaic sediments at the head of Dome Bay. At the latter area at least one muskox, two wolves, and a variety of wildfowl were observed.

Geomorphic Processes

Active geomorphic processes include weathering by physical disintegration, mass wasting, fluvial, eolian, and coastal processes, as well as nivation. No attempt will be made here to weigh the relative efficacy of processes; instead, some of the more interesting processes and resultant landforms will be described.

Weathering

Bedrock, whether or not disturbed by mass erosion, is commonly physically disintegrated to at least a 1 to 2 m depth. This depth greatly exceeds the present maximum thaw depth; however, it does not necessarily imply a formerly thicker active layer. Breakdown could have been accomplished during accretion of ground ice, as well as by frost shattering.

The weathered residual material and rock surface is normally smooth; however, irregularities can develop. Hoodoos are common where certain better cemented units of the Isachsen Formation are flat lying. Development is assumed to result from nivation and fluvial erosion along joints, possibly aided by eolian abrasion. Near the base of the Christopher Formation in east-central King Christian Island, a row of strike-aligned mounds to 1.5 m high and 5 m in diameter (Fig. 95.2) can be traced across the island down to about 35 m elevation on both northern and southern coastal plains.

Figure 95.2

Concretions uncovered by weathering and erosion of surrounding Christopher Formation shale, King Christian Island.

Figures 95.3 and 95.4

Earthflows in marine-deltaic sediments, subsequent to rainfall of July 30-August 1, 1976, north of Jackson Bay, Ellef Ringnes Island. The basal shear zone is at the frost table.

The mounds are diagenetic concretions (H. R. Balkwill, pers. comm.) surrounded by shale. Their presence below the Holocene marine limit indicates either that they have been uncovered by weathering and mass erosion of surrounding shale in the course of the Holocene or, less likely, that the marine inundation and later regression of the shoreline was sufficiently rapid that the concretions were not planed off.

Mass Movement

Solifluction is undoubtedly an active process; however, lobes were rarely observed, possibly because of the generally low inclination of slopes and the sparse vegetation cover. Both sorted and vegetated stripes are common.

The earthflow is the most striking type of mass movement. It occurs preferentially on fine grained materials, on slopes from less than 1 to 10 degrees with or without vegetation, and most commonly occurs below the marine limit although numerous earthflows were noted above the marine limit on the Christopher Formation. The basal shear zone is at the frost table and usually is in clay or silty clay, but in places failure may occur in fine sand. The thickness of the sliding material is thus rarely greater than 50 cm although the area of material that moves may be as great as 500 m^2. On river banks, undercutting of the toe of the slope is also a factor, although earthflow failures rarely occur on steeply undercut banks. Saturation of the active layer takes place during snowmelt; however, the layer is probably too thin for earthflows to be initiated at this time. They occur later in the summer, provided unusually heavy or extended rainfall saturates the active layer which is then at maximum thickness. It is likely that in summers with only light rain no failures occur.

In 1976 many earthflows occurred north of Jackson Bay during a period of intermittent rain (ca. 10 mm recorded at Isachsen, 100 km to the northwest) from July 30 to August 1, approximately two weeks after the end of snowmelt. Some slides reoccupied scars from earlier years. Figures 95.3 and 95.4 show a few of the many hundreds of earthflows activated in these two days on river banks cut in marine-deltaic sediments. No retrogressive thaw flowslides were observed, possibly indicating a general absence of massive ground ice.

Terrain Sensitivity

Although parts of the study area are highly susceptible to terrain disturbance, the net effect on the sparse vegetation and low density of wildlife probably would be slight. The few botanically diverse areas, however, should not be disturbed. The rivers seem an unlikely habitat for fish because of the short flow period, high sediment loads, and the acidic water where Kanguk shale is drained. Coastal waters were not considered in this study.

Of greater concern is the effect of a number of the geomorphologic processes on roads or pipelines. Some potential problems are: scouring of channel beds in the numerous water courses; initiation of gullies by concentrating runoff; failure of river banks; failure, particularly by earthflow, of any slope on the Christopher Formation or on fine grained coastal plain sediments; abrasion by windblown sand; corrosion, due to high acidity, on or downstream from the Kanguk Formation.

Quaternary History

No direct evidence of Pleistocene glacial erosion or deposition was found in the study area; topography appears to be chiefly a product of fluvial processes, mass wasting, and marine planation. A thick cover of weathered residual bedrock is present (though the rate of weathering is unknown); no morainal deposits have been found on the surface, incorporated in residual material, or underlying marine, deltaic, or fluvial sediments.

The flat lying gravel deposits (TQ_1), if remnants of widespread fluvial sediments, explain the presence of any exotic lithologies in lag deposits. These gravel caps also appear undisturbed by glaciation. A glacial origin for the gravel ridges (TQ_2) has yet to be proven, despite their esker-like form. If they are fluvioglacial sediments, the degree of erosion subsequent to emplacement makes a late Quaternary age unlikely.

Nevertheless, there is strong indirect evidence of glaciation. Some adjacent interisland channels have a trough-like form, and Balkwill *et al.* (1974) describe striations and striated erratics on adjacent Amund Ringnes Island, although the age of these features is unknown. The amount of uplift of King Christian and southern Ellef Ringnes Islands during the Holocene, however, seems only explainable by isostatic rebound from an ice cover of late Quaternary age.

England (1976) has suggested that much of this uplift is recovery from isostatic depression peripheral to the late Quaternary margins of the Greenland Ice Cap and enlarged Ellesmere and Axel Heiberg Island ice caps. On the basis of Walcott (1970), England considered the limit of the peripheral depression to be 180 km. King Christian Island, however, is 600 km beyond the suggested margin of Greenland ice, 200 km from the nearest coast of Axel Heiberg Island, and at least 400 km beyond the northern limit of Laurentide ice. Thus the 30 to 50 m or more of emergence in the Holocene is best explained by rebound from an ice cover over the islands and intervening channels, as in the concept of the Innuitian Ice Sheet proposed by Blake (1970), rather than by the influence of distant ice sheets. The lack of glacial landforms may be a consequence of the ice sheet being cold based (i.e. frozen to the underlying material), perhaps due to a low mean annual temperature. Another possibility is that locally thicker ice centres developed over higher land (the present islands) and there was little ice flow at these locations.

Radiocarbon age determinations are available for four surface samples collected by D.A. St-Onge on Ellef Ringnes Island. *Astarte borealis* valves from the southwest corner of the Noice Peninsula, at 22 m, are 7350 ± 200 years old (L-643B; St-Onge, 1965); *Astarte borealis* and *Hiatella arctica* from the south end of the Meteorologist Peninsula, at 33 m, are 8500 ± 200 years old (L-643A; St-Onge, 1965); a further determination on *Astarte borealis* shells, collected near the preceding sample L-643A, provided a date of 8370 ± 200 years (GSC-1846; Lowdon and Blake, in press); and driftwood in the same vicinity at 25 ± 5 m is 8320 ± 140 years old (GSC-999; Blake, 1970).

Two shell samples collected in 1976 have been dated. A *Mya truncata* valve on the surface at 42 m in south-central King Christian Island (77°45.25'N, 101°37.25'W) is 8900 ± 140 years old (GSC-2386). The sample dated was taken from one of a number of valve clusters and fragments, each representing one or two paired valves, surrounded by silty clay. The sediment, of marine or deltaic origin, extends to 48 m in elevation at this location. This age determination provides a minimum age and elevation for the highest Holocene sea level.

Mya truncata valves taken 8 m below the top of an exposure in deltaic sediments with the surface at 33 m, 12 km north of Jackson Bay, Ellef Ringnes Island (78°12.25'N, 100°46.75'W) are 7640 ± 120 years old (GSC-2383). In the exposure, which is typical of deltas on the coastal plain, interbedded silt, sand, and minor clay overlie massive dark grey clay with shells, which in turn unconformably overlies Eureka Sound Formation sediments. The shells were taken from the lower metre of clay. The age determination shows that the basal clay in at least this deltaic sequence is Holocene in age. It also provides a maximum age for sea level at 33 m at this location. The shells are surprisingly young in comparison with the three samples from a similar elevation at the south end of Meteorologist Peninsula – particularly the driftwood, which is usually contemporaneous with the shoreline on which it is deposited.

On southeast King Christian Island, shells and possible beach ridges are evidence of marine overlap to at least an 80 m elevation. Above ca. 45 m, however, shells tend to be notably thicker and more incrusted and pitted than those at lower elevations. Although no age determinations are available yet, the shells at higher elevations do appear similar to those collected elsewhere in the Arctic Islands which have provided infinite ages. The local abundance on the surface and the presence of pairs make glacial transport unlikely;

but with an ice cover and no basal ice movement, preservation of shells *in situ* is possible.

In the event of such an ice cover, then the late Quaternary high sea level would initially overlap the ice, and the only marine sediments to be expected would be in subsequent offlap deposits. On the other hand if no ice cover was present, some evidence of onlap would be expected in the form of sediments and anomalously old dates underlying Holocene offlap deposits. No such evidence has been found, and even if regressive shoreline processes eroded much of the onlap sediments, some should be preserved where deltaic offlap deposits were built beyond the (shallow) wave base – and it is in such deltaic deposits that many of the best exposures are cut.

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C.F.M. Lewis, S.M. Blasco, B.D. Bornhold, J.A.M. Hunter[1],
A.S. Judge[2], J.Wm. Kerr[3], P. McLaren, and B.R. Pelletier
Terrain Sciences Division

Introduction

A joint project by four Divisions of the Geological Survey of Canada and the Earth Physics Branch (Project EPB 5.2.1) was planned as a multidisciplinary marine program in the central Arctic, in the area of probable future pipeline and transportation development. The objective was to advance basic knowledge of seafloor and coastal materials and of underlying bedrock structure. The expedition planned to operate from *C.S.S. Hudson* for 30 days in the ice-infested waters of Barrow Strait and adjacnet channels. The proposed investigations related to surficial sediment distribution processes, coastal materials and processes, bedrock structures linking and dividing the Arctic Islands, and subsea thermal regime, with special reference to ice-bonded permafrost. The cruise also was planned to utilize high resolution geophysical techniques in order to determine the physical properties of near-surface seafloor sediments. Two launches, *Gull* and *Fulmar* were outfitted for diving and for sonar/seismic/sampling operations and to conduct nearshore programs in conjunction with *Hudson's* offshore operations. An evaluation of the Differential Omega positioning system was also scheduled by Bedford Institute's Navigation Group to take advantage of *Hudson's* operating location close to Resolute, where an existing ground station monitored variations in VLF radio wave propagation characteristics.

Unfortunately, the locale of the entire program had to be changed and its period cut in half when *Hudson* lost one of its two propellers near Prince Leopold Island en route to Resolute. Although normally capable of limited movement in ice, the research vessel was thereafter restricted to ice-free waters. Because of the abundance of sea ice in 1976 the program was recast in a contingency area — Lancaster Sound — where open water conditions generally prevailed (Fig. 96.1).

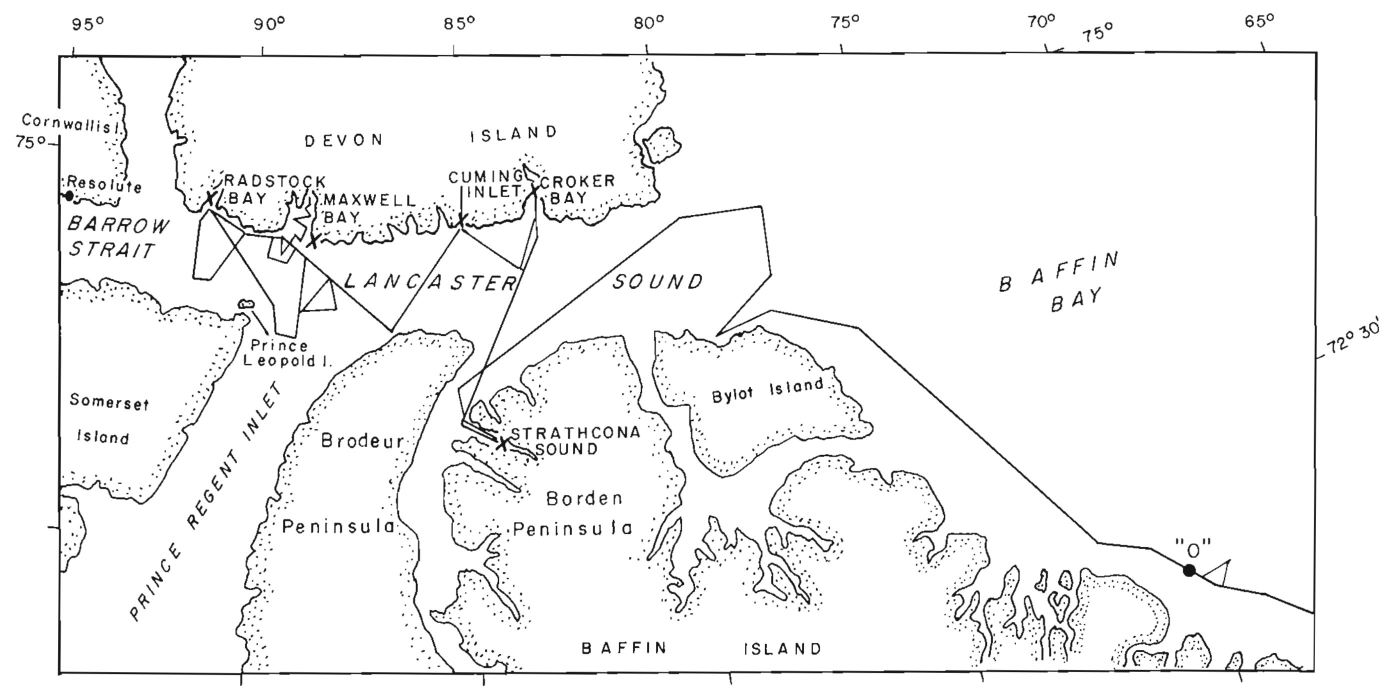

Figure 96.1. Sketch chart of *C.S.S. Hudson* tracks in Lancaster Sound area in 1976. The X's indicate nearshore study sites with the launches *Gull* and/or *Fulmar*.

[1] Resource Geophysics and Geochemistry Division

[2] Earth Physics Branch, Contribution No. 642.

[3] Institute of Sedimentary and Petroleum Geology, Calgary

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Geol. Surv. Can., Paper 77-1A (1977).

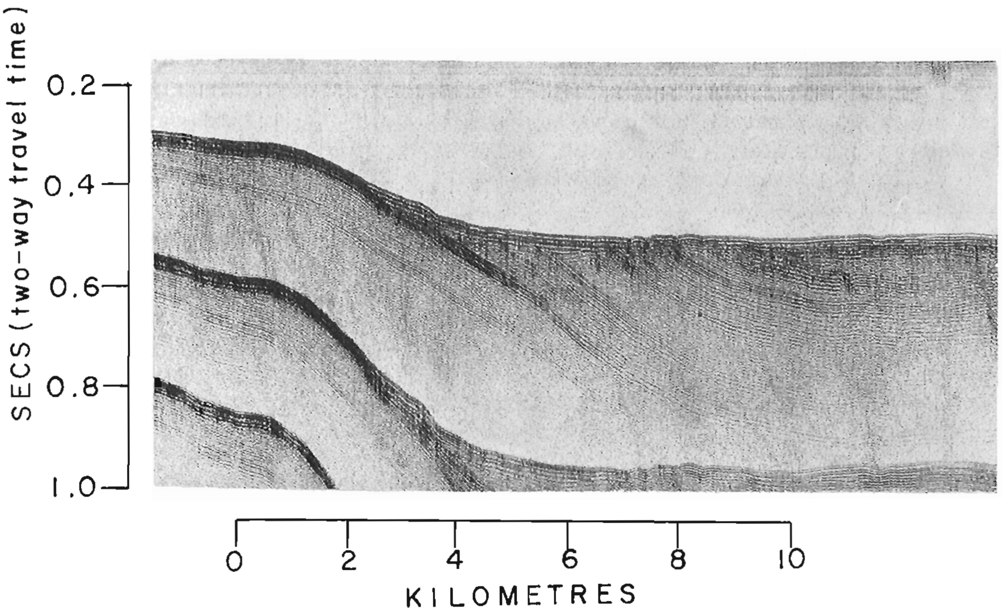

Figure 96.2.

Air gun seismic reflection record from western Lancaster Sound south of Radstock Bay. The profile crosses the nothern rim of Barrow Basin and shows Paleozoic carbonate rocks in the north (at left) overlain by Mesozoic-Tertiary clastics in the basin (at right).

All the foregoing planned activities then were undertaken within the abbreviated program in Lancaster Sound, except for seafloor rock drilling and suspended matter sampling. The following programs were undertaken:

(1) An analysis of surficial sediments was made by combining the results of high resolution seismic work with sediment coring. Surficial sediments were cored at selected sites and analyzed in the conventional manner. Physical sedimentary properties were determined remotely using the Huntec '70 Deep Tow seismic boomer (Hutchins *et al.*, 1976), a 20 m seabed seismic refraction array, neutron (porosity) probe, thermal gradiometer probe, gamma-gamma (density) probe, a spectrometer probe (for K, U, Th concentrations), a gamma-ray probe (for natural gamma radiation); and a seawater velocimeter probe.

(2) Representative nearshore sediment facies were examined by scuba divers with particular reference to bottom scouring by drifting icebergs and sea ice. This work will contribute to geological characterization of Arctic coasts and to the development of site selection criteria for future coastal structures.

(3) Determinations were made of seabottom temperatures, sediment and water thermal gradients, and sediment thermophysical properties, primarily for an assessment of the potential for ice-bonded subsea permafrost.

(4) Surficial sediment distribution and offshore bedrock structures were delineated by seismic, magnetic, and gravity profiling. The distribution of Cretaceous-Tertiary sedimentary basins and the fault-bounded margins of Lancaster Sound were of particular interest.

In addition to obtaining specific information about the Lancaster Sound region, the program also contributed to developing techniques and operational expertise for future application in other Arctic areas. The actual field operation in the Lancaster Sound area ran from August 19 to September 3. From September 4 to 13 *Hudson* was directed southwards for repairs with one stop on the Baffin Island shelf (at point '0', Fig. 96.1) to sample an oil slick Loncarevic, and Falconer, 1977).

Lancaster Sound

Seismic reflection profiling was undertaken in Lancaster Sound to: (1) establish the structural framework and its control on Mesozoic, Tertiary, and Quaternary sedimentation and (2) attempt to understand better the nature and history of Quaternary marine environments. Both conventional air gun seismic reflection techniques and the Huntec '70 Deep Tow System were used. The latter provided detailed records of the unconsolidated sediments lying on bedrock.

The main channels in the central Arctic are grabens, and their internal topography generally is controlled by bedrock structure (Kerr, in press). Bedrock structure that developed in a Cretaceous-Tertiary downfaulting episode provided a topographic framework which has been modified by erosion and deposition to give details of present bathymetry. A major Mesozoic-Tertiary basin of this origin occupies part of Lancaster Sound and is bounded on the north by the steep faulted southern margin of Devon Island. It has been identified by Daae and Rutgers (1975) and further defined by Bornhold and Lewis (1976). The basin occupies the northern half of western Lancaster Sound, has an area

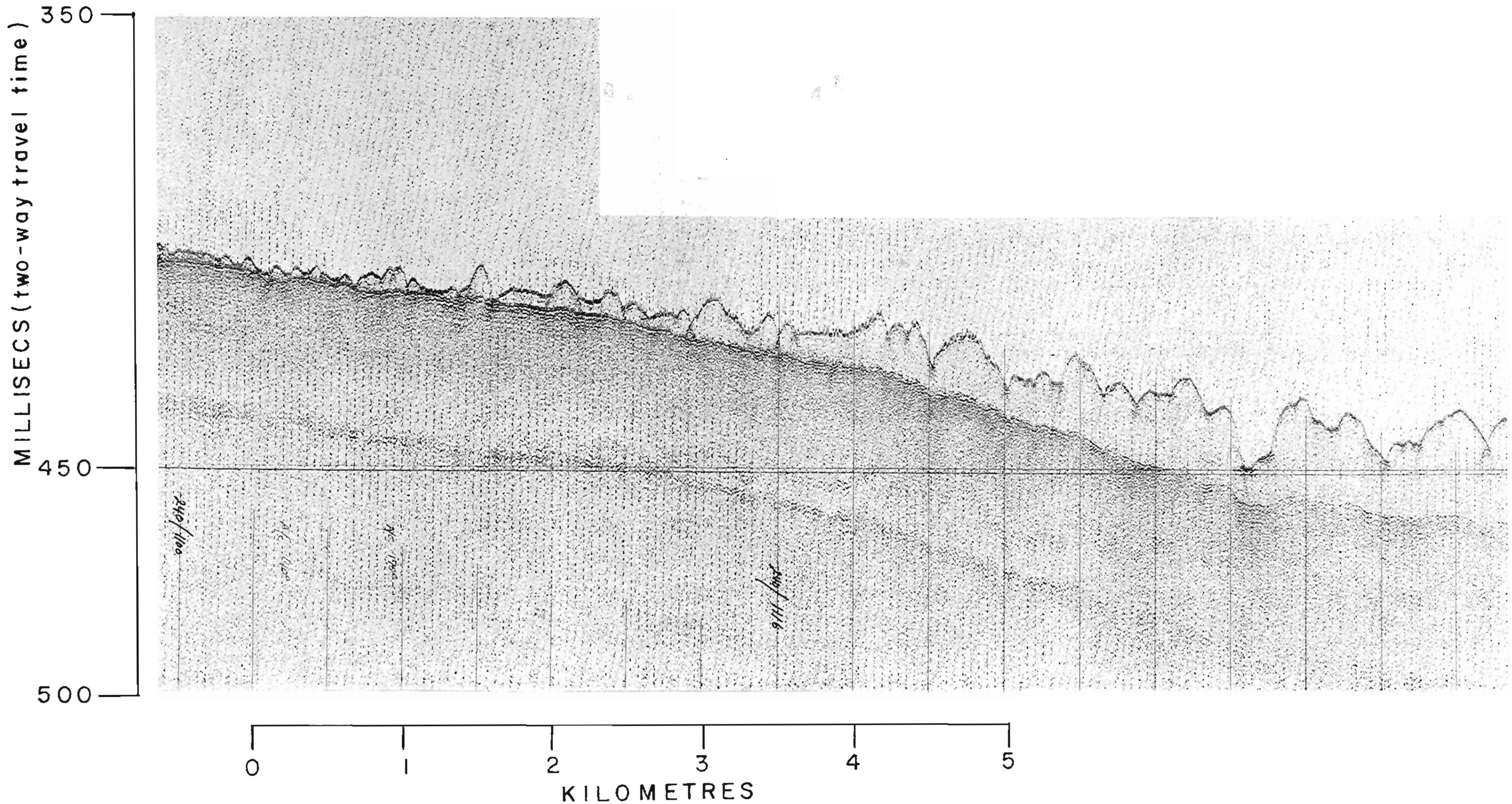

Figure 96.3. Huntec '70 Deep Tow seismic reflection record in western Lancaster Sound (approx. 74°04'N, 89°25'W) showing a thick accumulation of unconsolidated sediments overlying bedrock in deeper water but thinning to less than 5 m in shallower areas.

Figure 96.4. Seafloor photograph from western Lancaster Sound showing pebble-cobble pavement and evidence of strong current flow.

of 2400 km^2, and contains poorly consolidated clastic sediments at least 1300 m thick. Seismic reflection profiles (Fig. 96.2) also were obtained across the northern rim of Barrow Basin (Bornhold *et al.*, 1976; Bornhold and Lewis, 1976), which is a Mesozoic-Tertiary basin and also a prominent modern day topographic depression between Somerset and Devon islands. This basin occupies an area of approximately 700 km^2 and contains about 1000 m of clastic sediments in its deepest parts. The location of Barrow Basin may be controlled by bedrock structure. Its eastern margin may be related to the northward continuation of a fault suggested along the east coast of Somerset Island (Kerr and de Vries, 1977).

Earlier studies of the Quaternary sedimentary history of western Lancaster Sound (Bornhold and Lewis, 1976), based primarily on short gravity cores and seafloor photographs, concluded that net accumulation is occurring at the present time in only a few, small isolated basins, principally off northwestern Baffin Island and northern Somerset Island. Seismic reflection records (Fig. 96.3) obtained by the Huntec '70 System, combined with seabottom photography (Fig. 96.4), grab sampling, and coring, confirm the interpretation that the seafloor in western Lancaster Sound south of Devon Island is primarily undergoing erosion. The seismic profile (Fig. 96.3) shows hummocky topography developed on a thick sequence of unconsolidated sediments near the centre of the Lancaster Sound.

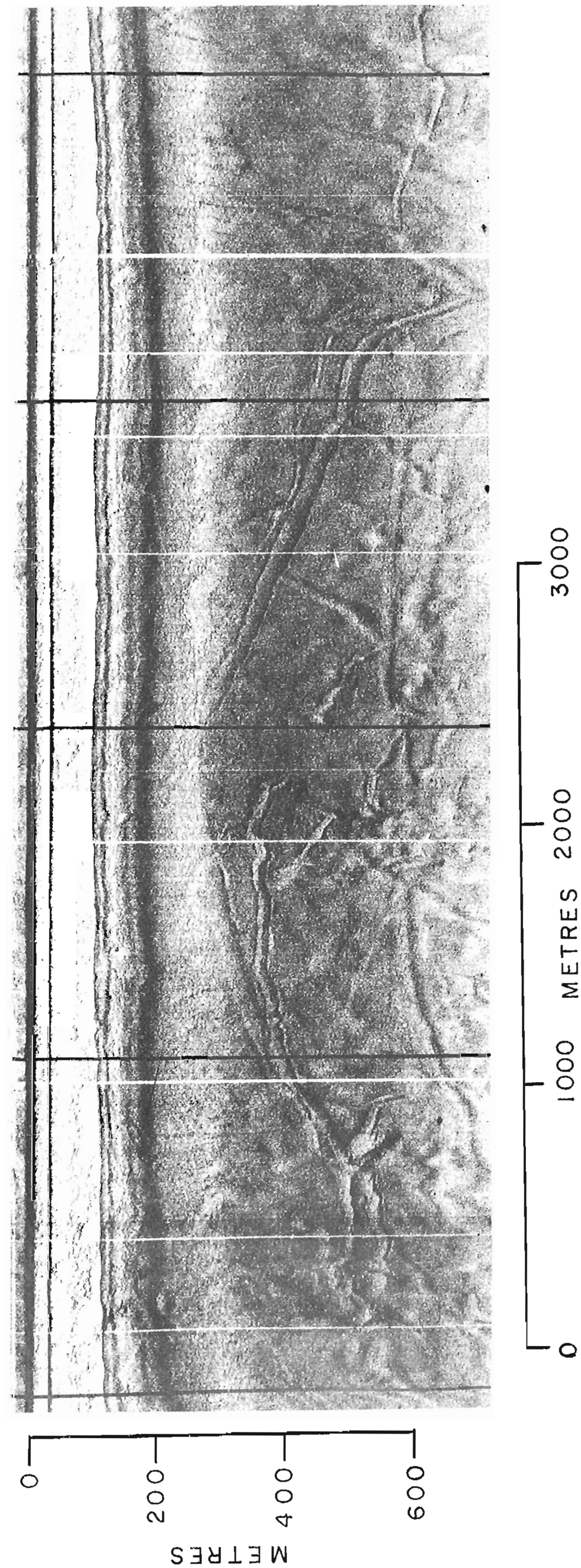

Figure 96.5 (opposite)

Side-scan sonograph off Fellfoot Point, southeastern Maxwell Bay, Devon Island, in approximately 130 m water depth showing numerous intersecting ice-scour tracks.

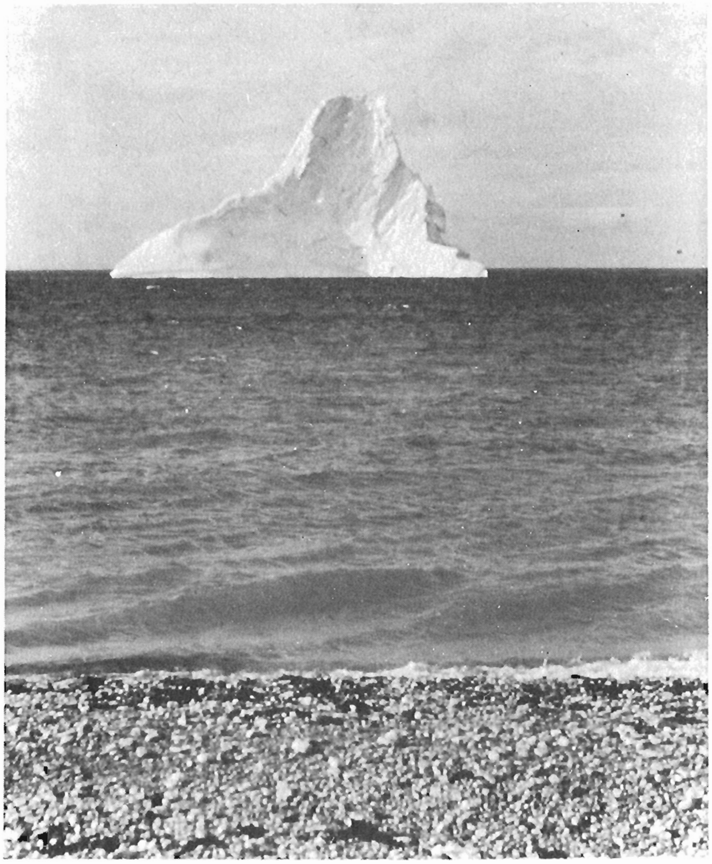

Figure 96.6. Grounded iceberg observed off Fellfoot Point, Maxwell Bay, in August 1973.

Farther north on the same profile, these sediments thin rapidly and finally are left as small, isolated, pebble- and cobble-mantled mounds on the underlying bedrock. The ubiquitous pebble and cobble pavements are remnants of older unconsolidated sediments from which the fine material has been winnowed by seafloor currents. The original sediments are presumed to have been glacial tills, overlain by pebbly marine muds which were deposited in an ice shelf environment during the early stages of deglaciation. Circulation in western Lancaster Sound and Barrow Strait at the present is too vigorous for fine sediments to accumulate, except in isolated depressions and in easternmost Lancaster Sound where water is deep.

Side-scan sonar studies were carried out in several areas off the south coast of Devon Island in an effort to determine the significance of ice scour. Sonographs off Fellfoot Point (southeastern Maxwell Bay) revealed scours (Fig. 96.5) in water depths of 50 to 130 m. A grounded iceberg was observed in this area in August 1973 (Fig. 96.6). Side-scan sonographs in water depths more than 150 m off southwestern Devon Island also showed long, curved, and sinuous scour tracks (Fig. 96.7) of unknown age. It is possible that these deep-water scour tracks are old, possibly relict, since they may be inscribed onto a nondepositional seabed and would not be obliterated by normal sedimentation.

The question of scourage, particularly in these relatively deep waters, is extremely relevant to engineering design of seabed pipelines; if relict, the scours represent no threat, but if they are modern, future ice scouring poses a distinct hazard to pipeline integrity.

Devon Island Fiords

Marine geological and geophysical studies were carried out from *C.S.S. Hudson* and the launches *Fulmar* and *Gull* in several fiords — Radstock Bay, Maxwell Bay, Cuming Inlet, and Croker Bay. Seismic reflection profiling by both air gun and the Huntec '70 Deep Tow System extended work done from *C.S.S. Baffin* in 1973 (Lewis *et al.*, 1974) and from *C.S.S. Hudson* in 1974 (Blake and Lewis, 1975). The objectives were: (1) to determine the structure of the fiords and relate this to larger structural elements in the Arctic Islands and (2) to determine the thickness, age, and character of unconsolidated sediments accumulating in the bays. The bays and inlets of southern Devon Island are dominated by north-south trending linear ridges and depressions, presumably horsts and grabens. At most, only a few metres of sediment have accumulated on the

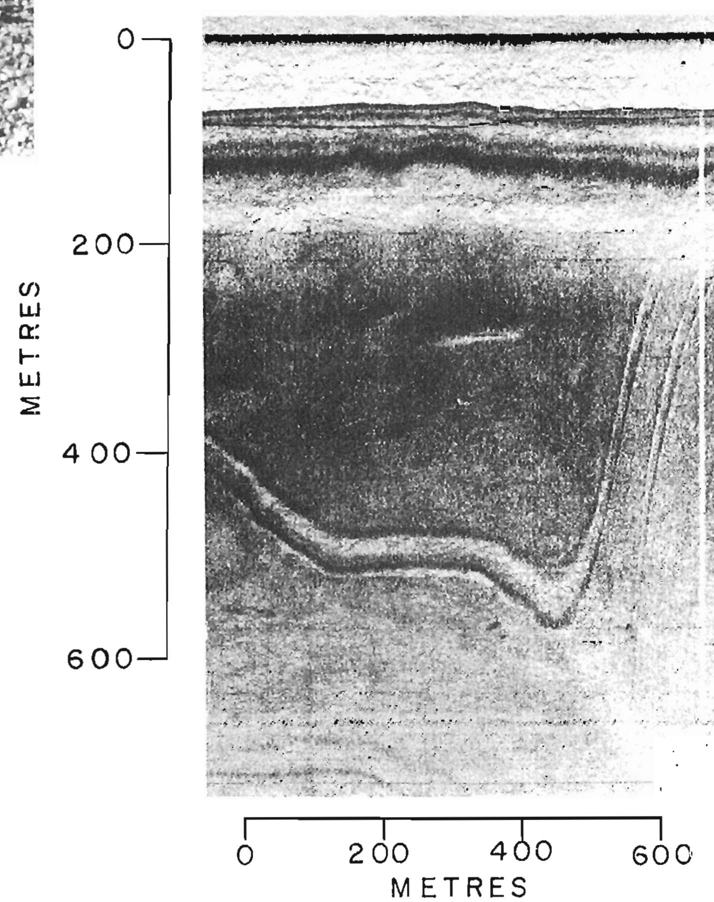

Figure 96.7. Side-scan sonograph in 145 m water depth in Lancaster Sound (74°31'N, 91°16'W) south of Radstock Bay. The record shows several ice-scour tracks.

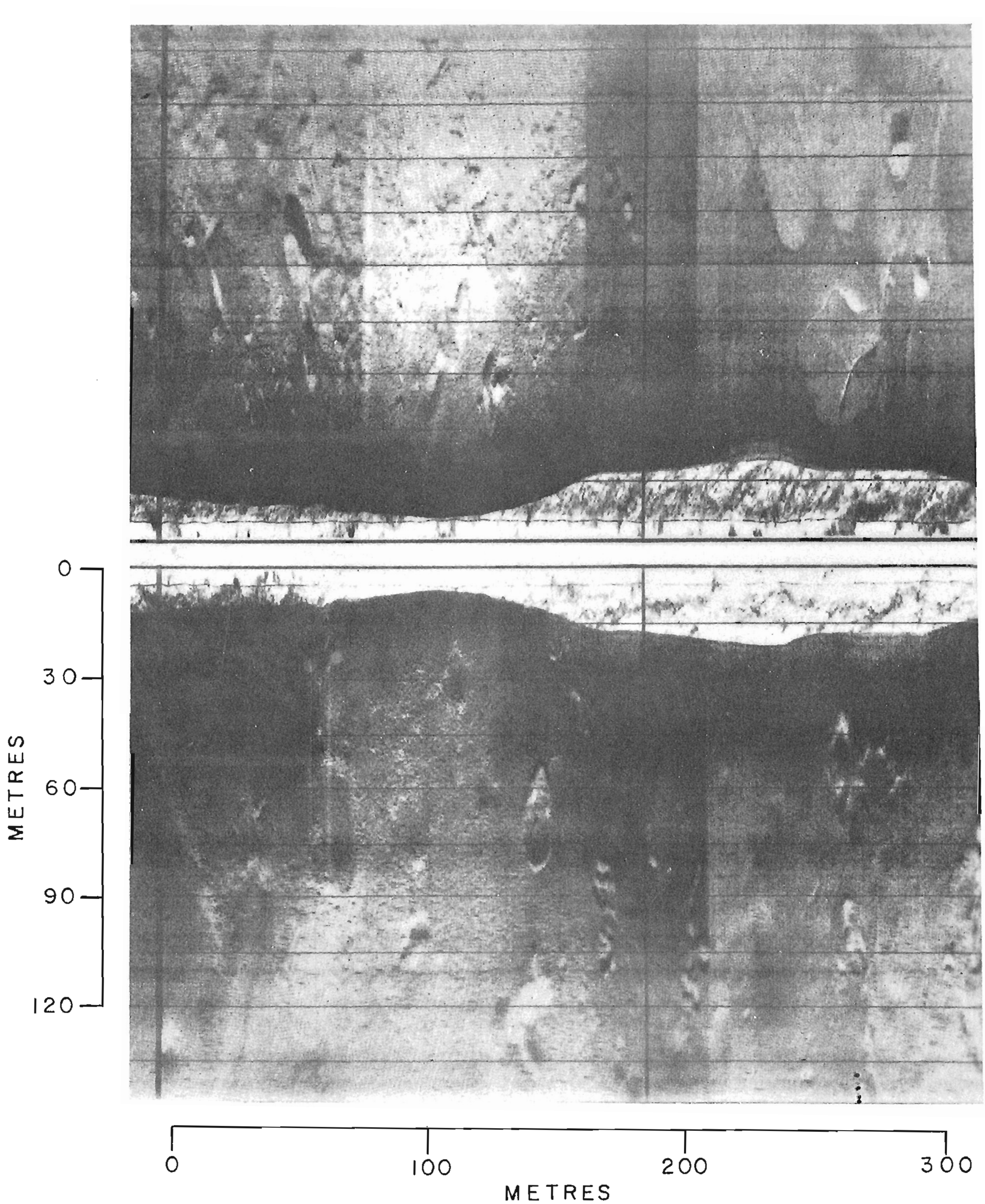

Figure 96.8. Side-scan sonograph in Radstock Bay (74°44'N, 91°10'W) in water depths between 10 and 40 m. The record shows numerous isolated crater scours presumably the products of periodic impact of ice in the shallow nearshore area.

Figue 96. 9.

Cobble pavement in nearshore region of Radstock Bay. Water depth approximately 15 m. Length of scale is 30 cm.

Figure 96. 10.

View of the interior of an ice scour at Dealy Point in Radstock Bay. The boulder and cobble debris has slumped down from the unstable oversteepened sides which are sloping at approximately 50 degrees. Water depth is approximatley 20 m.

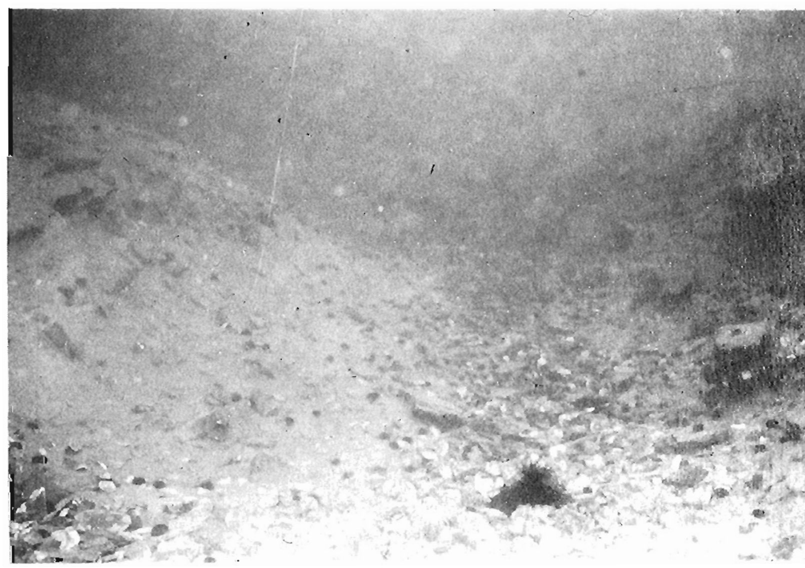

Figure 96. 11.

Diver using a Sonardyne Mark IV Underwater Rangemeter positioning system. A transponder held in its frame is in the foreground, and the diver is measuring the distance of a base line to a second transponder 200 m away.

ridges, but up to 30 m of fine muds and interstratified coarse sandy turbidites are common in the deep basins of these fiords. A rectilinear pattern of normal faults exists on southern Devon Island (Thorsteinsson, pers. comm., 1976). Thus structure, sculptured by erosion, may control the pattern of inlets on southern Devon Island.

A small embayment on the southwest coast of Radstock Bay and nearby parts of Barrow Strait were studied in greater detail using *C.S.S. Hudson* and both launches to investigate a potential harbour site. Echo sounding, seismic refraction, side-scan sonar, grab sampling, and coring were done from the *Fulmar*. Launch navigation was controlled by a Trisponder

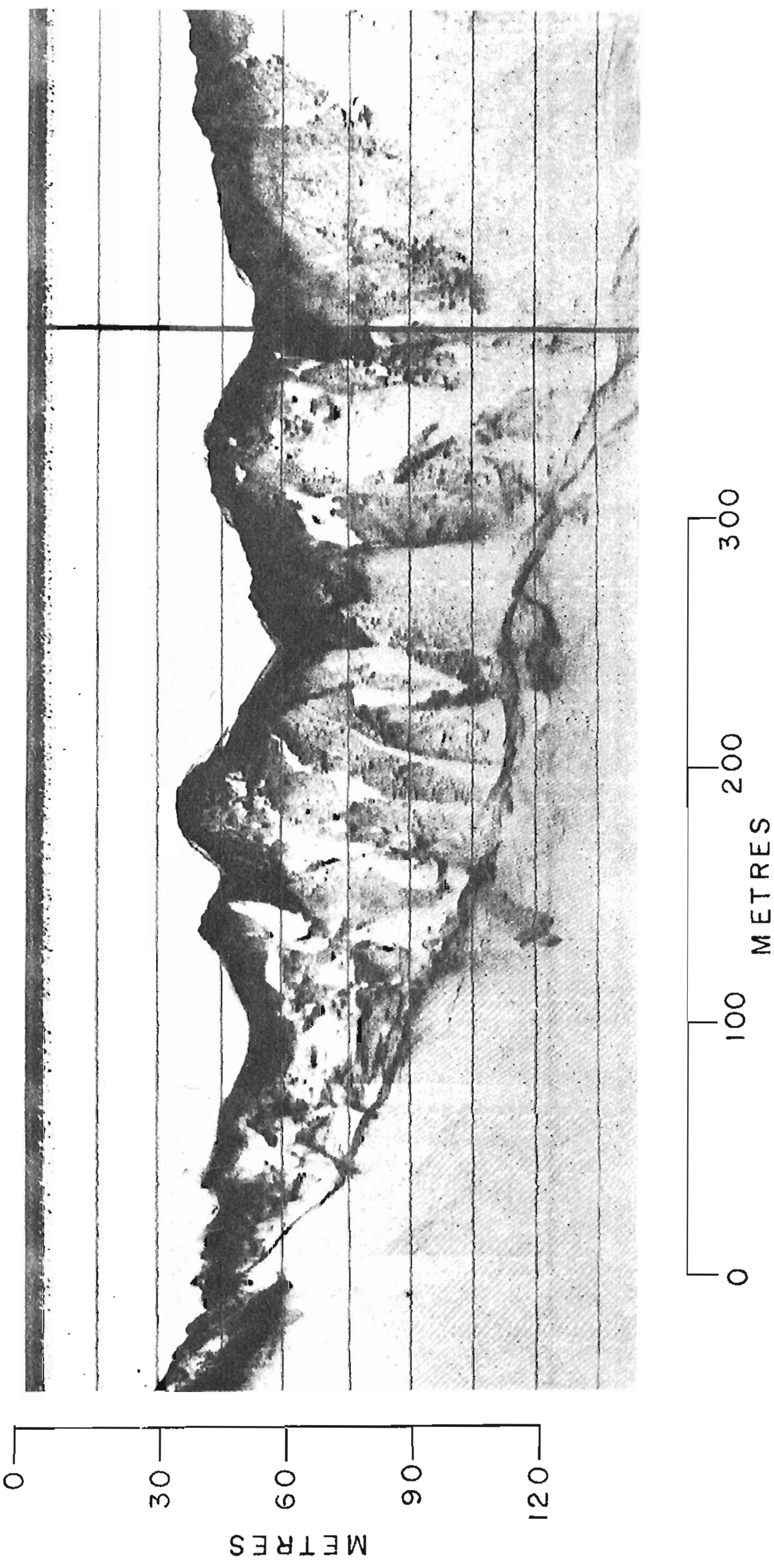

Figure 96.12.

Side-scan sonograph of the glacier in northwestern Croker Bay, Devon Island (74°53'N, 83°30'W) showing ledges and fissures in the glacier front. The seafloor was beyond the range of the side-scan sonar. The sonograph shows the morphology of only the underside of the ice.

range-range positioning system. Side-scan sonar records revealed a 'cratered' seafloor in nearshore waters believed to be caused by the impact of sea ice (Fig. 96.8).

Diving investigations, carried out from the *Gull*, at the potential port site at Radstock Bay included photography, coring, and biological sampling. In general, the nearshore bottom consisted of 80 to 100 per cent gravel and cobble pavement (Fig. 96.9) resting on steep slopes (30 to 45 degrees). Boulders were common, particularly adjacent to cliffed coastlines. The sedimentary matrix beneath the pavement contained a poorly sorted sandy mud, possibly till. It was difficult to obtain cores, and this would be done only after the surficial cobble pavement was removed. Ice scour was developed spectacularly on the seaward side of shoals and on exposed headlands (Fig. 96.10). The scours were observed to be approximately 3 m deep, 6 m wide, and 30 m long.

A Sonardyne, Mark IV Underwater Rangemeter and transponder system (Kelland, 1975) for positioning divers was used successfully. This instrument is a diver-operated acoustic range-range system that presents two way travel time between diver and transponder as a digital readout to a maximum of five different transponders (Fig. 96.11). The system range resolution is 0.1 ms or 150 mm at a propagation velocity of 1500 m/s; the useful range is approximately 300 m. This type of instrument is considered essential for future detailed underwater mapping. Three cores were located accurately with respect to two transponders.

Launch studies in Cuming Inlet and Croker Bay focused on the marine margins of three glaciers. Side-scan sonar, echo sounding, grab sampling, coring, suspended particulate matter collection, and diving observations were carried out on and near the ice fronts and on nearby shoal areas. Side-scan sonar profiles (Fig. 96-12) parallel to the ice fronts revealed the submarine morphology of the glaciers. Fissures extending more than 75 m into the ice and ledges of ice several tens of metres in length could be mapped readily using this technique.

Figure 96.13.

Proximal side of boulder moraine near margin of calving glacier front; depth is approximately 25 m.

Figure 96.14.

Prolific fauna and flora on a boulder-covered shoal. Depth approximately 7 m. Length of scale is 15 cm.

Poor visibility (less than 1 m) and water depths beyond the capabilities of SCUBA, precluded making abundant observations of the glacier margins. In all cases a very steep ridge of unstable boulder moraine was parallel to the ice front (Fig. 96.13). The slope on both proximal and distal sides was approximately 45 degrees. The seaward slope was mantled with silt derived from the release of rock flour from the melting glacier ice. The area of active sedimentation, however, was very limited; this conclusion followed from diving observations on a shoal less than 50 m from the glacier front, which contained an area of prolific fauna and flora in a nondepositional environment (Fig. 96.14).

Strathcona Sound

Piston coring, grab sampling, sea floor photography, physical oceanography, and geophysical studies in Strathcona Sound continued work begun in 1975 (Bornhold, 1976). Seismic reflection records (Fig. 96.15) permitted the delineation of several sedimentary basins. These basins are separated by northwest-southeast trending, linear bedrock ridges, for the most part devoid of sediments. Unconsolidated sediments up to 90 m thick and consisting of alternating red-brown muds and sandy turbidites were found to be accumulating in the intervening basins.

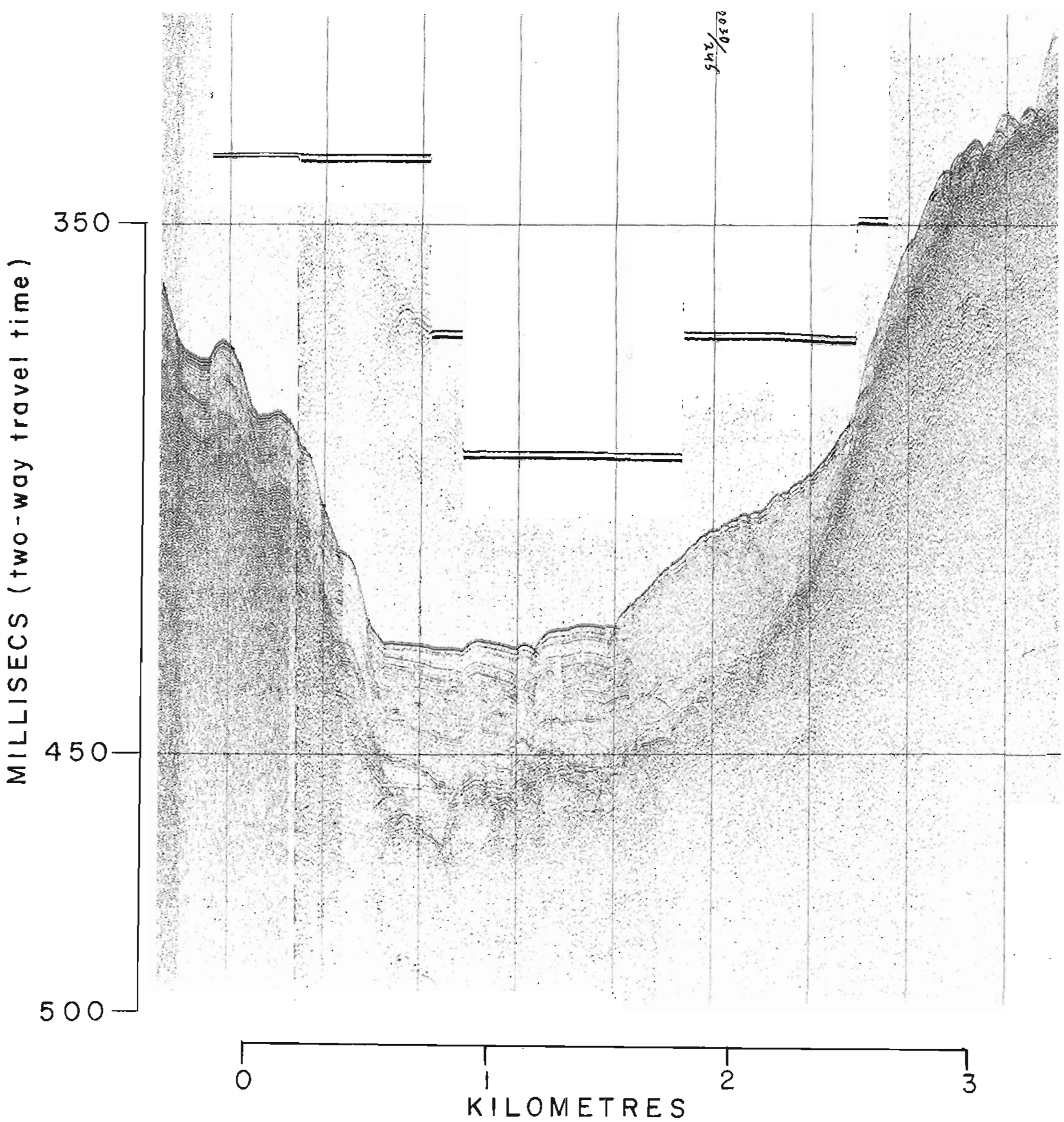

Figure 96.15. Huntec '70 Deep Tow seismic record in Strathcona Sound (73°06'N, 84°40'W) showing an accumulation of more than 35 m of muds and highly reflective sandy turbidites in the deepest part of the basin, structureless slumped sediments near the foot of the fiord wall, and nearly sediment-free valley walls.

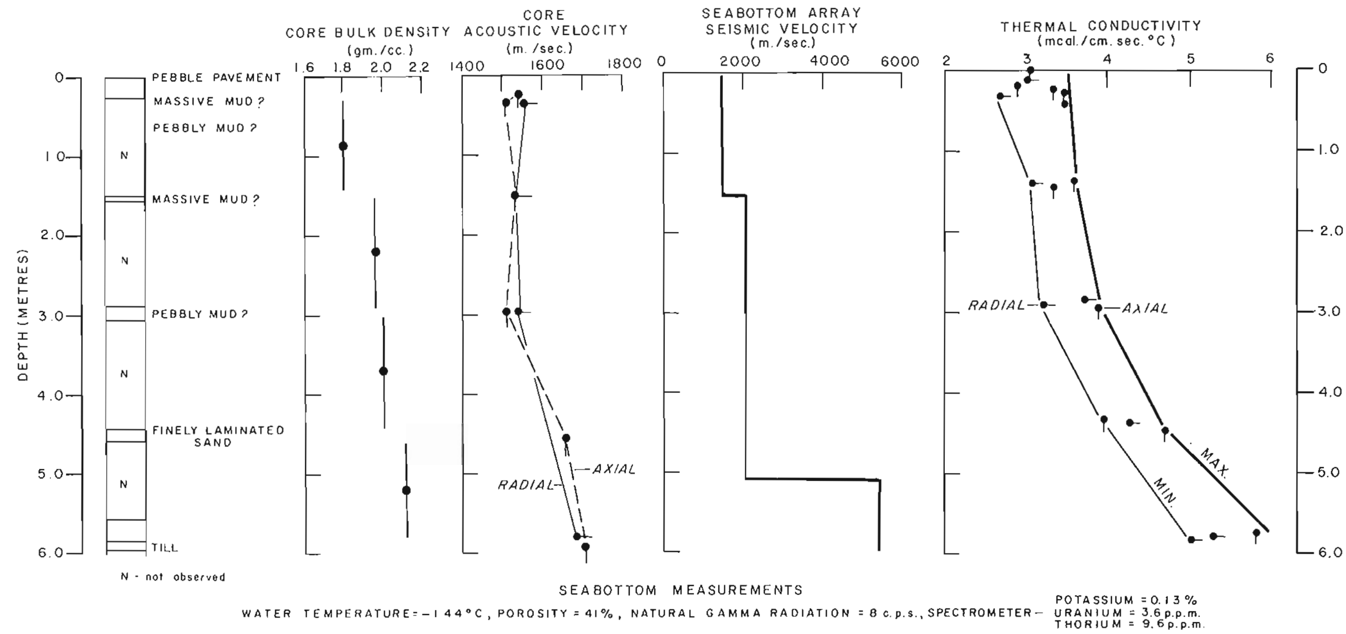

Figure 96.16. Seabottom properties and profiles of some physical parameters of a piston core from Maxwell Bay, Devon Island (74°35.6'N, 88°51.0'W). The core lithology is interpreted from X-radiographs of selected core segments. Values are preliminary and subject to revision.

Side-scan sonographs in the shallow water along the sides of the fiord reveal a few ice-scour tracks, ripple marks and bedrock outcrops. Channels, sediment dunes, and ripples were seen on the front of a small delta on the north side of the sound.

Bottom diving observations and coring attempts on both sides of the sound showed an impenetrable boulder and cobble pavement. The steep sides of the fiord continue underwater at slopes of 30 to 40 degrees in the nearshore. The nature of the fauna and flora is indicative of a nondepositional environment, but no ice scour was seen in the limited area of observation.

Physical Properties of Arctic Sediments

Co-ordinated study of the geotechnical, geological, and geophysical properties of Arctic marine sediments is one of the unique benefits of this multidisciplinary cruise. Nineteen piston cores representing a variety of surficial, glacial, and marine sedimentary environments were collected throughout the Lancaster Sound area including all the fiords visited. Sediment structure was given preliminary appraisal through on board radiography of unopened core liner tubes (Wahlgren and Lewis, 1977). Geological and geophysical samples and data were collected simultaneously at 6 of the 19 core locations to investigate the relationship between remotely sensed acoustic/seismic parameters and sediment geotechnical attributes. The ultimate goal is to improve the potential for high-speed prediction and mapping of sediment properties from analysis of reflected acoustic signals. A second goal, achieved through the measurement of thermophysical properties at the same locations, is to increase understanding of the subsea thermal regime and its relation to the sediments. The equipment deployed at these six test sites was that previously listed in the introduction.

An example of information generated aboard ship is given by the 6 m core from Maxwell Bay which is composed at the top of a pebble lag-veneer above silty mud and at the base of a compact pebbly grey till (Fig. 96.16). An obvious general increase with depth exists in the curves of each of the measured parameters: bulk density, sound velocity, and thermal conductivity (Fig. 96.16). Detailed analysis has yet to be made, but, for example, the consistent difference between axial and radial thermal conductivities suggests a horizontal rather than vertical preference in the orientation of sedimentary particles similar to that in a bedded sediment. Such large differences are not apparent in the acoustic velocity measurements made to date. This in itself may suggest that some geophysical tools are more effective in resolving certain fabric and lithological characteristics of sediment. Much data, however, remain to be analyzed and the differences in sediment parameters, as measured by different methods, require detailed study.

Axial and radial thermal properties were determined on board ship for up to 13 samples from each of 8 cores representing a range of sediment lithologies. Very few measurements of thermal properties of seabottom sediments in the Arctic Islands exist in the literature and yet these values are essential in the performance design of any seabottom or subseabottom structure. The measurements themselves require sample recovery and are tedious to do in large numbers; hence any extended correlations of, for example, the acoustic velocity to thermal conductivity as suggested by Goss and Combs (1974) could provide a useful saving in survey time since acoustic velocities can be measured underway. New data on the water temperature profile and the water/sediment temperature gradient were obtained in at least 12 stations throughout the work area. Many of these sites fill major gaps in the distribution of oceanographic summer bottom-water temperature data, thus assisting in the appraisal of both thermal and hydrodynamic bottom conditions.

Acknowledgments

The Atlantic Geoscience Centre at Bedford Institute of Oceanography supported this cruise in many ways: by arranging for ship time and by providing many of the scientific support personnel and equipment on the *C.S.S. Hudson*. Dr. D.I. Ross, K.S. Manchester, Dr. R.K.H. Falconer and their staff generously managed much of the pre-cruise preparatory work.

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Project 740068

S.H. Richard
Terrain Sciences Division

Field investigation and mapping of the surficial geology of the Huntingdon (31 G/1) map-area have been completed. A study of the bedrock geology and mineral resources of the map-area was carried out for the Geological Survey of Canada by Wilson between 1925 and 1941 and the results were published as part of her Ottawa-Cornwall geological map and report (Wilson, 1946). The unconsolidated Quaternary deposits in this area are well developed in surface extent and thickness over the Paleozoic limestone and sandstone platform of the St. Lawrence Lowlands and over the northern foothills of the Adirondack Uplands. The bedrock outcrops mainly in the northern foothills of the Adirondack Uplands in the southeastern part of the map-area, in a small dolomite and limestone upland west of Huntingdon in the central part of the map, and in the St. Lawrence Lowlands south of the Beauharnois Canal between St-Stanislas-de-Kostka and St-Louis-de-Gonzague in the northeastern part of the map-area near Valleyfield.

Glacial Deposits

Glacial deposits constitute about one-third of the land surface of the Huntingdon map-area. They occur mainly as an undulating to rolling or hummocky mantle of grey, calcareous, silty or sandy, compact ground moraine, mainly derived locally, covering flat lying bedrock. In the southeastern part of the map, where the glacial deposits overlie siliceous beds of Postdam Sandstone, the till is noncalcareous or only slightly calcareous. Till plains are commonly small in extent; the largest one lies in the southeastern part of the map-area at the foot of the Adirondack Uplands between Ormstown and the New York State border. A smaller one occurs just west of Huntingdon covering the top and slopes of a small tabular dolomite and limestone upland. Most of the till occurs as small areas of low relief, hummocky or ridged moraine separated from each other by lower areas of clay flats, marshes, and peat bogs, or by low sand dune plains. Till also commonly occurs as single, isolated mounds or knolls rising, sometimes barely, above the surface of the intervening shallow basins filled with marine sediments. Some small areas of surface till are underlain at or near the surface by tabular outcrops of flat lying sedimentary bedrock.

West of Port Lewis, Maplemore, and Trout River, the morainal deposits occur mainly as parallel till ridges up to one mile long and are nearly all trending northeast-southwest (Fig. 97.1). These are well developed west of Rivière-Beaudette and between St-Anicet and Fort Covington in the western part of the map-area. They are not present east of Rivière-Beaudette and Beaver Crossing. Their usual northeast-southwest trend is roughly perpendicular to the trend of the numerous drumlins that are developed on the west side of St. Lawrence River. Two of these glacial ridges are more prominent and better developed than the others: one is located in the northwestern part of the map-area at Rivière-Beaudette; the other is located at Beaver Crossing in the southwest corner of the map-area (Fig. 97.1). Many borrow pits dug into both of them show that they are composed of several different lithological units.

In the Beaver Crossing ridge, the most abundant material is a thick unit of glaciofluvial, unfossiliferous outwash gravels and sands up to 25 m (80 feet) thick. In the Hunsinger pit, the largest active gravel pit in the ridge, 12 to 18 m (40 to 60 feet) exposures show beds of gravels and cobbles alternating with beds of pebbly sand on the southeastern side of the ridge (Fig. 97.2). The many large boulders and the small degree of sorting in the lowermost gravel bed strongly suggests an ice-frontal origin. The uppermost beds are generally more sandy and in one place include a faulted and folded lens of unfossiliferous, interbedded, bluish grey clay and silt and buff sandy silt.

In the same gravel pit, a section cut into the northwestern side of the morainal ridge exposes deformed beds of unfossiliferous outwash sands and gravels overlain by, and in some places replaced by, severely folded and faulted beds of fossiliferous marine clays, silts, and sands (Fig. 97.2). Many parts of this material contain pebbles and coarse sand distributed throughout, and some of the material is compact and till-like in appearance. The fossils found in this material, *Balanus hameri* and *Hiatella arctica* (all fossils were identified by the author), are not in original growth position but appear to have been disturbed, broken, and redistributed throughout the unit. Farther south along the same section this fossiliferous, deformed unit wedges out and is replaced by undisturbed fossiliferous coarse sands and fine gravels interpreted as beach deposits. These beach sands and gravels lie at about 10 to 12 m (35 to 40 feet) below the ridge crest in horizontal beds; the marine fossil shells present include *Balanus hameri* and *Hiatella arctica.*

Higher in the unfossiliferous glacial outwash gravels and sands, in the same section on the northwest side of the morainal ridge, are exposed isolated pockets of compact diamicton and of fossiliferous diamicton (Fig. 97.3). Above these are fossiliferous marine beach gravels and sands which cap the crest and sides of the ridge and vary in thickness between 2 and 6 m (7 and 20 feet). The shells in this unit are undisturbed and commonly are found in growth position and include *Macoma balthica, Mytilus edulis, Hiatella arctica,* and *Balanus crenatus.*

In the Vallée pit (Fig. 97.2) located in the same ridge but 1 km to the northeast, the same varied lithological units are present. The most abundant material

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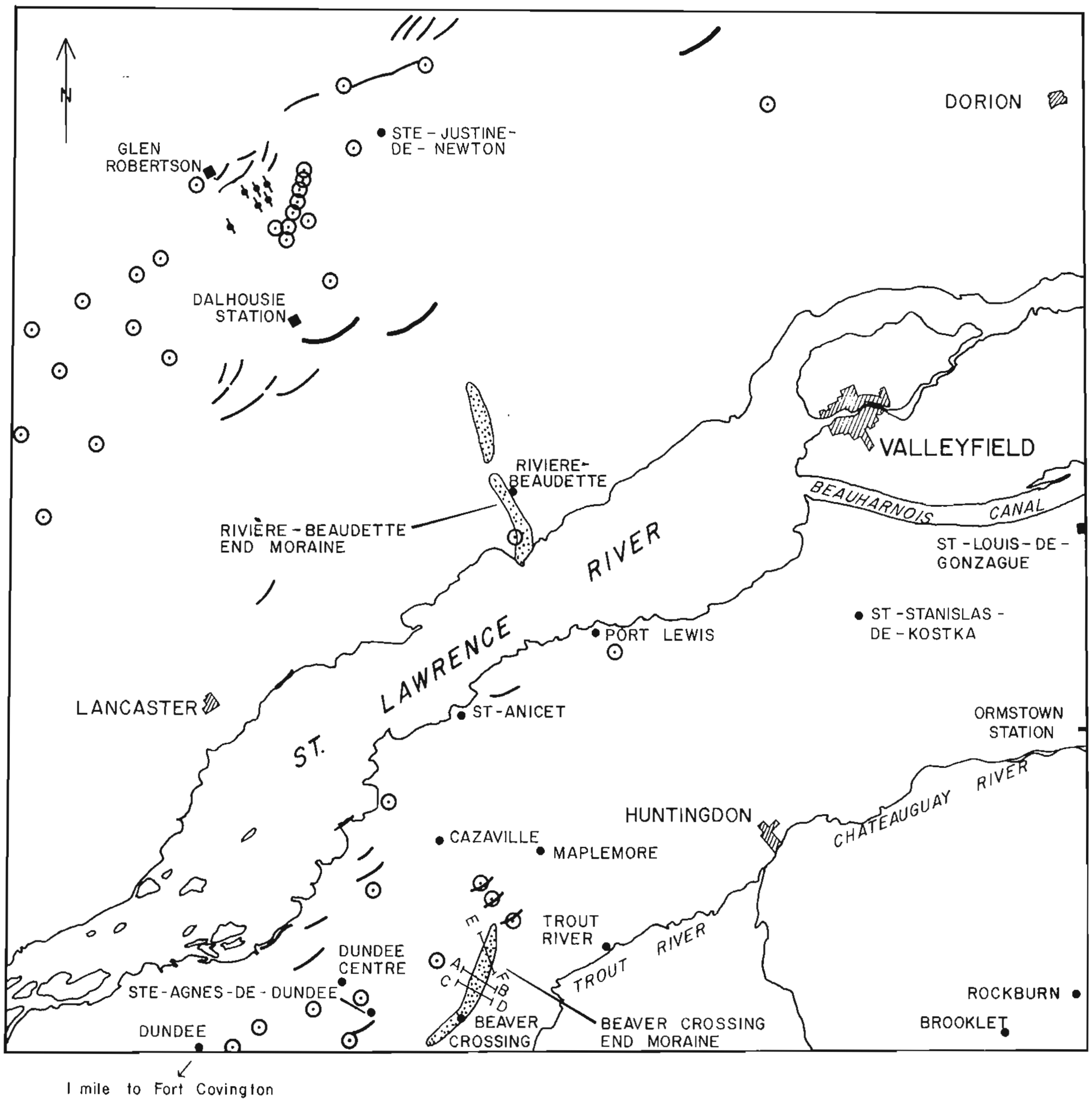

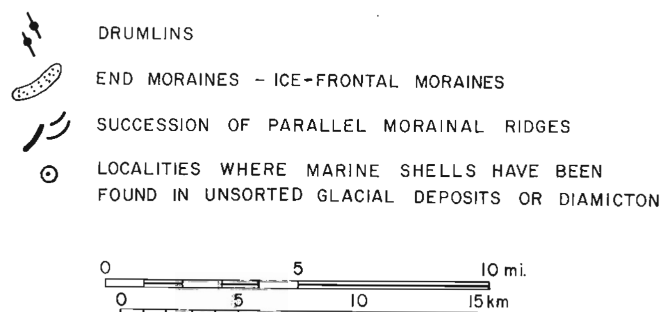

Figure 97. 1.

Location map of Valleyfield-Huntingdon area showing positions and trends of glacial features such as drumlins, major end moraines, successions of parallel morainal ridges, and localities where marine shells have been found in unsorted glacial deposits or diamictons. A-B, C-D, and E-F are cross-sections through the Beaver Crossing end moraine (*see* Fig. 97. 2).

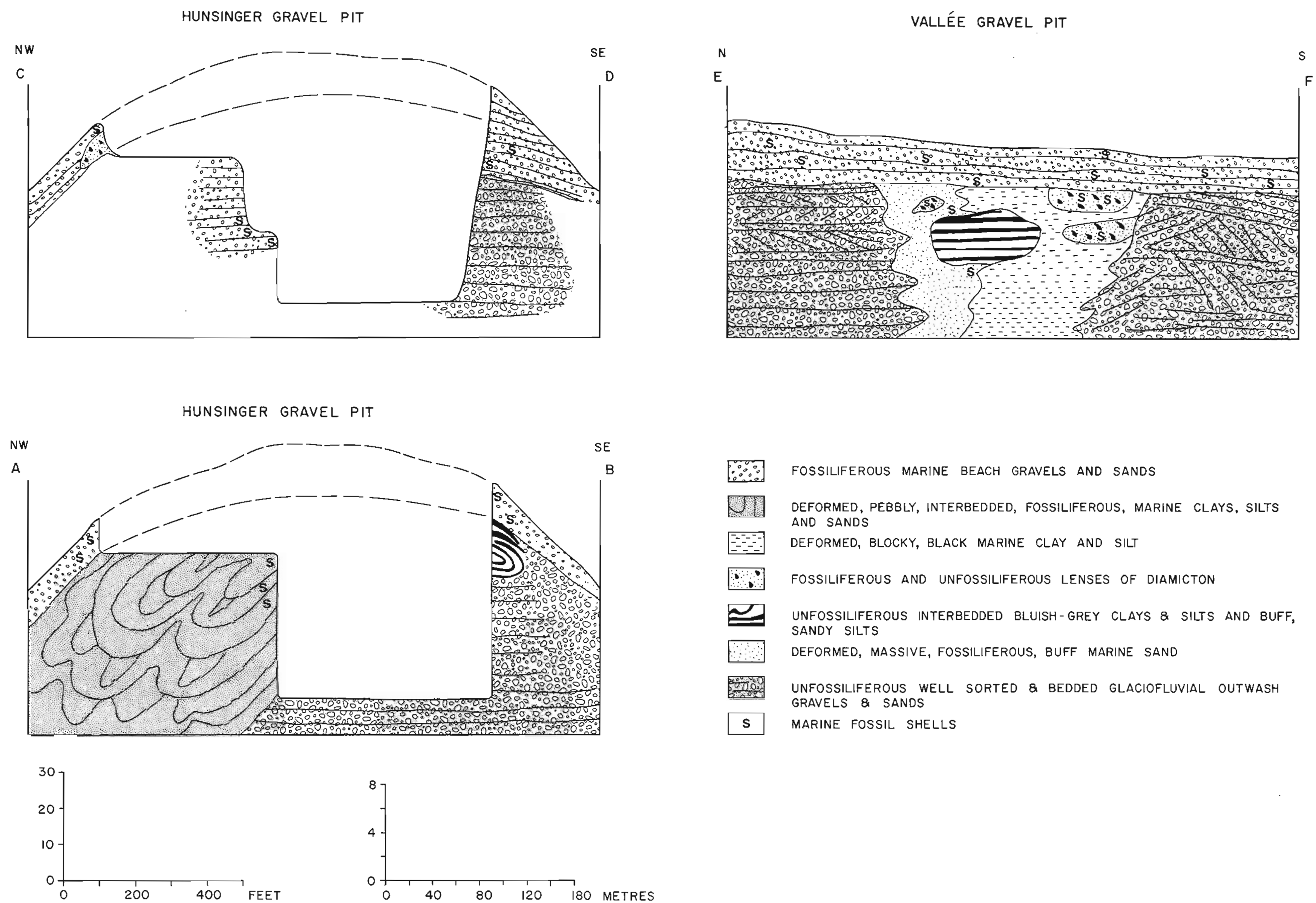

Figure 97. 2. Schematic section diagrams through Beaver Crossing end moraine showing nature and position of lithological units exposed in Hunsinger and Vallée gravel pits near Beaver Crossing, Huntingdon County, Quebec (*see* Fig. 97. 1 for location).

Figure 97. 3

Lens of fossiliferous unsorted material exposed inside the core of the Beaver Crossing ice-frontal moraine, 1 km northeast of Beaver Crossing, Huntingdon County, Quebec.

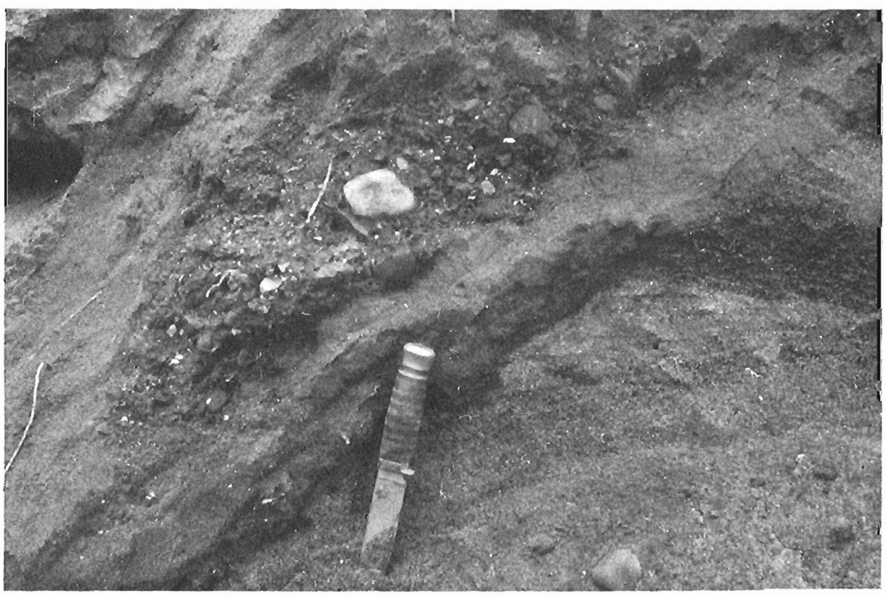

Figure 97. 4

Lens of fossiliferous clayey unsorted material included in and completely surrounded by beds of deformed medium and coarse marine sands. The photo was taken 2 km northeast of Beaver Crossing, Huntingdon County, Quebec.

is unfossiliferous outwash gravels and sands up to 12 m (40 feet) thick. In one part of the exposed section topset and foreset bedding in well sorted sand suggest a deltaic paleoenvironment. As in the Hunsinger gravel pit, this sand unit wedges out towards the northeast and is overlain and gradually is replaced by a thick unit of dark grey to black massive marine clays and silts which was deformed by faulting and folding. Lenses of fossiliferous compact diamicton occur within this unit and also near the top of the glaciofluvial deposits which overlie and gradually replace this unit towards the northeast (Fig. 97. 4). A lens of unfossiliferous interbedded clays, silts, and fine sands, similar to those present in the Hunsinger pit, occurs in the upper part of this section within the black massive marine clay and silt unit and just 1 to 2 m (3 to 6 feet) under a fossiliferous marine beach sands and gravels unit capping the ridge (Fig. 97. 5).

Sections exposed in the numerous borrow pits throughout the Rivière-Beaudette morainal ridge show the presence of essentially the same lithological units with the same stratigraphic relations. Between these two morainal ridges and the St. Lawrence River, marine shells have been found in hummocky or ridged glacial deposits at many localities (Figs. 97. 1 and 97. 6). Similar occurrences were noted throughout the Cornwall and southern part of the Alexandria map-areas on the west side of the St. Lawrence River (Fig. 97. 1).

A typical example of these occurrences is about 1 km north of Fort Covington, New York near the United States border in an area of well developed hummocky moraine called the Fort Covington moraine by MacClintock and Stewart (1965, p. 6). Here a 2 m (7 foot) road-cut through a morainic knoll exposes fossiliferous diamicton. This diamicton, considered to be till, is the only sediment present throughout this small section. *Mya arenaria* and *Hiatella arctica* were recovered from an elevation of 50 m (165 feet) a. s. l.

East of the Beaver Crossing morainal ridge and of the Fort Covington moraine, a large number of fresh exposures in the glacial till have been found which do

Figure 97. 5

Lens of unfossiliferous interbedded, bluish grey clay and silt and buff, sandy silt exposed inside the core of the Beaver Crossing end moraine, 2 km northeast of Beaver Crossing, Huntingdon County, Quebec.

Figure 97. 6

Marine shells exposed in the clayey matrix of a ground moraine deposit behind or west of the Beaver Crossing ice-frontal moraine, 1. 5 km southwest of Ste-Agnès-de-Dundee, Huntingdon County, Quebec and 5 km east of Fort Covington, New York State, U. S. A.

not include shells. This striking contrast between areas of abundant fossils and areas of no fossils tends to suggest that the glacial deposits in these two parts of the map-area are of a different age and were laid down under two different environments. The presence of marine shells seems to indicate that the Beaver Crossing and Rivière-Beaudette moraines and the smaller moraines to the west were formed in one of three possible ways: 1) Formation took place between 12 000 to 11 800 years B. P. (Denny, 1974, p. 36; Elson, 1969; McDonald, 1968; Prest, 1970) with the ice sheet retreating towards the north from the Adirondack Foothills while maintaining an active ice front in the advancing Champlain Sea. 2) They were formed between 11 300 and 11 000 years B. P. by a late post-Fort Covington ice readvance (Terasmae, 1965, p. 36; Richard, 1975) into the Champlain Sea basin after the marine waters had occupied the area for a period of about 1000 years. 3) The fossiliferous diamictons are in no way related to glacial activity but are the product of normal marine reworking (accompanied by slumping) of glacial deposits.

If the shell occurrences are related to glacial events, ^{14}C dating should make it possible to determine to which one of the two above mentioned geological time periods these glacial deposits belong. If the diamictons are not of glacial origin, the ^{14}C dates merely will provide maximum ages for retreat of the Champlain Sea below the elevation of the sample sites. The fact that there are no marine fossils associated with the glacial deposits in the central and eastern parts of the map-area could be taken to indicate that they predate the formation of the Champlain Sea and hence are at least 12 000 years old. This could also be interpreted as indicating that this part of the map-area was deglaciated earlier than the western part.

Marine Sediments

Marine sediments deposited in the Champlain Sea, following the withdrawal of Wisconsin ice, are more widespread than the glacial deposits and constitute approximately one-half of the land surface of the area surveyed. These sediments consist mainly of grey, massive, blocky, unctuous, calcareous, fossiliferous marine clays and silts. They occur to an elevation of 50 m (170 feet) a.s.l. in the central part of the map-area and to an elevation of 60 m (200 feet) in the southern and northwestern parts. These deposits form gently sloping clay plains whose surfaces are primary surfaces of marine sedimentation. The most extensive deposits are found mainly in the St. Lawrence River valley on both sides of the river west of Valleyfield and in the Trout and Chateauguay river valleys in the Huntingdon and Ormstown area. In these areas, the marine clay plains are fairly extensive but they are commonly interrupted by low mounds, knolls, or hummocks of glacial till rising above the surface of the clay flats.

Two fairly large bodies of marine sand are found in the map-area; one is located just north of Rivière-Beaudette, and the other extends north and south of the town of Cazaville in the western part of the map. The sand is well sorted and medium to fine grained, and the presence of marine shells indicates that it was deposited during the submergence of the area by the Champlain Sea. The lowest occurrence of marine shells in shoreline sands of the Champlain Sea in the region north of the St. Lawrence River is 1.5 km northwest of Rivière-Beaudette, where shells of *Mya truncata, Balanus hameri, Hiatella arctica,* and *Macoma balthica* were recovered from an elevation of 58 m (190 feet) a.s.l. Two kilometres northwest of Cazaville shells of *Macoma balthica* and *Mya truncata* were recovered from sands at an elevation of 55 m (180 feet) a.s.l., the lowest elevation at which marine shells have been recovered from shoreline sands in this part of the St. Lawrence River valley south of the river.

During submergence by Champlain Sea, several fossiliferous sand and gravel beach ridges were constructed around the higher parts of the emerging glacial topography. The most numerous and best developed of these abandoned beaches are found on the northern slopes of the Adirondacks in the southeastern part of the map-area up to an elevation of 160 m (530 feet) at Brooklet near Rockburn just north of the New York State border. No marine shells have been recovered from these raised beaches at a higher elevation than 87 m (285 feet) a.s.l. The highest marine shells recovered in a raised marine beach in the map-area were found in the beach gravels and sands at the crest of the Beaver Crossing end moraine at an elevation of 87 m (285 feet) a.s.l.

Post-Champlain Sea Freshwater Sediments

Below the elevation of 50 m (170 feet) a.s.l., a veneer up to 2 m thick of freshwater alluvial or lacustrine sand has been found on both sides of St. Lawrence River lying over the marine clays. Nonmarine shells have been recovered from these sands at several localities and will be submitted for identification and radiocarbon dating. These deposits formed following withdrawal of Champlain Sea at a time when lowermost parts of the map-area were occupied by a body of freshwater. This body of water has been called Lampsilis Lake by Elson (Elson, 1960; Prest, 1970, p. 723-24; Terasmae, 1965, p. 27, 37), after fossil shells of the mollusc *Lampsilis Siliquoidea* found in its sediments.

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98. ARCHEAN SUBAQUEOUS AUTOCLASTIC VOLCANIC ROCKS, ROUYN-NORANDA AREA, QUEBEC: CLASSIFICATION, DIAGNOSIS AND INTERPRETATION

Erich Dimroth[1]
Regional and Economic Geology Division

Introduction

Detailed interpretation of volcaniclastic rocks has always been one of the central problems of Archean research. Rocks in most of the Rouyn-Noranda area show little schistosity and suffered only very low grade metamorphism. Consequently, their primary textures and structures are relatively well preserved. These circumstances offer ideal conditions for the study of volcaniclastic rocks. In this report, I will discuss the

Figure 98. 1.

In situ brecciation of a flow-ribboned band in rhyolite flow breccia. The monolithologic nature of the breccia, *in situ* brecciation, and transitions of brecciated and massive facies are the most important characteristics of flow breccias. Rhyolite flow, south of Lake Dufault.

[1]Sciences de la Terre, Université du Québec à Chicoutimi, Chicoutimi, Qué.

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Table 98.1

Diagnosis of fragmentation mechanism

	GEOMETRY	OTHER FACIES IN SAME ERUPTION UNIT	RELATION TO OTHER FACIES	FRAGMENTS	FRAGMENT SHAPE	SPECIAL FEATURES OF FRAGMENTATION
Friction breccias	Breccia forms part of mappable flow or dome.	Massive and flow-banded facies always present. Hyalotuff may be present.	Complex (*see* Ref.) transitions of breccia into massive and flow-banded facies.	Monomict, except for foreign blocks picked up at base of flow.	Angular to irregularly corroded	*In situ* brecciation, perlites occur.
Hyaloclastites	Breccia forms top (rarely base) of lava flow.	Massive and pillowed facies. Hyalotuff may be present.	Breccia overlies massive or pillowed facies. Transitions in massive or pillowed facies.	Monomict. Pillow fragments, globules, granules (*see* references and Table 98.3).	Depending on vesiculation (*see* text and Table 98.3).	*In situ* brecciation, perlites.
Magmatic explosion	Fall-back: Cones close to fissure; flows and fall-out: sheets.	None; Breccias may be precursor to lava flows.	None.	Polymict. Essential: Pumice and phenocrysts. Accidental: Various.	Essential: Pumice lumps, pumiceous matrix, bubble-wall shards. Accidental: Angular, rarely rounded.	No *in situ* brecciation, no perlites.
Hydro-magmatic explosion	As	magmatic	explosion	Polymict. Essential and accidental fragments not readily distinguished.	Angular; re-entrant angles common. Some fragments of subvolcanic rocks may be rounded.	No *in situ* brecciation, no perlites.
Steam explosion	Fall-back: local piles of breccia.	Fall-back breccia grades into and is rooted in the source rock.	Top: Polymict fall-back breccia. Middle: Breccia mostly composed of source rock. Base: Brecciated source rock.	Variable stratigraphically.	As hydromagmatic explosion.	Brecciated source rock shows *in situ* brecciation.

REFERENCES: Parsons (1969), Heiken (1972), Dimroth *et al.* (1975a, b), Carlisle (1962).

classification, diagnosis and interpretation of autoclastic volcanic rocks, that is of flow breccias and pyroclastic rocks.

Existing classification of subaerial autoclastic rocks (for example Parson, 1969) are not easily applied to the subaqueous environment. On the other hand, present classifications of subaqueous autoclastic rocks are very incomplete. Therefore a new system of the classification of subaqueous autoclastic rocks has been developed (Dimroth *et al.*, 1975); this system is based on a sharp differentiation between the mechansims of fragmentation and of emplacement of autoclastic rocks. It will be presented in table form, the following text serving mainly to illustrate Tables 98.1 to 98.3.

Mechanisms of fragmentation

Basically, fragmentation of autoclastic rocks can take place by three mechanisms:

1) Friction breccias form where parts of an actively moving flow (generally the upper surface layer) deform by brittle fracturing where other parts (generally the interior) deform by plastico-viscous flowage.

2) Hyaloclastites form where fragmentation is due to thermal strain within a differentially cooled solid.

3) Fragmentation by explosion is due to rapid expansion of vapor. Three cases must be distinguished.

3a) Magmatic explosion is caused by rapid exsolution and expansion of gases dissolved in a magma.

3b) Hydromagmatic explosion is caused by rapid expansion of vapor formed by contact of sea water and an actively moving lava.

3c) Steam explosion is caused by rapid expansion of vapor formed on contact of sea water with a cooling, solid, not actively moving magmatic body.

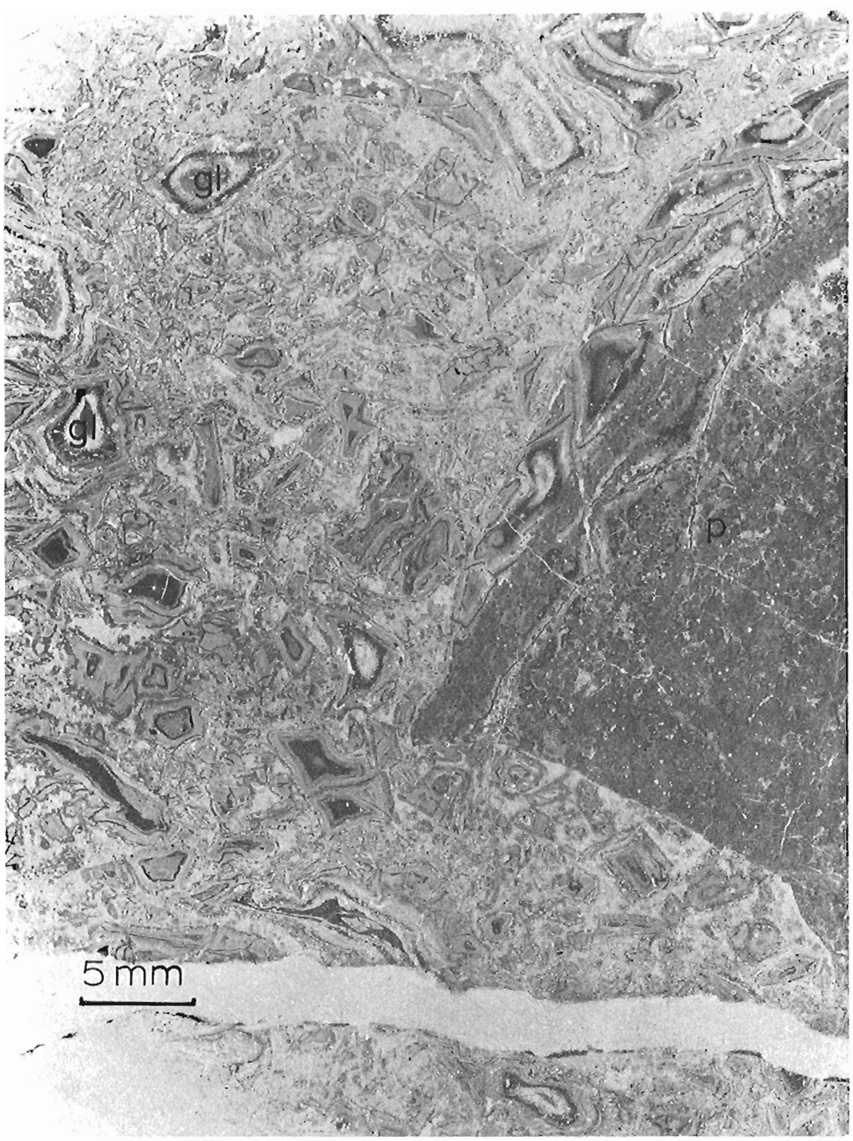

Figure 98.2.

Typical broken-pillow breccia: Pillow fragment (p), globules (gl) and granules. Section ONT-HY3.

Mixed mechanisms are common. Diagnosis of the fragmentation mechanisms is possible based on a combination of petrographic and geologic observations (Tables 98.1 and 98.3).

Mechanisms of emplacement

Four processes are responsible for the ultimate emplacement of autoclastic rocks:

1) Passive emplacement at the top or base of an actively moving lava flow.
2) Avalanching of fragments, for example at the front of an actively moving flow.
3) Pyroclastic flow: Emplacement by mass flow (turbidity current, grain flow, mud flow).
4) Pyroclastic fall: a) Fall-back of fragments emplaced (essentially) on a balistic trajectory and b) Fall-out emplaced by settling of pyroclastic fragments from a suspension of pyroclastic material.

Analysis of the geometry and of sedimentary structures permits diagnosis of the emplacement mechanism. However, diagnosis of pyroclastic fall-out from distal pyroclastic flows is not possible at present since settling from suspension is the main process of deposition in both cases. Both should show graded bedding. However, distal pyroclastic flows could show erosional bases which should be absent from pyroclastic falls. Furthermore, distal pyroclastic flows should show Bouma-cycles, whereas sedimentary structures should not follow in a regular sequence in pyroclastic falls. Current directions should be

Figure 98.3.

Detail of Figure 98.2. Note *in situ* brecciation (arrow), and the typical convex-concave shapes of sideromelane shards.

consistent and should be down-slope in pyroclastic flows, whereas this need not be the case in pyroclastic falls. These criteria have been obtained from a theoretical analysis of the problem; actual cases have not been studied.

Effects of sea-floor metamorphism

Different parts of one and the same flow do not, in general, devitrify in the same way; therefore most flow breccias appear to be polymict whereas in fact all their fragments are derived from the same lava. Differentiation of polymict and monomict breccias is simple in porphyritic rocks: All parts of one flow contain essentially the same proportion of phenocrysts having about the same size and shape. Repetition of flows containing similar phenocrysts is exceptional. Thus, breccias in which all fragments contain a similar proportion of similar phenocrysts are monomict even if the bulk composition of fragments is not the same. Characteristic devitrification effects permit recognition of monomict breccias in aphanitic rocks.

Basalts and andesites: Typically, pillows consist of three zones: (1) An outer zone of chilled glass devoid of quench crystals; during devitrification, this glass is replaced by chlorite, less commonly by quartz. (2) The outer zone grades abruptly into a chilled phase containing quench crystals of plagioclase. Albite spherulites nucleate at the quench crystals, producing a light weathering, feldspar-rich outer zone. Where numerous phenocrysts are present, the spherulitic zone is absent. (3) Zone 2 grades imperceptibly into microcrystalline basalt or andesite by a gradual increase of the size and number of crystallites. Simultaneously,

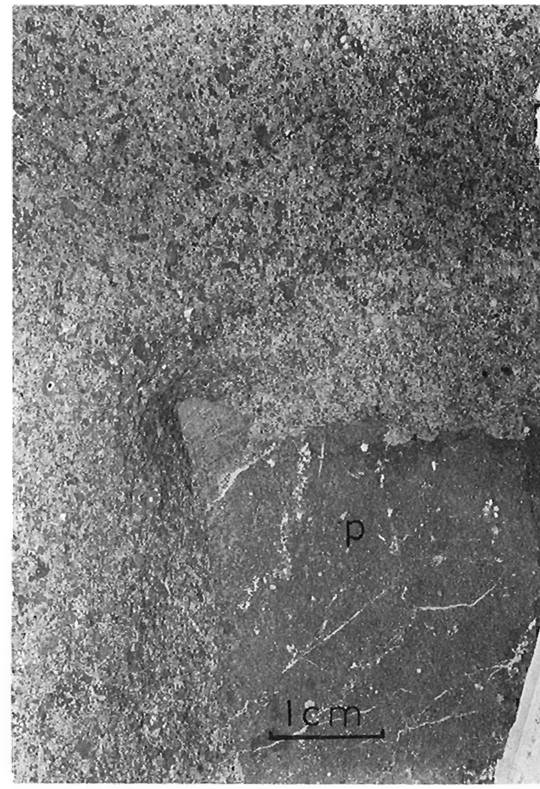

Figure 98. 4. Broken-pillow breccia with strongly vesiculated pillow fragment (p). Section HERE 2A.

Figure 98. 5.

Detail of Figure 98. 4. Note the irregular shape of the pumiceous shards in the matrix of the broken pillow breccia. Interpretation as a hyaloclastite is based on the presence of pillows, gradation into pillowed andesite downward and absence of sedimentary structures.

the quench crystal morphology is lost. Consequently, monomict basalt and andesite breccias (hyaloclastites and hyalotuffs) contain three types of fragments: (a) chloritized sideromelane shards, (b) feldspar-rich spherulitic fragments and (c) fragments of microlitic lava.

Rhyolites and dacites: Typically, the massive and flow-layered facies and the large fragments devitrify to a white-weathering rock mainly composed of quartz and albite, whereas their microbreccia matrix devitrifies to a darker weathering rock containing more chlorite and sericite. Both materials contain similar phenocrysts in similar proportions.

A network of silicified and albitized fractures may be superposed on both the matrix and fragments of rhyolite flow breccias.

Autoclastic rocks of Rouyn-Noranda area

The criteria outlined have permitted interpretation of the origin of autoclastic rocks in the Rouyn-Noranda area (Table 98.3). Few comments are required. Rhyolite flow breccias (type Ia) form parts of rhyolite domes and flows, and show complex relations to other facies described by De Rosen-Spence (1976) and Provost (*in* Dimroth *et al.*, 1975a, b). Individual rhyolite flows are readily mapped, because the size and proportion of phenocrysts remain approximately constant within flows, but vary from flow to flow. Hyalotuffs are hyaloclasstites that have been emplaced by mass-flow mechanisms. Petrographically, they resemble the hyaloclastic matrix of underlying broken-pillow breccia. They are monomict; where prophyritic, their phenocrysts are similar to phenocrysts of the underlying flow. Thus, they are believed to have formed where minor hydromagmatic explosion took place simultaneously with hyaloclastic fragmentation at the top of an actively moving flow or at the eruptive fissure. The hyalotuffs are part of the underlying flow.

Figure 98.6. Graded bedding in hyalotuff. Section B49-1A.

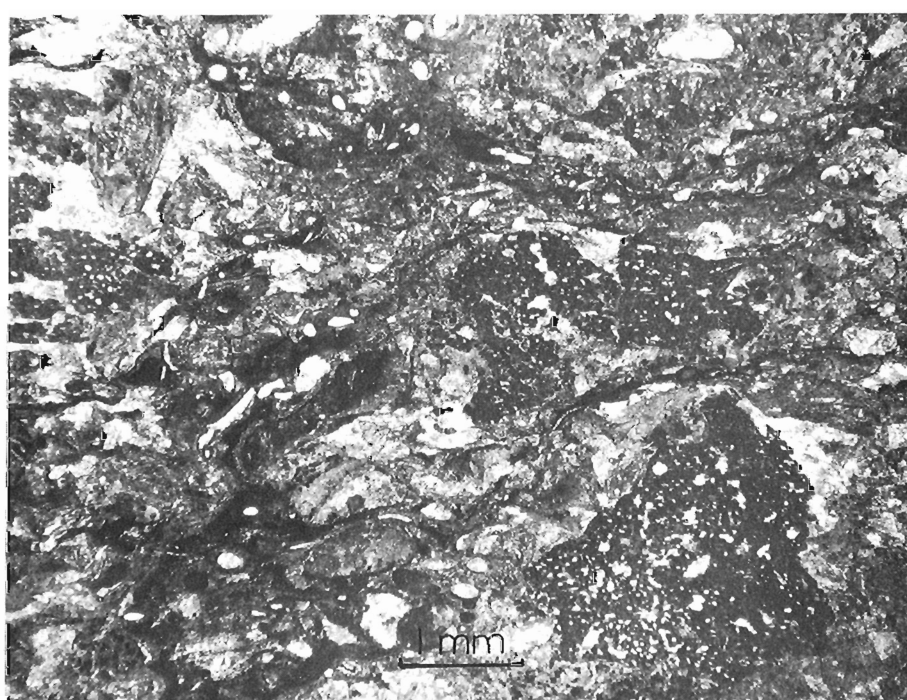

Figure 98.7.

Detail of Figure 98.6. Note the similarity of the fragment shapes to pumiceous shards (Fig. 98.5). The fragments contain phenocrysts of the same size as the underlying flow and flow breccia.

Table 98. 2

Emplacement mechanisms

	STRUCTURES	REFERENCES
Passive emplacement	Breccia grades into massive, flow-layered or pillowed part of the same flow. Complex flow structures may be present. No sedimentary structures. *In situ* brecciation.	Côté and Dimroth (1976 Provost (*in* Dimroth *et al.*, 1975a, b)
Avalanching	Large-scale foreset beds.	Provost (*in* Dimroth *et al.*, 1975a, b)
Pyroclastic fall-back	Local breccia piles. Poor layering; poor sorting; very rapid change of grain size. No *in situ* brecciation.	Dimroth *et al.* (1975a, b)
Pyroclastic flow	Sheets; sedimentary structures of subaqueous mass-flows and turbidites (Walker and Mutti, 1973). No *in situ* brecciation.	Tassé (*in* Dimroth *et al.*, 1975a, b).
Pyroclastic fall-out	*See* text.	

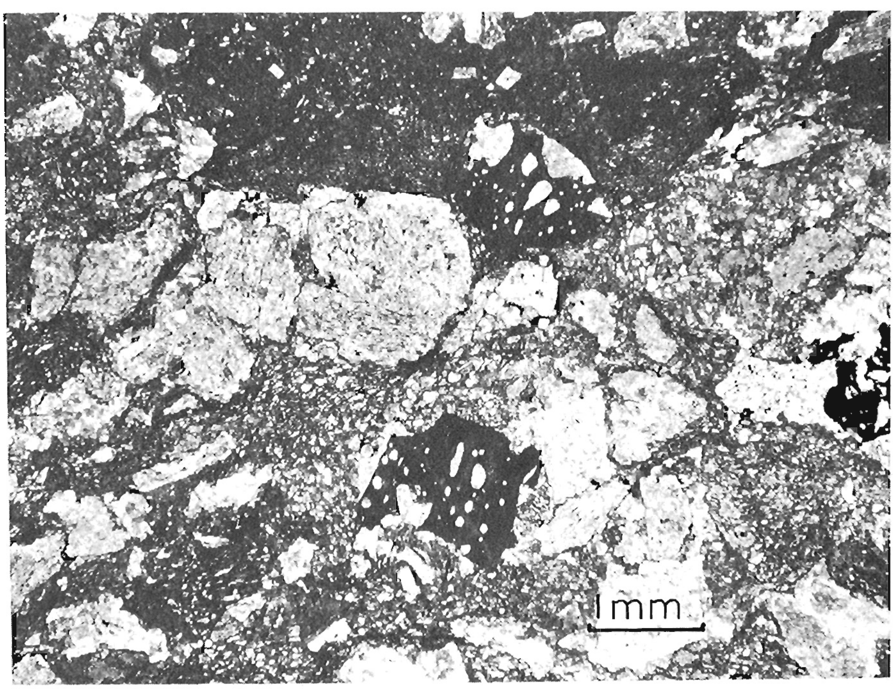

Figure 98. 8.

Pumice flow. Accidental fragments and feldspar phenocrysts embedded in a matrix of pumice. Graded bedding is present on a large scale, as are parallel and oblique laminations and other sedimentary structures. Section B28-17B.

Table 98.3

Autoclastic rocks observed in Rouyn-Noranda area

	ROCK TYPE	MECHANISMS	SPECIAL OBSERVATIONS	REFERENCES
I A.	Rhyolite flow breccia	Friction breccia; emplacement passive and by avalanching.	*See* references for details.	De Rosen-Spence (1976) Provost (*in* Dimroth *et al.*, 1975a, b)
I B.	Rhyolite hyalotuff	Minor hydroexplosion at top of advancing flow. Emplacement as pyroclastic flow.	Hyalotuff contains same phenocrysts as underlying flow, except where finer grained than phenocrysts.	Provost (op. cit.)
II A.	Isolated and broken pillow breccia	Thermal strain; emplacement passive.	Matrix is sideromelane shard hyaloclastite where pillows poorly vesiculated; is pumiceous hyaloclastite where pillows well vesiculated; imbrication.	Carlisle, 1962, Côté and Dimroth 1976
II B.	Sideromelane shard hyaloclastite	As II A.	Globules: spherical to edge-rounded bodies. Granules: sharp-edged, commonly knife-edged fragments of globules and pillows.	Carlisle 1962
II C.	Pumiceous hyaloclastite	As II A (minor hydromagmatic explosion as component in fragmentation process ??).	Odd-shaped vesiculated fragments, pillow rims, pillow fragments, bubble-wall shards.	Dimroth *et al.*, 1975a, b
II D.	Sideromelane shard and pumiceous hyalotuffs	Minor hydromagmatic explosion at top of advanced lava flows. Emplacement as mass-flow.	Generally derived from pumiceous hyaloclastite.	Dimroth *et al.*, 1975a, b
III A.	Pumice flows	Magmatic explosion, emplacement as pyroclastic flow.	Good marker horizons.	Tassé (*in* Dimroth *et al.*, 1975a, b)
III B.	Pyroclastic flow due to hydroexplosion	Hydromagmatic explosion. Emplacement as pyroclastic flow.	Some are good marker horizons.	Tassé
III C.	Pyroclastic fall-back	Hydromagmatic and steam explosion.		Dimroth *et al.*, 1975a, b

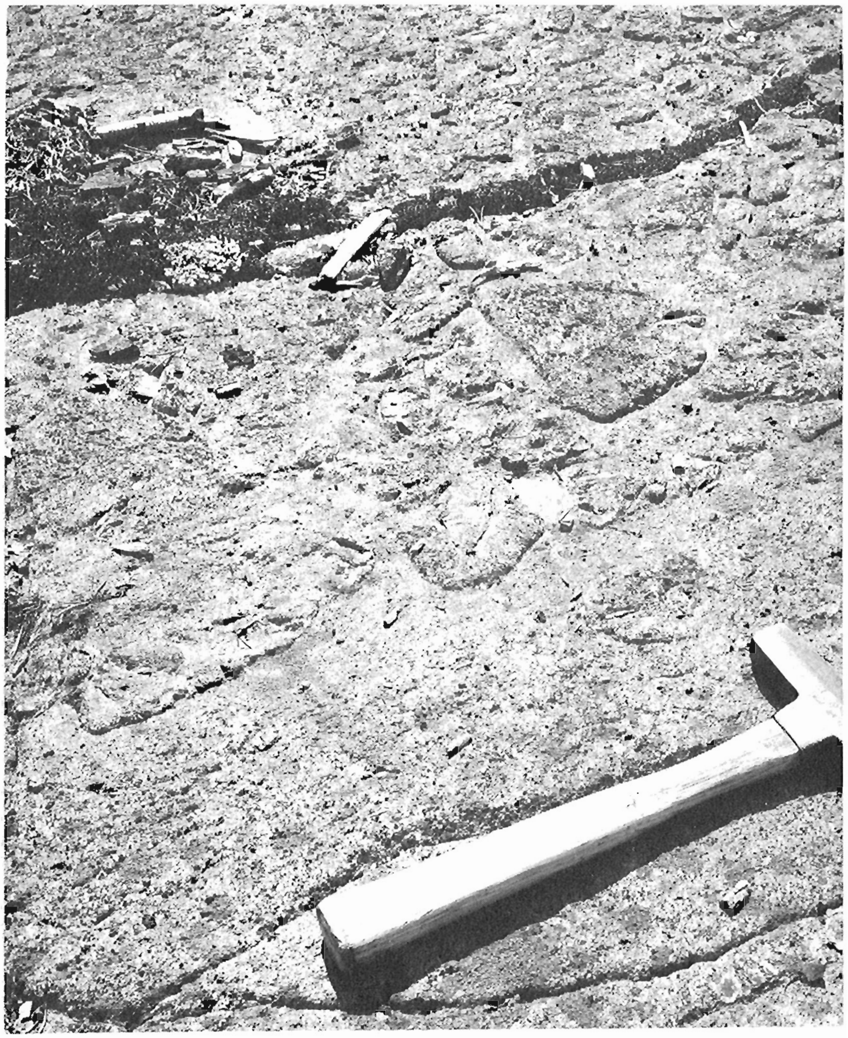

Figure 98.9.

Pyroclastic flow formed by hydro-explosion. Note the angularity of fragments. Graded bedding (normal and reversed), parallel and oblique laminations, and other sedimentary structures are present. Reneault.

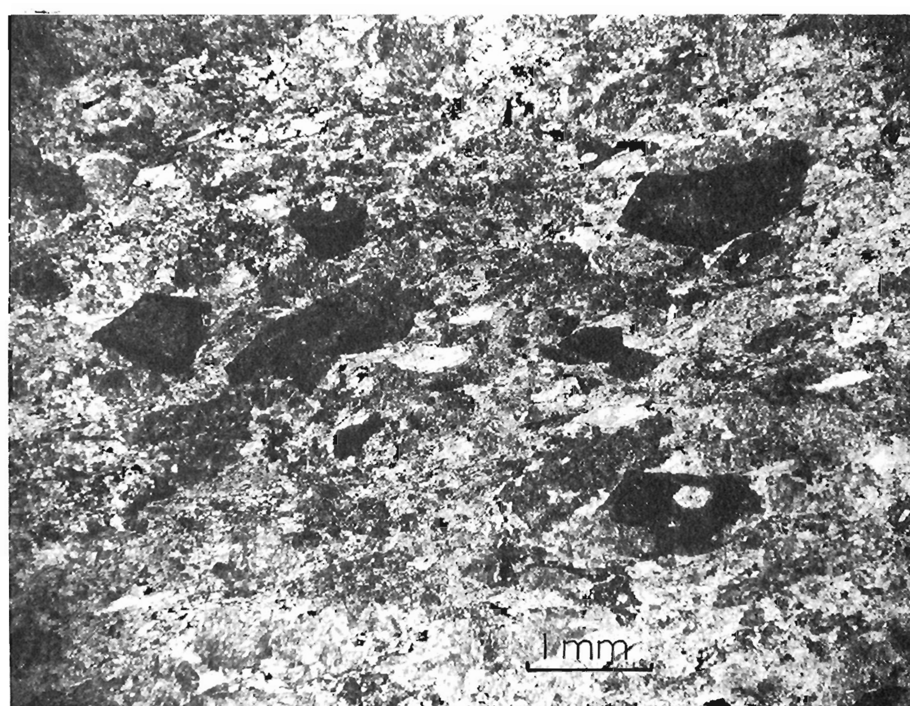

Figure 98.10.

Fine grained ash-tuff from same outcrop. Note angularity of fragments and absence of pumice. Section D11-6B.

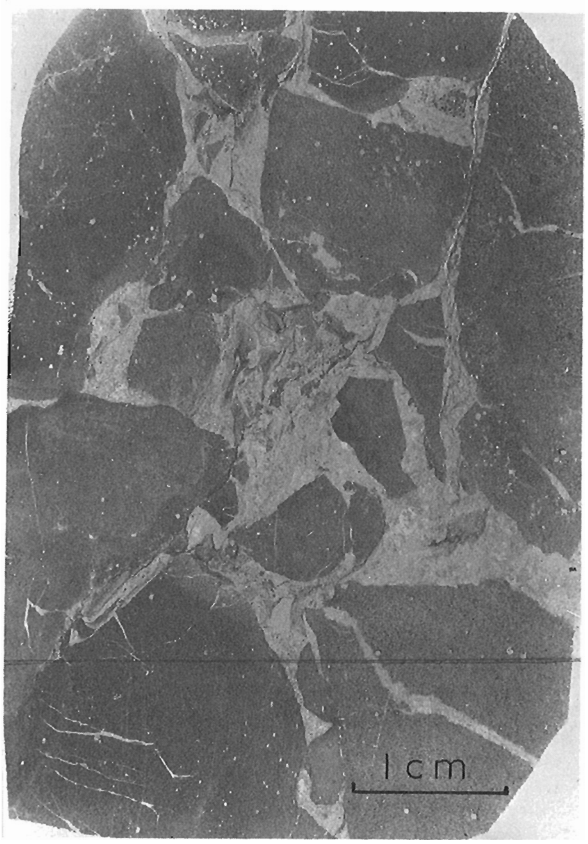

Figure 98.11. "Basaltic" hyaloclastite. This type of hyaloclastite overlies massive flows with thin hyaloclastic flow-top. A few shards of sideromelane are present between basaltic fragments. Section ONT-4.

Acknowledgments

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Project 760039

B.D. Loncarevic and R.K. Falconer
Atlantic Geoscience Centre, Dartmouth

During the 1976 cruise of *C.S.S. HUDSON* to the Arctic an oil slick was observed off Baffin Island. The occurrence is off Scott Inlet, near the northern edge of the Hecla and Griper Bank in the vicinity of 71°24'N, 70°15'W (Fig. 99.1). It is possible that this oil slick is caused by a seepage along edges of flat-lying sedimentary beds, where these have been truncated and eroded by glaciers emerging from Scott Inlet. The economic significance of the observation of this oil slick is not clear. The laboratory analyses are not complete and only preliminary results are available. Nevertheless, it is considered important to report this discovery now.

Regional investigations in Baffin Bay include studies of surficial geology, shallow stratigraphy, and deep crustal structure. As a part of this program *C.S.S. HUDSON* carried out a seismic reflection profiling line in the north-south direction over the Hecla and Griper Bank on 10 August, 1976. The sea was covered by 6-8 tenths of one year old ice fragments. The ship was positioned using a satellite navigator aided by frequent fixes on three grounded icebergs. The largest of these was almost 100 m high above the sea surface and over 200 m in length of waterline. Its position was fixed as 71°20'N, 70°22'W.

Shortly after the beginning of the seismic line a large chunk of ice tangled the four cables streamed astern the ship: air gun, hydrophone eel, magnetometer and towed side-scan sonar. The ship had to stop to unravel the spaghetti-like mass. While working on the quarter deck several people noticed a number of small oil slicks around the ship. Later the bridge reported noticing these oil slicks ahead of us, so they were not caused by any leak from the ship.

Small blobs of light material were coming to the surface at a rate of one every three to five minutes. The material had very low viscosity and dispersed into a thin film quickly covering an area of up to 1-2 m^2. No bubbles were observed so it is not likely that there was any gas. When dispersed on the sea surface, the material produced a multicoloured sheen, usually associated with liquid hydrocarbons. The strong southward current and wind driven surface circulation was moving the slicks away and they could be observed over an area of a couple of square kilometres. Concentrations could be observed locked in small pools or embayments between ice flows.

Some ten hours later, on completion of the seismic line, we returned to the area where the slicks were first observed. Although it was the middle of the night it was easy to see the slicks in the Arctic twilight, in the same position where they were first observed. An *ad hoc* sampling program was quickly organized: two short bottom cores, two mid-water Nisken bottle samples

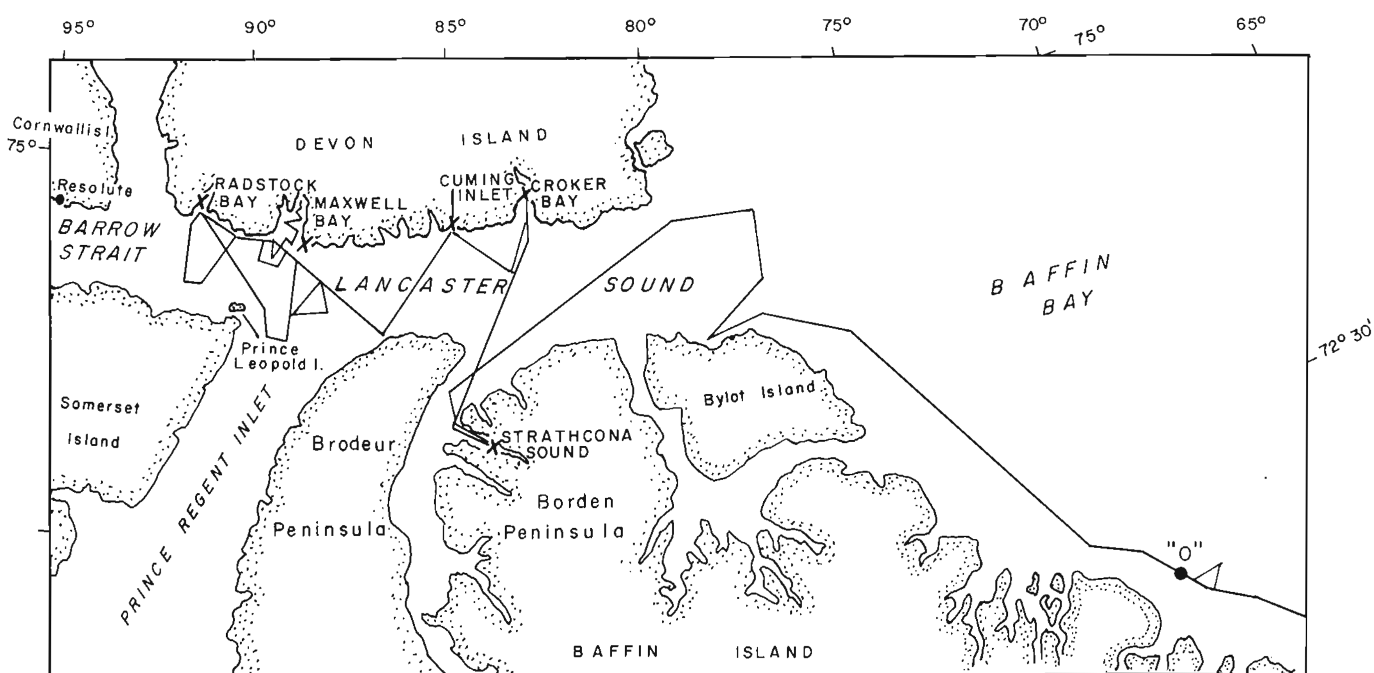

Figure 99.1. Oil slick was found at point marked with an "O" off the east coast of Baffin Island.

From: Report of Activities, Part A;
Geol. Surv. Can., Paper 77-1A (1977)

and some sea surface samples were taken. Sampling the sea surface was difficult. The first attempts with a bucket were unsuccessful. Dropping the bucket in the water quickly dispersed the slick. Samples were eventually obtained by lowering a bunch of tissue paper and skimming the surface thus soaking up slicks.

On completion of the Arctic program *C.S.S. HUDSON* passed through the area on its way south 25 days later on 4 September 1976. The oil slicks were detected again. By reference to the grounded icebergs it could be determined that the position of slicks had not changed over the 25-day period in spite of the southerly current.

The analysis of the cores did not detect any hydrocarbons. The analysis of the sea water indicates an order of magnitude higher concentration of hydrocarbon than observed as background level on the Nova Scotia Shelf (40-60 ppm). The gas chromotographic studies of surface samples continue.

Although we refer in this report to "oil slicks", evidence regarding the nature of these hydrocarbons is inconclusive at the present time. The three possible causes of the observed slicks may be: (a) natural crude; (b) animal (whale?) oil; (c) refined oil (snowmobile lost through ice, a sunken boat, or a crashed plane?). The fact that the slick was observed over a large area would argue against (b) and (c). RCMP detachment in Resolute did not have any report of a recent crash or sinking in the area but search of historical records was not undertaken. The laboratory investigations continue but it is questionable whether the samples collected will be sufficient to provide definitive answers.

ERRATA

Report of Activities, Part C
Geol. Surv. Can., Paper 76-1C

Paper 76-1C, rep. 32, p. 177, left hand column, equation should read:

$$\tan \tfrac{1}{2}E = \sqrt{\frac{\sin (s - d) \sin (s - f)}{\sin s \sin (s - e')}}$$

Paper 76-1C, rep. 23, pages 124 and 125 are interchanged

AUTHOR INDEX

Bracketed numbers refer to individual reports within this publication