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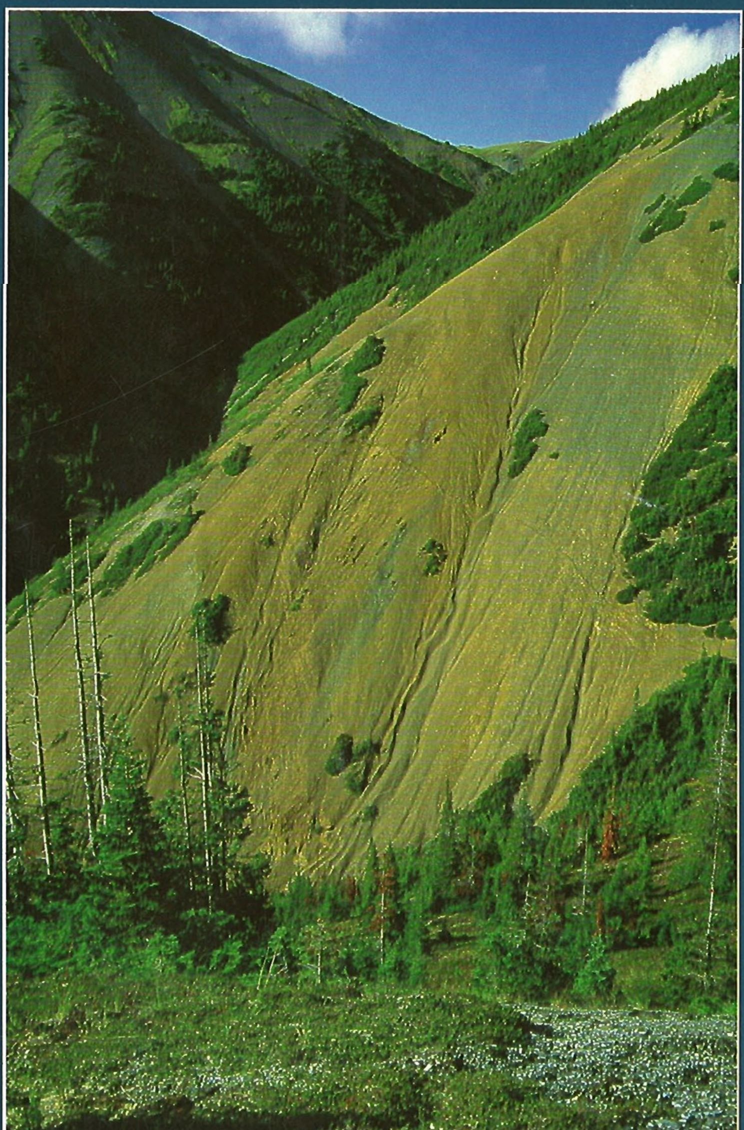

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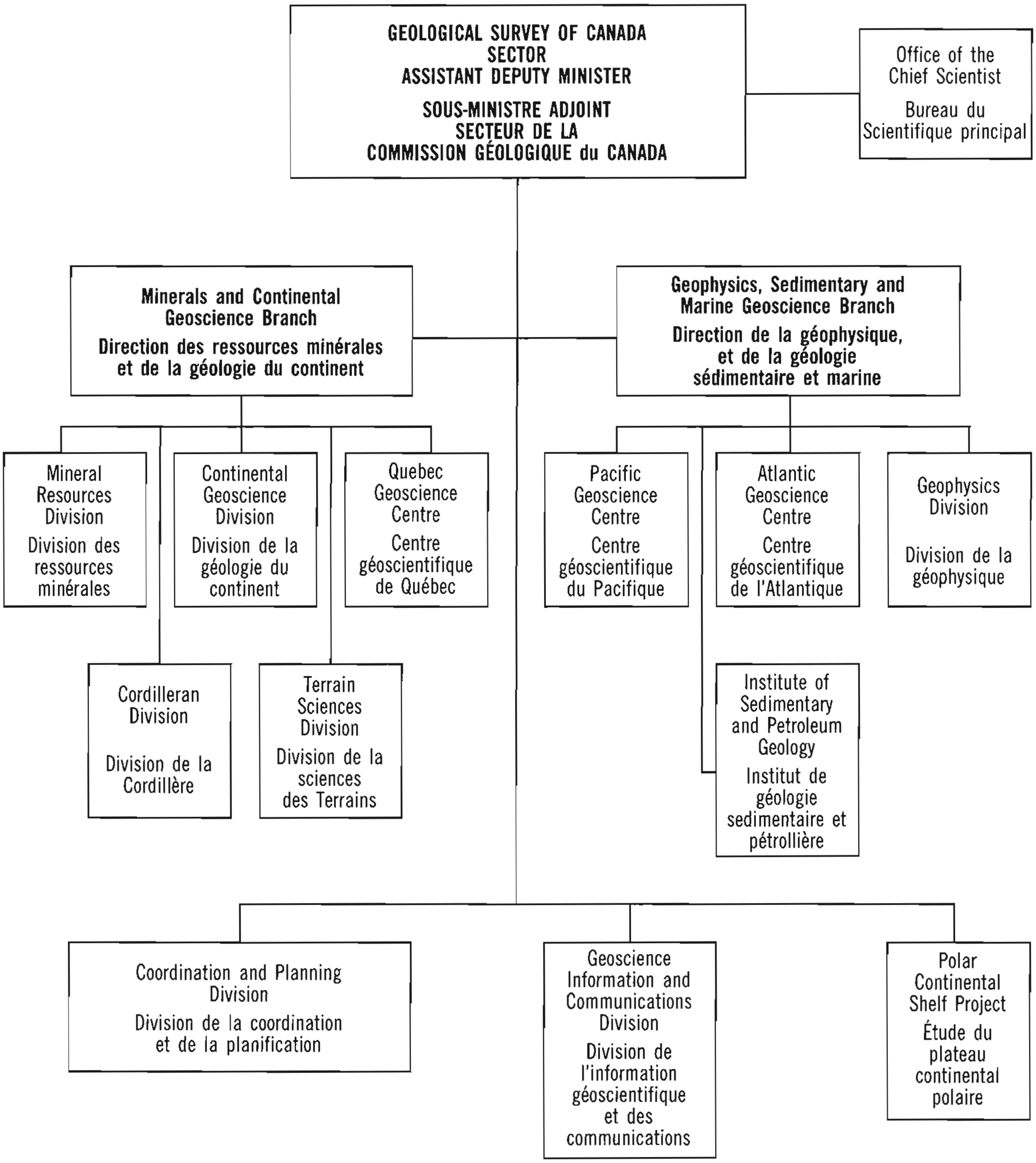
Cover description

View of a pale brown weathering barite "kill zone", Driftpile deposit, Gataga district, Northeastern British Columbia. (Photo by Suzanne Paradis, August 1994; GSC 1994-788 B).

Description de la photo couverture

Vue d'une «zone morte» dépourvue de végétation, de barytine à couleur d'altération brun pâle, au gisement Driftpile, district de Gataga, dans le nord-est de la Colombie-Britannique. (Photo prise par Suzanne Paradis, août 1994; GSC 1994-788 B).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Progress report and field activities of the Fraser Valley hydrogeology project, British Columbia

Brian D. Ricketts
Cordilleran Division, Vancouver

Ricketts, B.D., 1995: Progress report and field activities of the Fraser Valley hydrogeology project, British Columbia; in Current Research 1995-A; Geological Survey of Canada, p. 1-5.

Abstract: Progress on the development of groundwater databases and regional hydrostratigraphic mapping in the lower Fraser Valley are reported. The field program was designed to evaluate how different geophysical methods might be used for groundwater exploration, and to expand the existing well water database with geophysical data.

Résumé : Un rapport d'avancement des travaux entrepris pour la mise sur pied de bases de données sur les eaux souterraines et la cartographie hydrostratigraphique régionale dans la vallée inférieure du fleuve Fraser est présenté. Le programme sur le terrain a été conçu pour évaluer comment différentes méthodes géophysiques peuvent être utilisées pour explorer les eaux souterraines et accroître la base de données actuelle sur les eaux de puits en utilisant les données géophysiques.

INTRODUCTION

Construction of groundwater databases and regional hydrostratigraphic mapping in the lower Fraser Valley (Fig. 1) have been underway for about 18 months. The first phase of the project will be completed in March 1996 (Ricketts and Jackson, 1994). An extensive field program has been conducted since February 1994, with the involvement of several research groups, the co-operation of local government agencies including the City of Surrey, the Corporation of Langley, and the Corporation of Matsqui, and private land owners. The field work is designed to evaluate different geophysical methods with respect to groundwater exploration, and to expand the database and provide new information on hydrostratigraphic architecture.

THE DATABASE

Location, technical, and geological data from approximately 4300 water wells covering an area from White Rock and Surrey to Langley, have been imported into ARC/INFO® (a GIS platform). This represents about 20 per cent of the available water well data in the lower Fraser Valley; the remaining data will be added in the next 18 months. Ultimately, public access to these data and the accompanying hydrogeological interpretations will be made through public terminals (Ricketts and Liebscher, 1994). In the interim, different components of the database are being released as GSC Open Files (e.g., Dunn and Ricketts, 1994).

The database structure in ARC/INFO (also in Paradox/dBASE) has been outlined by Woodsworth and Ricketts (1994). Most of the original data are contained in "verbatim" drilling company reports curated by the Groundwater Section, British Columbia Ministry of Environment, and require extensive "cleaning" before analysis and mapping can proceed. Macros have been written to automate cleaning procedures. Programs have also been developed to graphically display individual wells, and to generate lithostratigraphic/hydrostratigraphic profiles (Ricketts and Dunn, 1995). This is a necessary step to producing subsurface maps, and three-dimensional models of specific stratigraphic units such as aquifers.

FIELD ACTIVITIES

Field activities for the Fraser Valley Hydrogeology project during 1994 employed a variety of geophysical methods, encompassing an area from central Surrey to Sumas Valley (Fig. 2): (1) Borehole drilling, piezometer installation and geophysical logging; (2) Cone penetrometry (CPT); (3) Electromagnetic surveys (EM); (4) Shallow, high resolution reflection seismic surveys; and (5) Ground penetrating radar (GPR). Few, if any of these methods have been used on a regular basis in British Columbia for aquifer and groundwater mapping. Their use in the Fraser Valley project therefore has two principal aims: to evaluate and compare the methods, individually and combined, for groundwater studies in general, and to add geophysical information to the existing water well database. Each method measures different

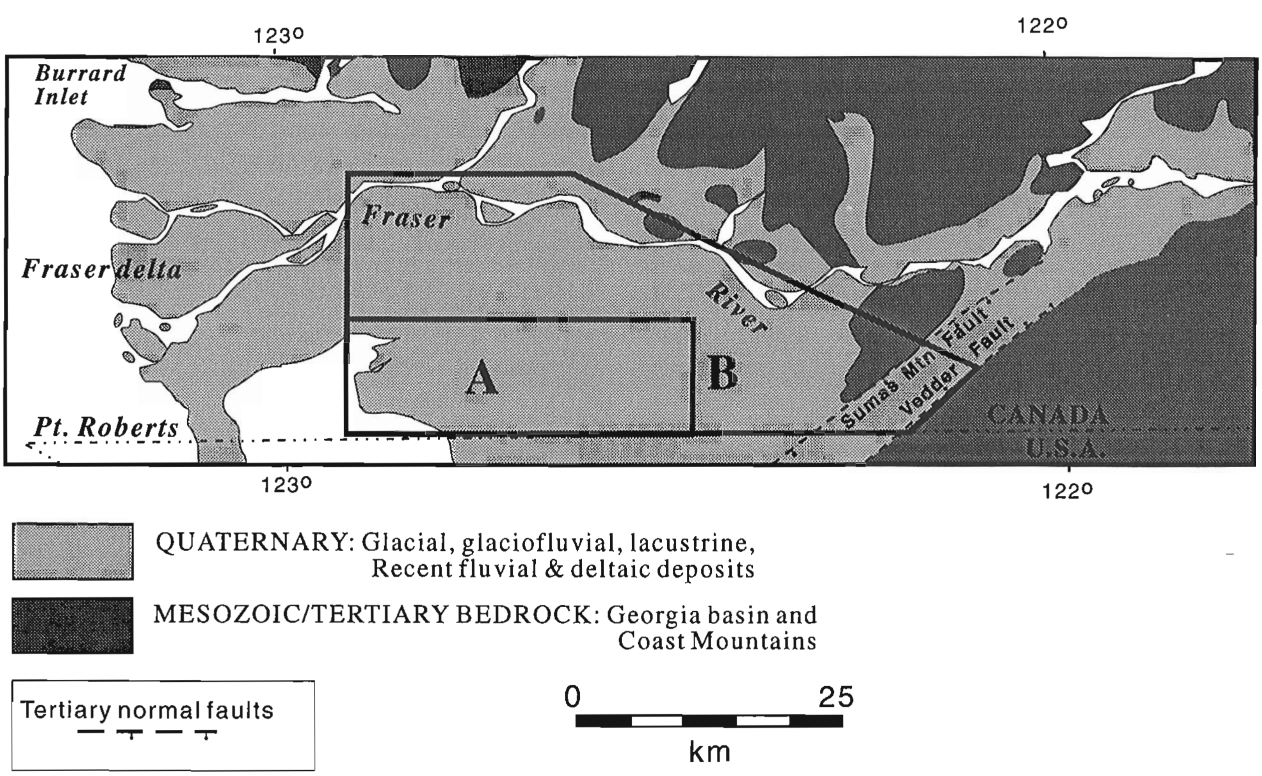

Figure 1. *Location map of Fraser lowlands. Area A indicates the current extent of the digital database. Area B indicates the general coverage of field work and the final extent of the database coverage.*

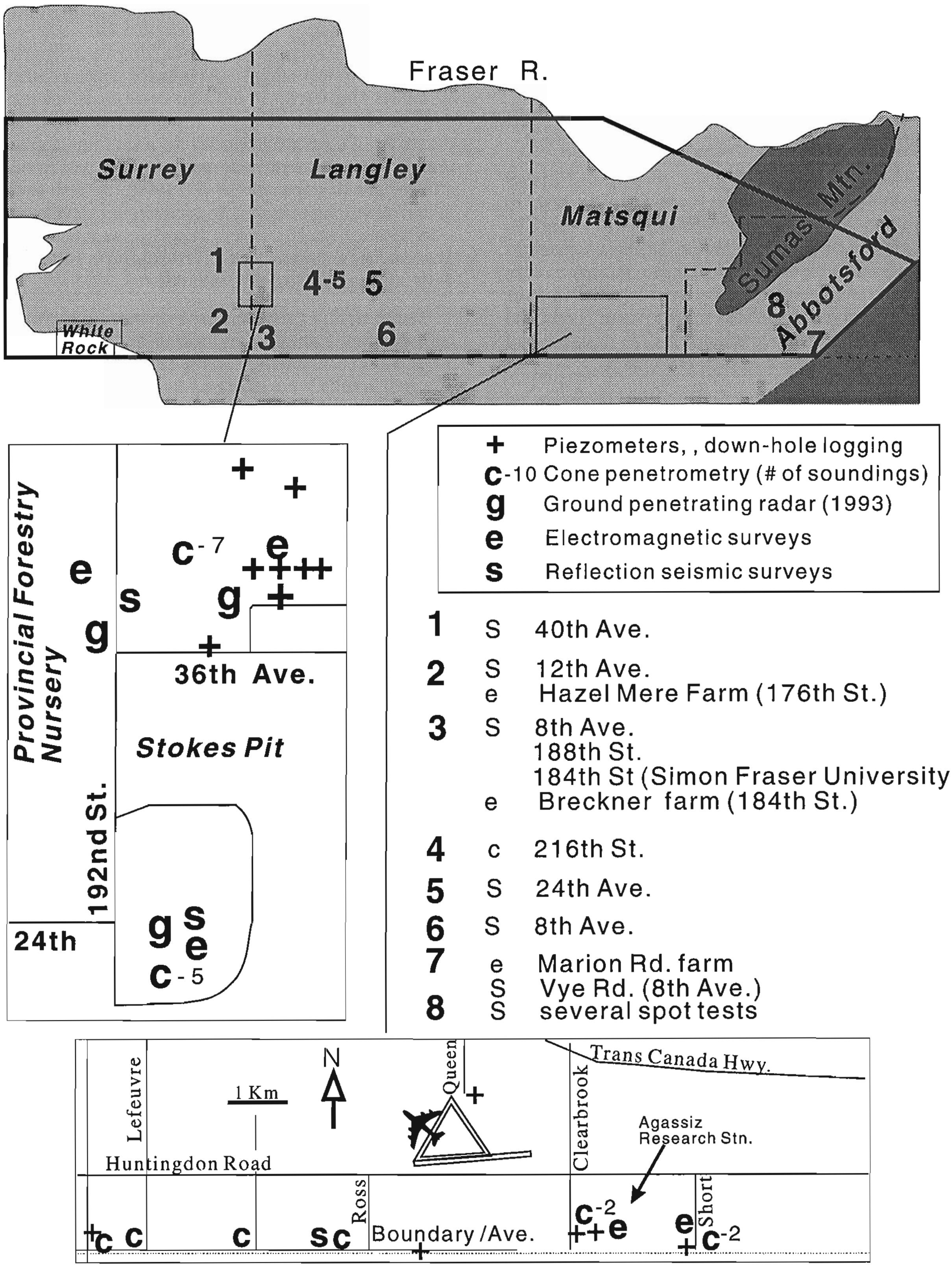

Figure 2. Map showing generalized locations (numbered) of field activities for 1994, with details of the Brookswood and Abbotsford areas.

properties of sediments and fluids and an important direction of future research is the degree to which each data set compliments the others (Fig. 3).

Piezometer installation and borehole logging
(Hugh Liebscher, Environment Canada; Robert McDonald, GSC Sidney)

Thirteen boreholes were drilled, under contract to Novasol Environmental Consultants and Field Drilling Ltd, in Brookswood and Abbotsford unconfined aquifers (Fig. 2). Well depths ranged from 15 m to 30.5 m. Drilling was mostly in sand and gravel of the Sumas Drift, using a cable-tool rig and bailer. Nests of three piezometers were installed in 10 of the boreholes to record fluctuations in static water levels, and to determine local hydraulic gradients. A 0.76 m screen and sand-pack was set at the base of each piezometer.

Three wells, one in the Abbotsford aquifer and two in the Brookswood aquifer, contain a single, fully screened piezometer. These are currently being used by Raghava Dasika, Civil Engineering, University of British Columbia (UBC), to investigate the residence time of nitrate derived primarily from manure storage piles, and its vertical distribution in the groundwater column.

Downhole geophysical measurements were made using conductivity, magnetic susceptibility, and natural gamma probes (MacDonald and Ricketts, in prep.). Preliminary results show clear identification of the water table (conductivity log) and subtle fining and coarsening trends in sand and gravel beds recorded by the gamma logs. The conductivity logs also provided local calibration for the EM surveys (below).

Cone penetrometry
(In-Situ Testing Group lead by Dick Campanella, Civil Engineering, UBC)

Cone penetration testing was conducted, under contract to the In-Situ Testing Group, in and adjacent to the Brookswood and Abbotsford aquifers (Fig. 2). The objective was to establish the utility of cone penetrometry for acquiring detailed information on aquifer hydraulic characteristics and stratigraphic architecture (Campanella et al., 1994). A new system for in-situ measuring of hydraulic conductivity, developed by the UBC Group, was also tested. Test sites were chosen to complement data collected during GPR studies in 1993 (Rea et al., 1994 a,b), and the logged boreholes.

Dense aquifer gravel and sand encountered by the cone in both aquifers restricted the use of the resistivity module; about 300 m of soundings (several locations) were achieved. The best results were obtained from fine grained deposits, mainly sand, silt and clay, but including diamictons. Deep penetration in test sites along the eastern margin of Brookswood aquifer show the method is particularly useful for delineating the pinchout margins of coarse grained aquifers. In addition to recording stratigraphic information and pore pressure, the CPT accurately estimates parameters such as location of the phreatic surface, determination of in-situ gradients, and hydraulic conductivity.

Electromagnetic surveys
(Melvyn Best, GSC Sidney; Brian Todd, GSC Ottawa; Deirdre O'Leary, UBC)

Approximately 6 km of EM survey were completed using 80 m and 20 m loops (Fig. 2). Depths of stratigraphic resolution, commonly 100-200 m, depend on the resistivity of

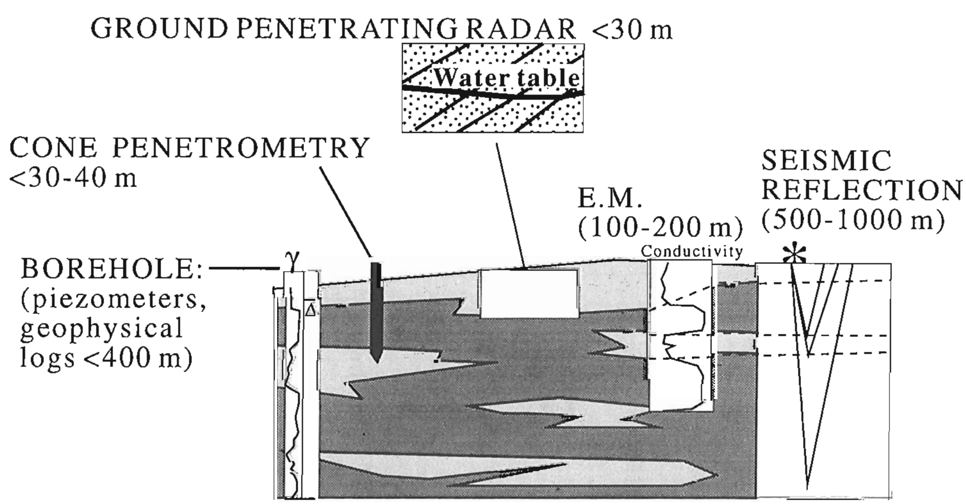

Figure 3. Schematic representation of the different exploration methods showing depths of effective stratigraphic resolution, against a background of aquifers (light grey) and aquitards.

individual layers and resistivity contrast between layers. Sites were selected to provide as much geological variability as possible. Some of these sites coincide with those used for the GPR and seismic surveys so that direct comparison of data sets can be made (Best et al., 1995).

High resolution reflection seismic surveys
(Susan Pullan, Ron Good, GSC Ottawa)

High resolution reflection seismic surveys, totalling 12-14 line kilometres, also were conducted at several locations in the Fraser Valley. Like the EM surveys, locations that provided geological variability were chosen to test the usefulness of the method for determining hydrostratigraphic architecture in the region. In contrast to the EM and GPR surveys, the best results were obtained along transects remote from the unconfined sand-gravel aquifers, with reflections ranging from approximately 40 m to 1 km depth (Pullan et al., 1995). As expected, the shallowest resolvable reflections in the coarse-grained unconfined aquifers coincide approximately with the deepest reflections obtained in the GPR surveys.

Ground penetrating radar
(Rosemary Knight, Jane Rea, Geophysics and Astronomy, UBC)

Ground penetrating radar surveys this year build on the results obtained in 1993 surveys (Rea et al., 1994a,b) to determine the nature of radar reflections with respect to sediment properties such as fabric, texture, and lithofacies, and hydraulic properties like degree of water saturation, porosity and hydraulic conductivity. Because the presence of water strongly effects the radar response in low conductivity materials, detailed measurements of water content will be made at selected gravel pit exposures using Time Domain Reflectometry. The results will permit better definition of the water table and other radar reflectors, and of the heterogeneities in unconfined aquifers.

ACKNOWLEDGMENTS

The viability of the Groundwater Project depends on the participation, assistance, and co-operation of many organizations and individuals: Dirk Tempelman-Kluit, Bert Struik, Bob Turner, Glenn Woodsworth, John Clague, and John Luternauer (GSC Vancouver); Mel Best, Bob MacDonald, and Tark Hamilton (GSC Sidney), Susan Pullan, Jim Hunter, Brian Boyd, Scott Dallimore, and Ron Good (GSC, Terrain Sciences, Ottawa), Gwendy Hall and Judy Vaive (GSC, Mineral Resources Division, Ottawa); Hugh Liebscher and Basil Hii (Environment Canada); Bernie Zebarth (Agriculture and Agrifood Canada); Mayur Kothary (C.R.T.C. Surrey); Paul Matysek and Peter Bobrowsky (B.C.M.E.M.P.R. Geological Survey Branch); Al Kohut, Rodney Zimmerman, and Leigh Ringham (Water Management Branch, B.C. Ministry of Environment, Lands and Parks); Steve Martin and Kerry Schneider (B.C. Ministry of Health); Tony Willingdon (Provincial Forestry Nursery, Surrey); Tom Heath (Greater Vancouver Regional District); Ken Bennett and Wayne Clifton (City of Surrey); Peter Scales, Neil Calver and Rick Walters (Township of Langley); Ed Regts and Peter Andzans (Township of Matsqui); John Psutka (B.C. Hydro); Dick Campanella and Raghava Dasika (Civil Engineering, UBC); Mike Davies (Klohn-Crippen Ltd.); Rosemary Knight and Jane Rea (Astronomy and Geophysics, UBC), Mike Roberts (Geography, SFU); Steelhead Aggregates, Central Aggregates, and Matsqui Township for access to gravel pits; Novasol Environmental Consultants and Field Drilling Ltd; Dick McNicol (Aqua Flow Ltd. and Canadian Groundwater Association); Bill Gray (Hazel Mere Farm, Surrey) and Raymond Breckner (Breckner Farm, Surrey). Jim Monger is thanked for reviewing the manuscript.

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Geological Survey of Canada Project 930035-01

The groundwater database, Fraser Valley, British Columbia

Brian D. Ricketts and David Dunn
Cordilleran Division, Vancouver

Ricketts, B.D. and Dunn, D., 1995: The groundwater database, Fraser Valley, British Columbia; in Current Research 1995-A; Geological Survey of Canada, p. 7-10.

Abstract: A flexible and expandable database has been developed for groundwater studies in the Fraser Lowlands. Currently, the database contains information from about 4300 water wells. The data are being used in a GIS platform to develop 2-dimensional hydrostratigraphic cross-sections and subsurface maps, and 3-dimensional models of specific aquifers.

Résumé : Une base de données souple et extensible a été créée pour les études des eaux souterraines dans les basses terres du Fraser. À l'heure actuelle, la base de données contient des informations concernant environ 4 300 puits d'eau. Les données servent dans un modèle de système d'information géographique à établir des coupes hydrostratigraphiques en deux dimensions, de même que des cartes de la subsurface et des modèles en trois dimensions de formations aquifères spécifiques.

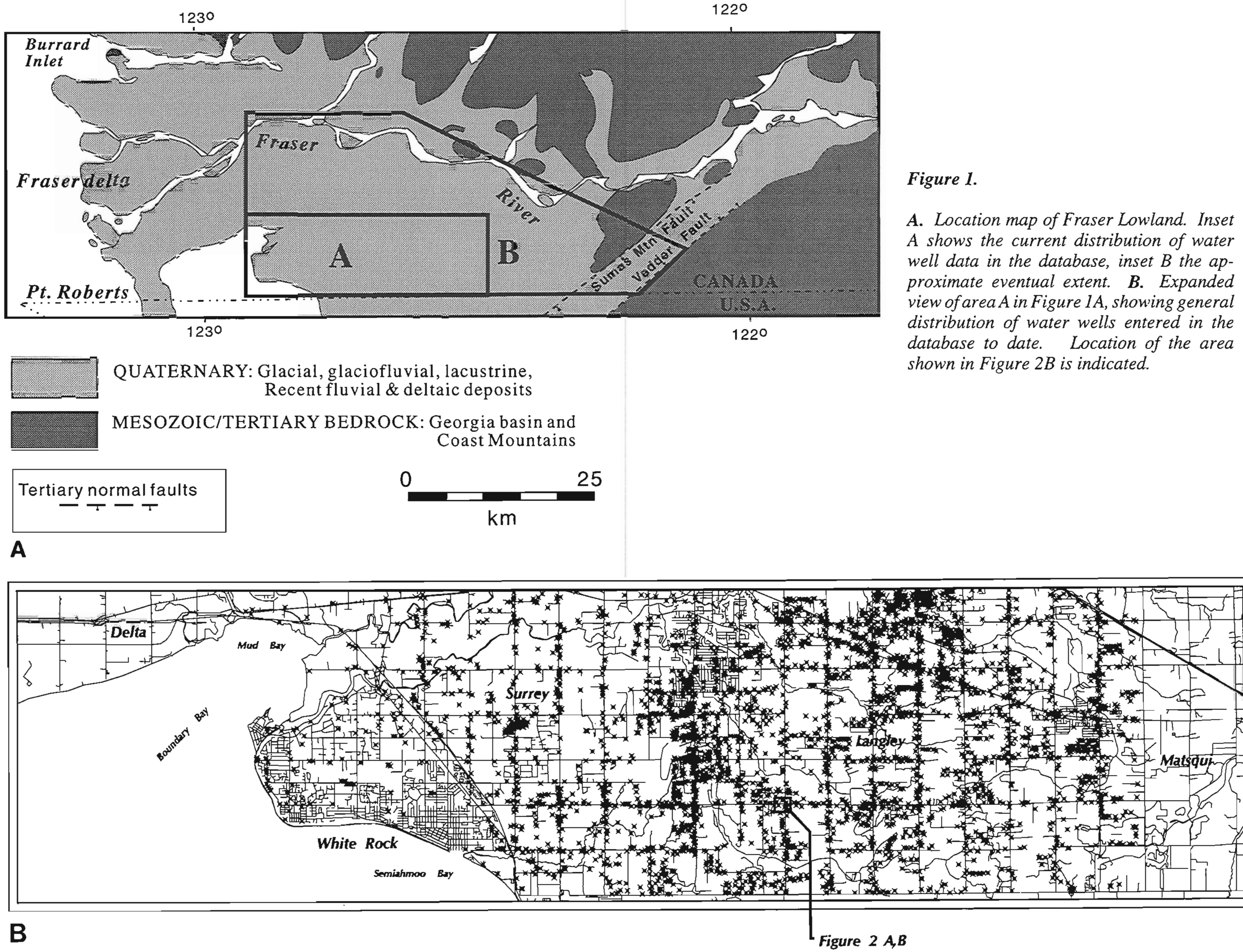


Figure 1.

A. *Location map of Fraser Lowland. Inset A shows the current distribution of water well data in the database, inset B the approximate eventual extent.* ***B.*** *Expanded view of area A in Figure 1A, showing general distribution of water wells entered in the database to date. Location of the area shown in Figure 2B is indicated.*

INTRODUCTION

In 1993, a groundwater database was developed for GIS and spread-sheet platforms to facilitate regional groundwater mapping and research in the Fraser Valley (Woodsworth and Ricketts, 1994). The database is flexible and expandable, keeping in mind three important aspects of the project: (1) easy access to the data for digital mapping and hydrostratigraphic modelling, (2) transferability to other database formats, and (3) eventual public access to the hydrogeological components of the database. This report summarizes the progress made in developing mapping and modelling

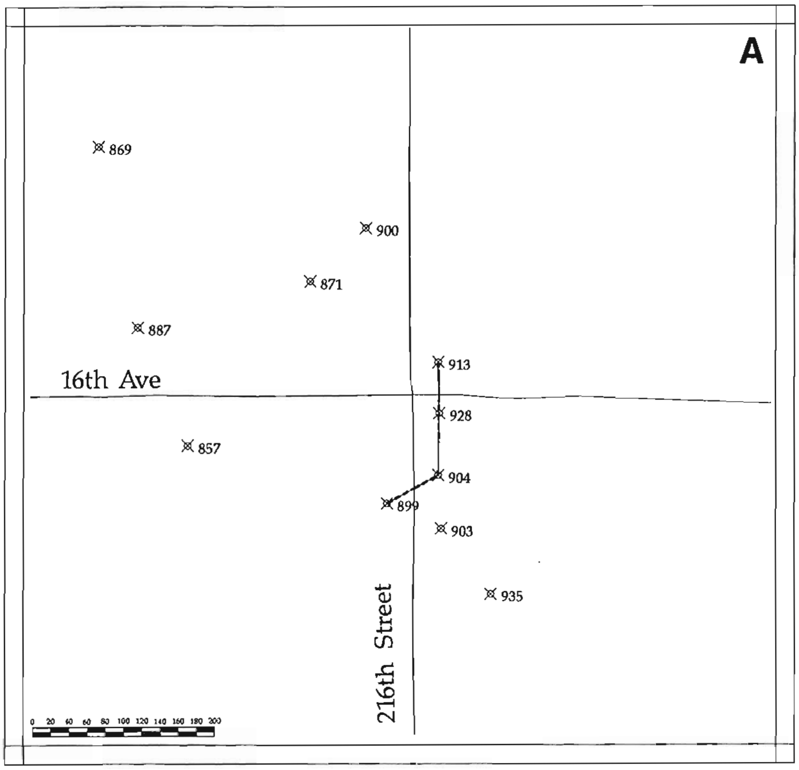
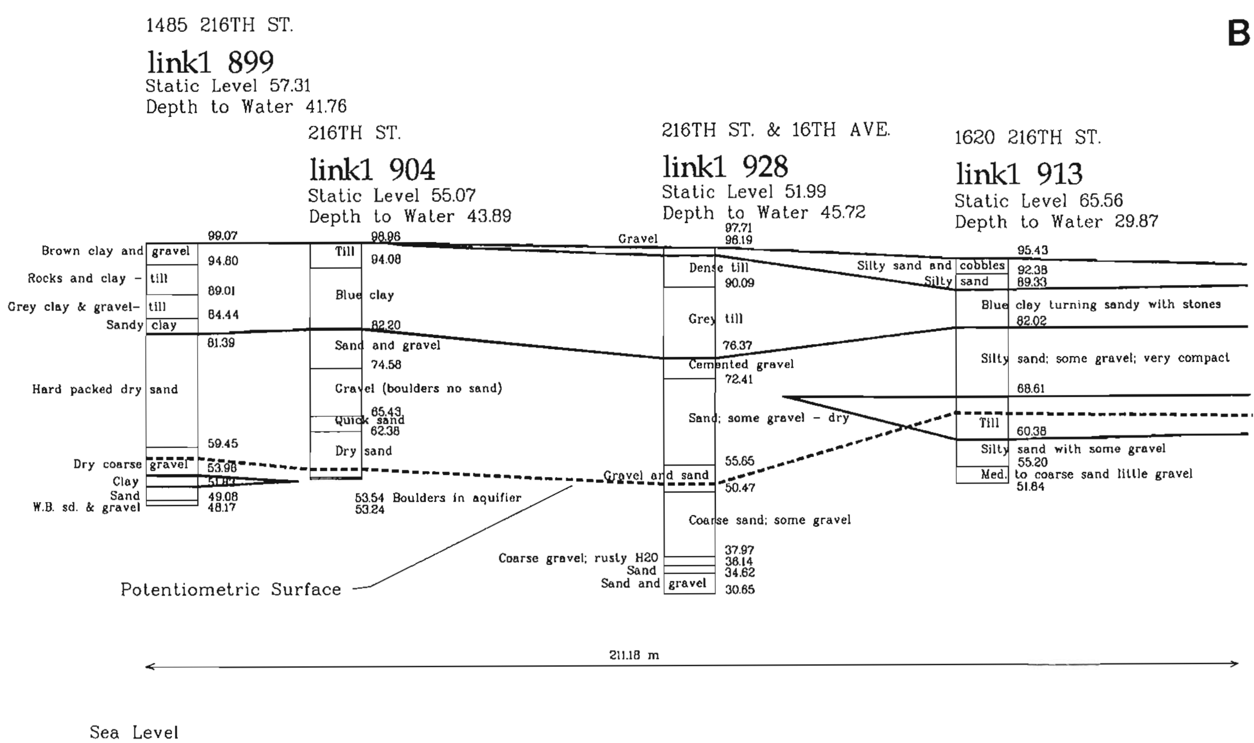


Figure 2.

A. *Detailed map showing water wells along example transect in Figure 2B. Well numbers indicated by LINK1. The location of the map is indicated in Figure 1B. Scale in metres.* ***B.*** *Profile of wells and hydrostratigraphic units along the transect in Figure 2A. Well depths in metres relative to sea level. The top of each well coincides with local elevation of the land surface. Lithological descriptions are the verbatim reports from original well logs. Depth to water is the depth from the well head, recorded by the driller - note that in most cases it is not recorded whether this measurement was taken from the top of a stand-pipe or from ground level; in calculating the static level we assume it refers to ground level. Static level is calculated relative to sea level. Hydrostratigraphic units are delineated by the heavy black lines (represented as arcs in the ARC/INFO^R coverage).*



programs for the GIS (ARC/INFO^R) platform. Data management and analysis are initially being undertaken for part of Surrey and Langley municipalities (Fig. 1A,B). The database will be expanded in the next 18 months to include most of the lower Fraser Valley.

Contents of the database

In designing the database structure, account was taken of published recommendations from the Federal-Provincial Working Group on Groundwater (1991), and existing GSC and United States Geological Survey conventions. The bulk of the data used to date consists of about 4300 water well records provided by the British Columbia Ministry of Environment. Three linked files are extracted from each well record: a well location file, a file containing technical details such as well depth, static level and flow rate, and a drill log file detailing lithological units. All well locations have been digitized to provide UTM coordinates (not present in the original data). UTM coordinates have been converted from the NAD27 to NAD83 datum.

Well-head elevations are not provided in the original water well data. To obtain these values we have interpolated well locations on a digital elevation model (DEM). The DEM was derived from topographic coverages in the B.C. Ministry of Environment TRIM (Terrain Resource Information Management) digital maps (1:20 000 scale), which contain landform, elevation, cultural and cadastral data.

Most of the original water well data consists of "verbatim" copies of drilling company reports. Although these records provide an adequate basis for quick perusal of a small number of wells, they are not suitable for complex analysis and querying. Considerable effort has been expended to standardize and "clean" the data, particularly the drilling records, prior to hydrostratigraphic analysis.

In addition to the water well data, the surficial geology of Fraser Lowlands has been digitized (Dunn and Ricketts, 1994). These maps (NTS areas 92G/1, 92G/2, and parts of 92G/3, G/6, G/7, 92H/4) were originally published by J.E. Armstrong and S.R. Hicock at 1:50 000 scale.

HYDROSTRATIGRAPHIC MODELLING

Lithological descriptions in the well-log file, derived from drilling company reports, form the basis for construction of digital stratigraphic columns, 2-dimensional profiles and subsurface maps, and 3-dimensional models of aquifers and aquitards. Programs have been written (by David Dunn) in the GIS platform to automate selection and graphical representation of water wells, either individually or with many wells along a prescribed transect (Fig. 2A,B). A common example in this project is the selection of wells within a 200 m corridor (or envelope) along streets, rivers, or other linear features. Wells can be displayed with lithological descriptions, depths with respect to a datum (usually sea level), and technical data such as static level. The geographic separation of wells is drawn to scale in the profiles. Well-head elevations are drawn relative to sea level.

Correlation of particular hydrostratigraphic or lithostratigraphic units along transects is made visually; there is too much variation in the descriptions of lithological units among different well reports to warrant automation of this step. When correlations have been established, stratigraphic profiles are built (Fig. 2B). In this project, we have used as our starting point the earlier, seminal studies of hydrostratigraphic architecture by E.C. Halstead and H. Liebscher (Halstead, 1986). The profiles will form the basis for constructing subsurface maps and 3-dimensional models.

ACKNOWLEDGMENTS

Water well data are curated by the B.C. Ministry of Environment. Rodney Zimmerman is thanked for providing these data. Glenn Woodsworth and Bertrand Groulx have continued to provide assistance as the database has evolved. Rob Cocking and Stephen Williams shared their GIS expertise as we developed this critical component of the hydrogeology project. Glenn Woodsworth is thanked for critically reviewing the paper.

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Geological Survey of Canada Project 930035-01

Preliminary results from a shallow seismic reflection survey, Lower Fraser Valley hydrogeology project, British Columbia

Susan E. Pullan, Ronald L. Good, and Brian D. Ricketts[1]
Terrain Sciences Division

Pullan, S.E., Good, R.L., and Ricketts, B.D., 1995: Preliminary results from a shallow seismic reflection survey, Lower Fraser Valley hydrogeology project, British Columbia; in Current Research 1995-A; Geological Survey of Canada, p. 11-18.

Abstract: The results of the testing phase of a shallow seismic reflection survey carried out in 1994 as a component of the Lower Fraser Valley hydrogeology project are summarized in this paper. The quality of the seismic reflection data obtained varied widely across the survey area, depending primarily upon surface geological conditions. In some areas where the surface materials were fine grained and water saturated, excellent data were obtained and continuous seismic profiles from these areas are expected to yield subsurface structural information to depths of up to 800 m below surface. Other areas proved to be unsuitable for the application of shallow seismic reflection techniques.

Résumé : Les résultats de la phase d'essai d'un levé de sismique- réflexion à faible profondeur réalisé en 1994 dans le cadre du Projet d'hydrogéologie de la vallée inférieure du Fraser sont présentés. La qualité des données de sismique-réflexion a varié beaucoup dans la région étudiée, surtout en fonction des conditions de la géologie de surface. Dans certaines régions où les matériaux de surface étaient à grain fin et saturés en eau, on a obtenu d'excellentes données, et les profils sismiques continus de ces régions devraient fournir des informations sur la structure souterraine jusqu'à des profondeurs de 800 m. D'autres régions ne se sont pas avérées propices à l'application des techniques de sismique-réflexion à faible profondeur.

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INTRODUCTION

The Quaternary materials of the lower Fraser Valley, from Boundary Bay south of Vancouver, British Columbia, to the Sumas Valley east of Abbotsford, contain many important aquifers, some unconfined (such as the Abbotsford and Brookswood aquifers), and others confined by diamictons, glaciomarine and glaciolacustrine deposits. The Lower Fraser Valley hydrogeology project was initiated in April 1993, with the aim of characterizing the aquifers and aquitards in terms of their 3-dimensional geometry, stratigraphic position and sedimentary character, hydraulic properties, recharge characteristics, and groundwater types (Ricketts and Jackson, 1994). In order to obtain some of the baseline data needed to achieve this aim, many different types of data are being utilized in the groundwater mapping process, including water well records (Woodsworth and Ricketts, 1994; Ricketts, 1995), new drilling and water sampling, cone penetrometry, and geophysical surveys (Rea et al., 1994; Best et al., 1995).

Geophysical methods provide a non-destructive means of obtaining data related to the physical properties of the subsurface, and a way of expanding the one-dimensional information obtained by boreholes into a two- or three-dimensional picture. This paper reports on the preliminary results of a shallow seismic reflection survey carried out in the lower Fraser Valley in July and August of 1994.

Shallow seismic reflection techniques depend on contrasts in the density and seismic velocity of sediments, and so provide a potential means of mapping the subsurface structure of unconsolidated sequences. Such stratigraphic information can provide critical insight into past depositional environments, and the lateral continuity (or inhomogeneity) of subsurface units, and in this sense, shallow seismic reflection profiles can play an important role in hydrogeological studies (Pullan et al., 1994).

METHOD

Shallow seismic reflection techniques involve laying out a series of receivers (geophones) on the ground surface, and recording ground motions at these locations as a function of time in response to the detonation of a small explosive/shotgun source or the impact of a weight drop or sledge hammer. Energy travels along the ground surface, is refracted along interfaces such as the water table where large increases in seismic velocity occur, and is reflected from subsurface horizons such as lithological boundaries where there is a contrast in acoustic impedance (product of density and seismic velocity).

Because of the interference of direct, refracted and reflected events in the near surface, seismic reflection profiles cannot yield information on very shallow structure (i.e., from

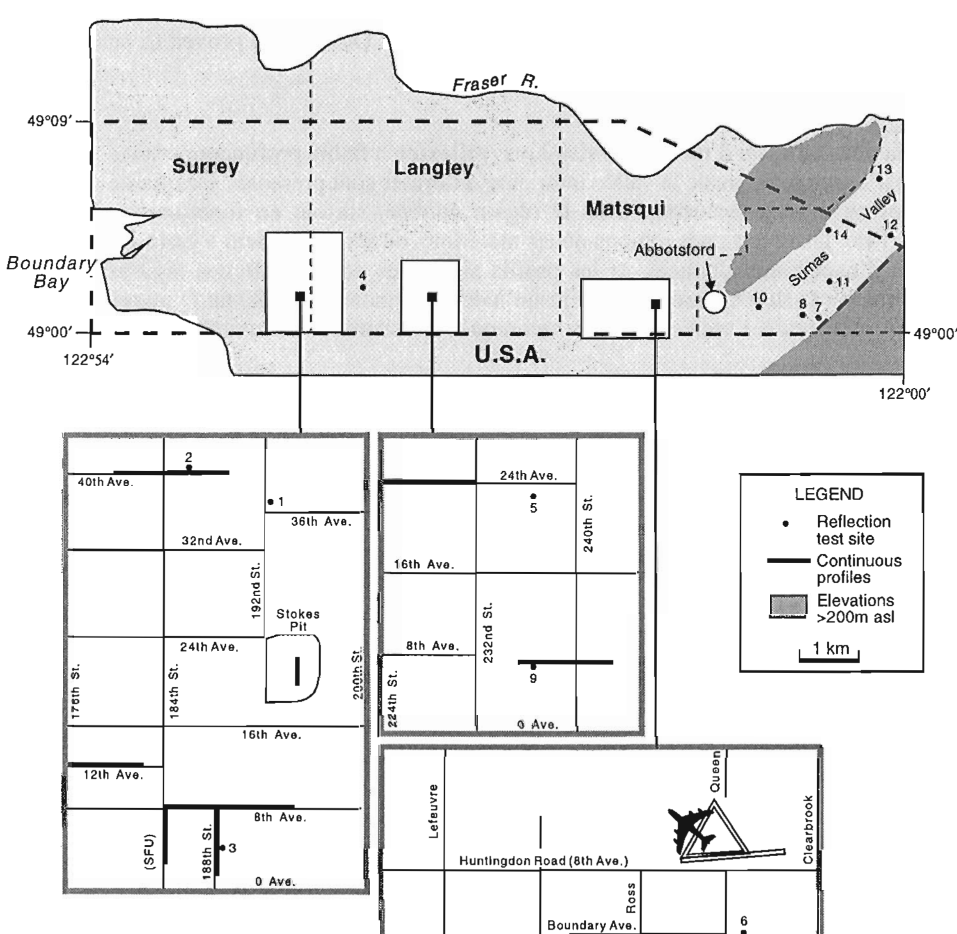

Figure 1.

Location map showing the 14 test sites and the 13 line-kilometres of seismic profiles collected during the survey.

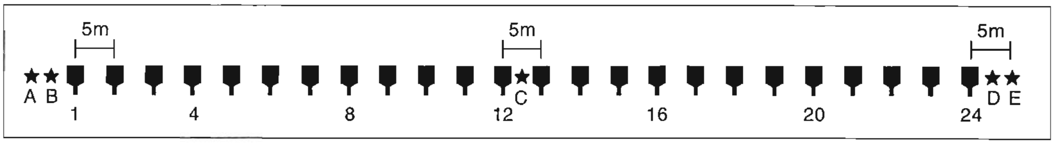

Figure 2. Source-receiver layout for test spreads. The 24 geophones are uniformly spaced 5 m apart, and data are recorded for shots 5 and 2.5 m off each end (shotpoints A, B, D and E), and in the centre of the spread (shotpoint C).

depths of 0 to 15-30 m below surface, depending on the geometry of the source/receivers and on the dominant frequency of energy transmitted by the ground). Where information on the sedimentary structures of the very near-surface is required, ground penetrating radar techniques should be considered. Seismic sections constructed from data such as recorded for this project, can yield subsurface structural information in the depth range of 20 m to several hundred metres below surface.

The Fraser Valley shallow seismic reflection survey involved both a testing and a production phase. Figure 1 shows the locations of the test sites and the profiles that were obtained during this survey.

During the testing phase, one 24-channel geophone spread was laid out at each site, and data were recorded to evaluate the quality (frequency and signal strength) of reflected energy, to determine the depths from which reflection signals were recorded, and to establish the recording parameters (source/receiver geometry, recording timescale, input filter settings etc.) and line priorities for the production phase. Tests were carried out at 14 sites across the survey area in four general locations (see Fig. 1). An effort was made to test sites located on all major surficial geological units (Armstrong, 1980, 1984; Armstrong and Hicock, 1980), as surface conditions strongly affect the frequency and signal strength of the energy that can be transmitted into the ground.

The production phase involved the recording of continuous 12-fold CDP (common depth point) data along the chosen survey lines (where "fold" refers to the number of traces stacked or summed to produce a single trace on the final section; see Steeples and Miller, 1990). This was accomplished by "rolling" 24 channels through 36 geophones laid out on the ground surface as the shot point was moved every 5 m. Approximately 12 line-kilometres of 12-fold CDP data were recorded in this manner in 21 working days with a 4 person crew; an additional 1 line-kilometre was recorded by a crew from Simon Fraser University. Two days were required for surveying relative elevations along the lines so that elevation corrections can be made on the final processed sections.

All data were recorded on a Geometrics ES-2401 24-channel engineering seismograph with an instantaneous floating point amplifier and 15-bit analog-digital conversion. Both the tests and CDP data were recorded with single 50 Hz vertical geophones per channel, 5 m geophone spacing, and a 12-gauge black powder blank shotgun shell detonated at the bottom of a 0.5-1 m hole as the seismic source (Pullan and MacAulay, 1987).

RESULTS

Processing of the CDP sections is currently underway. This paper presents some results from the testing phase of this survey. The tests involved recording data from 5 shots into a stationary 24-channel spread laid out in the ditch alongside the road (see Fig. 2 for source-receiver layout). The data from such a series of shots (see Figures 3 and 5) can be combined to generate a 96-trace section (1-2 fold), 120 m in length (see Fig. 4, 6).

The test results can be summarized on the basis of data quality and surficial geological units as follows:

1. sites that were shot on the relatively fine grained Capilano Formation sediments (glaciomarine, deltaic, and fluvial deposits) and recent floodplain deposits produced excellent reflection records, with dominant frequencies in the 150 Hz range and visible reflections on the field records to 800 milliseconds (corresponding to an estimated depth of 700 m below ground surface).
2. sites that were shot on the coarser grained Fort Langley formation (glacial and deltaic sediments) and Sumas drift (outwash, ice-contact and deltaic deposits), where the surface sediments were typically drier and water table was several metres below ground surface, produced fair to good reflection records, with lower dominant frequencies (~100 Hz), considerably more interference from ground roll (low frequency, large amplitude surface waves), and visible reflections on the field records down to 200-400 ms (estimated depths of 180-350 m below surface).
3. sites that were shot in the Sumas Valley (fine grained lacustrine deposits), all of which yielded extremely poor reflection records.

1. Example of excellent test site results

Figure 3 shows the field records obtained at site 3 (Fig 1; 188th Street approximately midway between 0 and 8th Avenues) where the surface materials are mapped as Capilano Formation sediments (marine silt-loam to clay-loam), and at this particular site consisted of an extremely compact damp silt/clay. These are exceptional shallow reflection records, with reflections in the frequency range of 150-300 Hz,

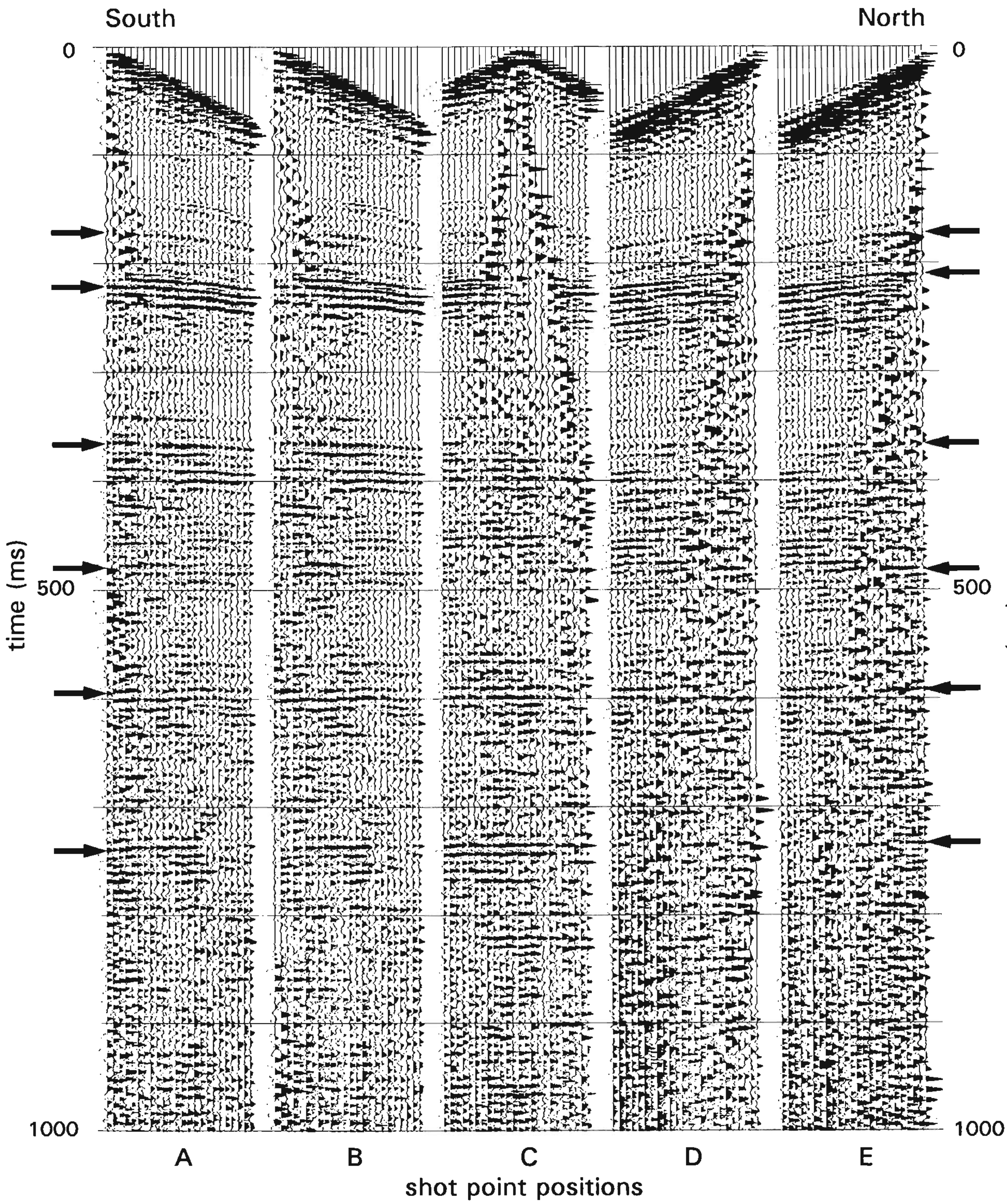

Figure 3. *Example of the 5 test shots recorded at site 3 where excellent high resolution seismic reflection data were obtained. These 1 second records were recorded with a 100 Hz analog low cut filter and a 60 Hz notch filter applied, and are displayed here with an AGC (automatic gain control) with a window of 250 ms applied to normalize amplitudes along the trace. Some major reflection events are indicated by arrows; the slower event (particularly evident in the centre shot record) is the ground-coupled air wave.*

virtually no interference from surface waves or a ground-coupled airwave, and clearly visible reflection events down to 800-1000 milliseconds (ms).

The records shown in Figure 3 have been sorted according to their common depth points (midpoint between source and receiver), a velocity analysis has been performed and the corrected traces have been stacked to produce the 96-trace section (1-2 fold) plotted in Figure 4. The section shows a distinctive, high-amplitude reflection at approximately 200 ms (~180 m depth), as well as numerous other events from 160 ms to 1000 ms. All horizons appear to be horizontal on this scale (120 m). An estimated depth scale is shown on the right hand side of the section; these depth estimates are not well-constrained in the lower portion of the section where the spread length is small in comparison to the depths and there is little moveout on the reflected events on the seismograms.

It was concluded that a shallow seismic reflection profile in this area, conducted with the equipment and recording parameters used in this test, would provide excellent high resolution subsurface structural information to depths of up to 800 m below ground surface (Quaternary and Tertiary sequences).

2. Example of good test site results

Figure 5 shows the field records obtained at site 5 (Fig. 1; 236th Street just south of 24th Avenue) where the surface sediments are mapped as Fort Langley Formation (glaciomarine stony clayey silt to silty sand). The surface material at this site appeared to be a dry sandy diamicton. Reflections are visible on these records down to times of 320 ms (approximately 250 m depth), but the reflection events are less coherent and generally lower in amplitude than on the records shown in Figure 3, and higher-amplitude surface waves interfere with deeper reflections, particularly at small source-receiver separations.

A section produced from these records is shown in Figure 6. The blank areas angling across the section from each side are caused by a mute of the airwave and groundroll. The most prominent events on this profile are a high- amplitude reflection at 80 ms (~70 m depth), and a deeper reflection at 300 ms (~260 m depth), both apparently dipping slightly to the north. Some weaker reflections are also visible (e.g., at 200 ms and 270 ms), though these are not generally coherent across the entire section with this low fold data. With the improvement in the signal to noise ratio expected with 12-fold data, even such weak events should be well- defined on the final seismic profiles.

It was concluded that a shallow seismic reflection profile in this area would provide good subsurface structural information to depths in the order of 250-300 m below surface.

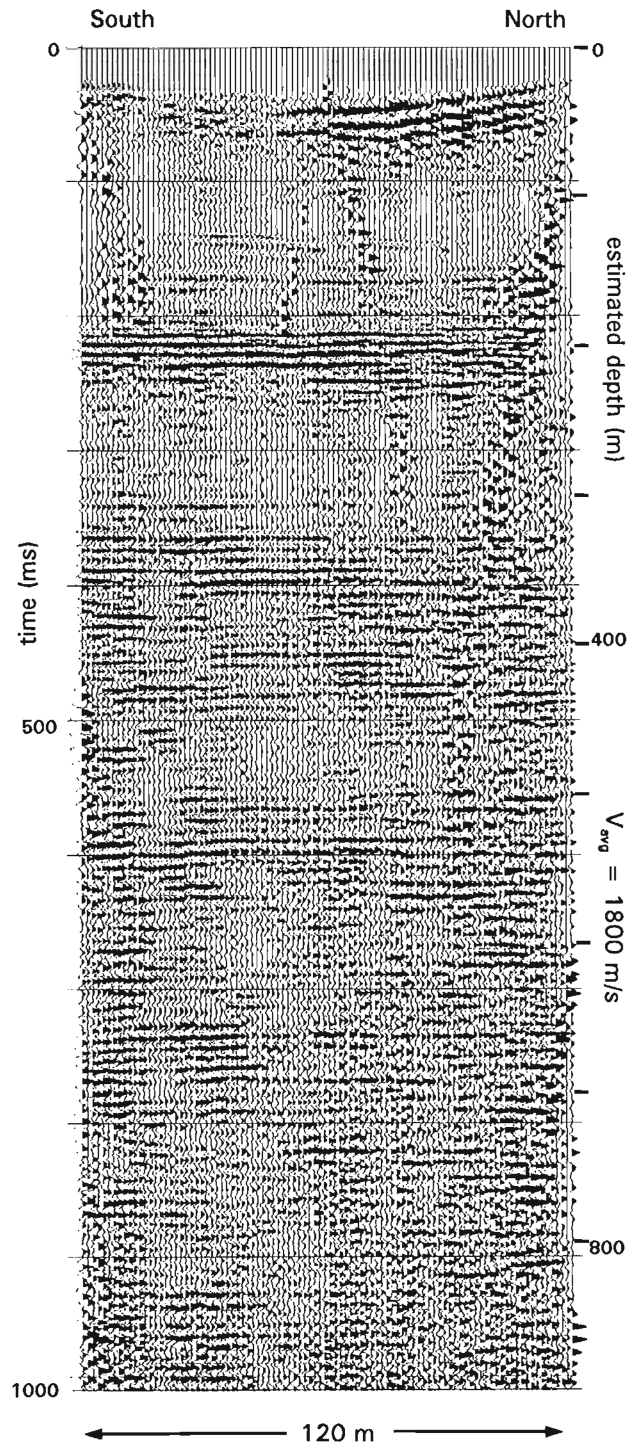

Figure 4. 96-trace section (1-2 fold) constructed from records shown in Figure 3. The depth scale has been calculated from the results of the velocity analysis conducted to create this stacked section.

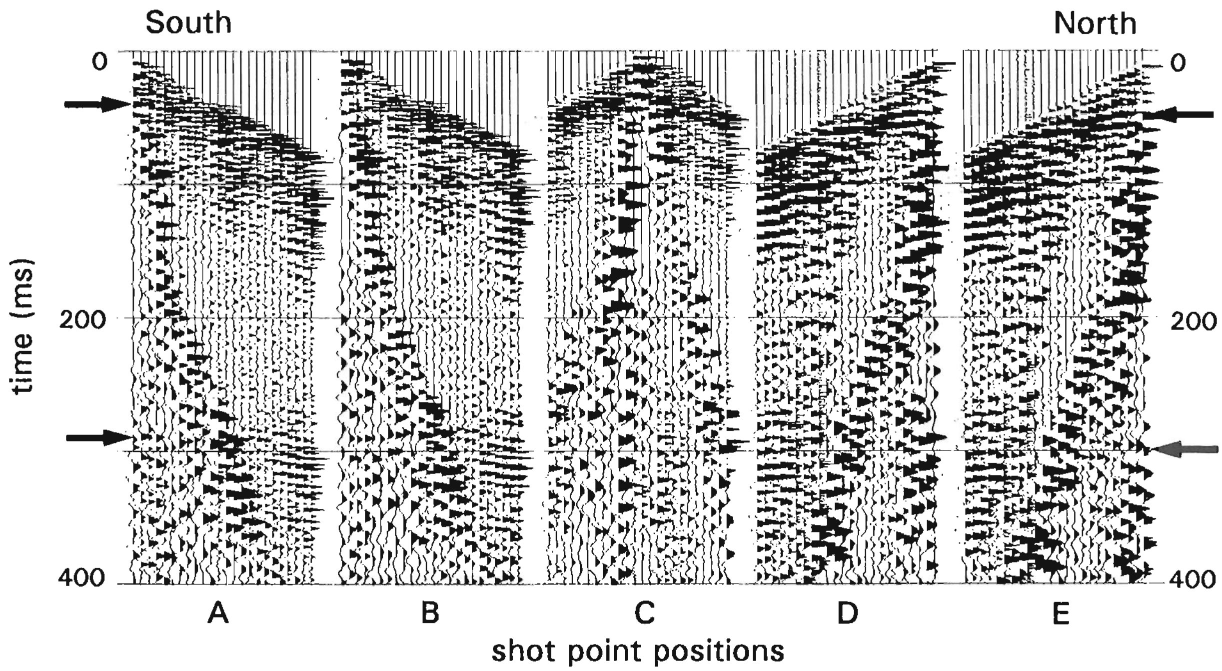

Figure 5. Example of the 5 test shots recorded at site 5 where good seismic reflection data were obtained. These 400 ms records were recorded with a 100 Hz analog low cut filter and a 60 Hz notch filter applied, and are displayed here with an AGC (250 ms window) applied to normalize amplitudes along the trace. Some reflection events are visible on these records, but the signal strength is lower and there is considerably more interference from surface waves than on the records shown in Figure 3.

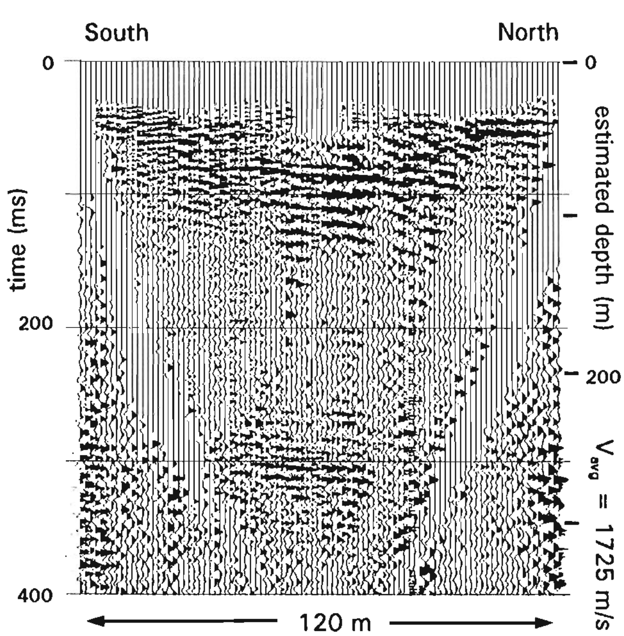

***Figure 6.** 96-trace section (1-2 fold) constructed from records shown in Figure 5. The depth scale has been calculated from the results of the velocity analysis conducted to create this stacked section.*

3. Example of poor test site results - Sumas Valley

Figure 7 shows the field records obtained at site 10 (Fig. 1; south-central region of the Sumas Valley) where the surface sediments are mapped as post-glacial lacustrine deposits (silt to clay). This is a typical suite of test records from the Sumas Valley, where, despite the fine grained surface sediments, very poor reflection records were obtained. The signal strength and frequency of the energy that are transmitted into the ground were extremely low. Such records suggest that there is a highly attenuating medium in the near-surface, such as peat or disturbed material (fill), but the actual cause of the poor transmission of high frequency seismic energy into the ground in this region is unknown at the present time. It was concluded that attempting to obtain a reflection profile in the Sumas Valley during this survey was not worthwhile.

CONCLUSIONS

This paper has examined some of the results of the testing phase of a shallow seismic survey conducted in support of the Lower Fraser Valley hydrogeology project. The tests have shown that the quality of shallow seismic reflection records varies widely across the survey area, depending primarily on the surface geological conditions. Excellent data were

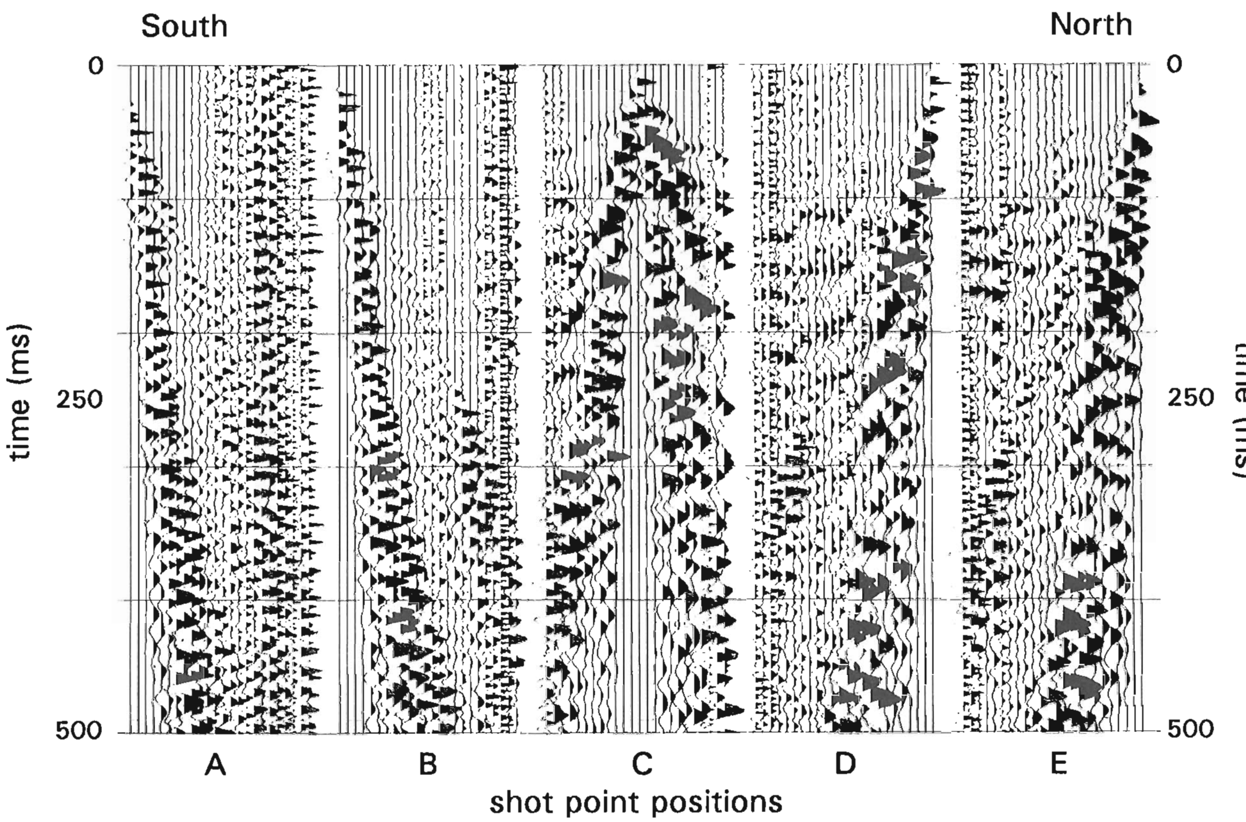

Figure 7. Example of the 5 test shots recorded at site 10 in the Sumas Valley where very poor seismic reflection data were obtained. These 1 second records (only 500 ms are plotted here) were recorded with a 100 Hz analog low cut filter and a 60 Hz notch filter applied, and are displayed here without any AGC applied. These records show very low signal strength, much interference from surface waves and other ambient noise, and no visible reflection events

obtained in areas where the surface materials were fine grained and water saturated (Capilano Formation sediments and recent floodplain deposits), and in these areas shallow seismic reflection profiles may provide detailed subsurface structural information to depths of up to 800 m below surface. As expected, the data quality was not as high in areas where the surface sediments were drier and coarser (Fort Langley Formation and Sumas Drift). However, good high-resolution reflection data were obtained at test sites in these areas, and it is expected that shallow seismic reflection profiles collected here will provide subsurface structural information to depths of up to 300 m below surface. Unfortunately, data acquired in the Sumas Valley were extremely poor, in spite of seemingly excellent surface conditions (damp, fine grained sediments). The reasons for the poor data quality are not understood at this time. Because of the extremely poor test results, no seismic profiles were acquired in this area.

ACKNOWLEDGMENTS

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Geological Survey of Canada projects 920039 and 930035

Groundwater mapping using time-domain electromagnetics: examples from the Fraser Valley, British Columbia

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Best, M.E., Todd, B., and O'Leary, D., 1995: Groundwater mapping using time-domain electromagnetics: examples from the Fraser Valley, British Columbia; in Current Research 1995-A; Geological Survey of Canada, p. 19-27.

Abstract: Time domain electromagnetic (TDEM) surveys were carried out during the summer of 1994 as part of the Geological Survey of Canada's Fraser Valley hydrogeology project. The purpose was to examine the capability of TDEM systems to map surficial geology, for example ground water aquifers and impermeable zones such as clays, and water quality (salinity). Approximately 100 electromagnetic (EM) soundings were collected at 9 sites within the lower Fraser Valley and Fraser Delta using a Geonics EM-47 system operating in the central sounding mode. The sites were chosen to cover a range of geological conditions for evaluating the effectiveness of the EM system. Preliminary interpretation of the data indicates that electromagnetic methods can map surficial geological features for ground water to depths between 150 m and 200 m. Three soundings are included to illustrate typical EM responses obtained during the survey. Interpretation of the entire 100 soundings is ongoing and will be presented in the future.

Résumé : Des levés électromagnétiques dans le domaine temporel ont été menés au cours de l'été 1994 dans le cadre du projet d'hydrogéologie de la vallée du Fraser de la Commission géologique du Canada. L'objectif était d'étudier la capacité des systèmes électromagnétiques dans le domaine temporel à cartographier la géologie de surface, par exemple les formations aquifères et les zones imperméables telles les argiles, de même que la qualité de l'eau (la salinité). Les résultats d'environ 100 sondages électromagnétiques ont été recueillis sur neuf emplacements dans la vallée et le delta du Fraser au moyen du système Geonics EM-47 fonctionnant sur le mode de sondage central. Les emplacements ont été choisis de manière à couvrir une vaste gamme de conditions géologiques dans l'évaluation de l'efficacité du système électromagnétique. Une interprétation préliminaire des données montre que les méthodes électromagnétiques peuvent cartographier les éléments géologiques superficiels dans le cas d'eaux souterraines se trouvant à des profondeurs variant entre 150 m et 200 m. Trois sondages sont présentés pour illustrer les échos électromagnétiques typiques obtenus au cours du levé. L'interprétation des 100 sondages se poursuit et fera l'objet d'une présentation à venir.

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INTRODUCTION

Land use in the lower Fraser Valley has been steadily increasing during the last two decades. The valley has one of the fastest-growing populations in Canada and is one of the most important agricultural areas in British Columbia. Increasing land use during this period has put significant pressure on ground water aquifers in the lower Fraser Valley and the Fraser Delta. Consequently, new aquifers must be delineated and old aquifers must be better defined to develop their total potential. Many of the shallower, known aquifers are presently being utilized. New aquifers will therefore be deeper and more difficult to locate. Ground water contamination, particularly in unconfined aquifers, has become a contentious issue in the Fraser Valley. More information on flow direction of aquifers is required, as well as the location of impermeable boundaries and other barriers to flow, in order to understand contaminant and recharge problems. Since the regional geology is an essential component of these studies the Geological Survey of Canada has initiated a geological mapping program in the Fraser Valley.

As part of the Geological Survey of Canada's Fraser Valley hydrogeology program, time-domain electromagnetic (TDEM) surveys were carried out at 9 sites during the summer of 1994. The objective of these surveys was to investigate if electromagnetic methods can be used to map aquifers, to locate seals or aquitards, and to map the quality (salinity) of water. The sites were chosen to provide a range of geological environments and to ensure overlap between electromagnetic, reflection seismic, and ground penetrating radar surveys.

This project investigated whether TDEM methods could differentiate good aquifers (sands and gravels with high porosity and permeability) from poor aquifers (fine grained sands, silty sands with low porosity and permeability) and from seals and impermeable boundaries (clays) within the Fraser Valley. The project also examined whether electromagnetic methods could map the bedrock topography beneath sands and gravels that may provide preferential flow paths. Finally, the project studied whether EM methods could distinguish fresh water from brackish or saline water at depth.

ELECTROMAGNETIC METHODS

Electromagnetic (EM) methods map conductivity variations within the earth by inducing currents in the ground using a time-varying electromagnetic field. All EM systems consist of a transmitter that generates an electromagnetic (primary) field and one or more receivers that measure both the (secondary) fields generated in the earth by the induced eddy currents and the primary field. The two types of EM system in use at the present time are frequency-domain EM and time-domain EM. The transmitter of a frequency-domain EM system operates at a number of fixed frequencies, usually between 100 Hz and 20 kHz. The receiver(s) measure the amplitude and phase (relative to the transmitted magnetic field) or the in-phase and quadrature components of the secondary magnetic field at these fixed frequencies. The transmitter of a time-domain EM system transmits square waves or coded pulses. The receiver(s) measure the magnetic field as a function of time, usually when the current (primary field) is off.

Electromagnetic transmitters consist of a coil of area A with N turns of wire, each carrying a current I. The strength of the transmitted electromagnetic field is proportional to the dipole moment of the transmitter coil M = NIA. Electromagnetic receivers consist of coils with N turns of wire. They measure the voltage in the coil caused by the time-dependent magnetic field (Faraday's Law). A direct link is often made between the transmitter and the receiver(s) to ensure that time zero (time domain systems) and phase (frequency domain systems) are known. Sometimes this link is provided by calibrated crystal clocks.

The electromagnetic survey carried out this year as part of the Fraser Valley hydrogeology project used a time-domain electromagnetic system, the Geonics EM-47 (Geonics, 1991). The system was leased from the Ontario Geological Survey for this project.

Figure 1 is a schematic diagram depicting the layout of the transmitter and receiver for operating in the central sounding mode, the configuration used during the entire survey. The transmitter dipole consists of a single wire deployed in a square loop. The survey used a loop with sides equal to 20, 40 or 80 m with the wire carrying a current between 2.1 and 2.3 A. The transmitter controls the shape of the current and the frequency of the transmitted square wave. The receiver coil is a multi-turn coil with a diameter of approximately 1 m (coil area times the number of turns equal to 31.4 m^2). A reference cable is connected between the transmitter and receiver to control the timing of the transmitted waveform and the time during which the received voltage is measured. The receiver measures the vertical component of the magnetic field.

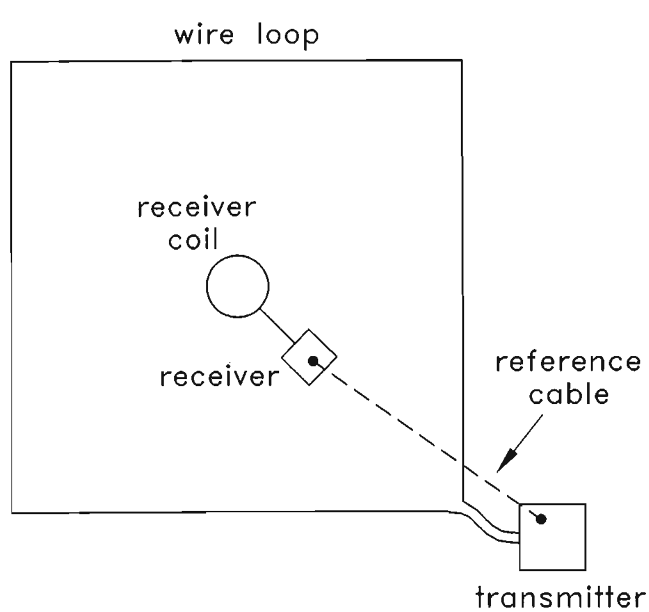


Figure 1. *Schematic of the Geonics EM-47 system in the central loop sounding mode.*

Figure 2a illustrates the time dependent current in the transmitter loop, the voltage in the receiver coil caused by this primary current (V_P) and the secondary voltage in the receiver coil caused by the induced currents in the earth (V_S). The transmitter current approximates a square wave with a sine-wave rise time and a linear ramp at the end of the square wave to turn off the current. The frequency of the square wave is equal to 1/T where T is the period of the square wave pulse. The period consists of a positive square wave of duration T/4, followed by an off-time of duration T/4. These are then reversed to give a total period equal to T. Three frequencies are available with the EM-47 system (UH = Ultra High frequency = 285 Hz, VH = Very High frequency = 75 Hz, and H = High frequency = 30 Hz).

As already discussed, the receiver is on when the transmitter is off. Consequently there is no primary voltage at the receiver that can interfere with the secondary magnetic field generated from the induced eddy currents. The receiver voltage in the EM-47 unit is measured in millivolts but is

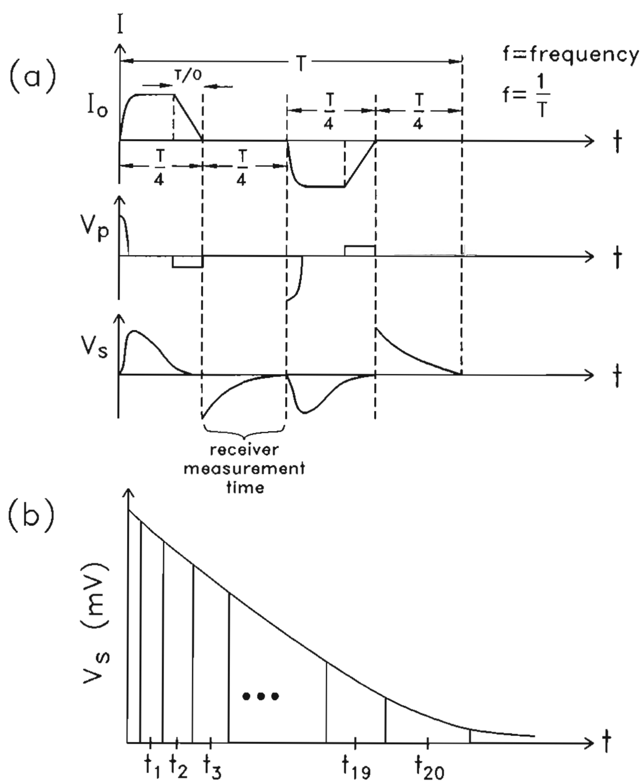


Figure 2. ***a)*** *The transmitter current wave form and the corresponding voltage in the receiver. The frequency of the transmitter current is defined to be 1/T where T is the period. The current is almost a square wave with a sine rise-time and a linear ramp turn-off time (T/O) which depends on the size of the transmitter loop. The secondary voltage caused by the induced eddy currents V_S are measured by the receiver during the off times of the current.* ***b)*** *A blow up of the secondary voltage at the receiver showing the time windows increasing with time. The time of a window is defined to be the time at the centre of the window.*

displayed as nanovolts (nV) per Am^2 (dipole moment) when plotted as a function of time using the TEMIXGL software from Interpex Limited, Golden, Colorado (TEMIXGL, 1994). The receiver voltage is measured in 20 time windows for each of the 3 square wave frequencies listed above. The times go from about 6 µs to 7 ms, with UH times going from 6.85 µs to 701 µs, VH times going from 48.3 µs to 2.825 ms and H times going from 100 µs to 7.04 ms. The receiver voltage is sampled logarithmically in time as shown in Figure 2b. Each voltage measurement is averaged over its time window and the time at the centre of the averaging window is defined to be the time for that window. The windows increase at longer times because the voltages are usually smaller.

The voltages are converted to apparent resistivity values using late time normalized voltages (Fitterman and Stewart, 1986; Stoyer, 1990). The apparent resistivity is defined to be the ratio of the measured voltage to the voltage that would be measured over a half-space of constant resistivity. Once the apparent resistivity versus time curves are computed, the data can be interpreted in terms of multi-layered earth models using standard forward and inverse mathematical modelling programs. A number of assumptions are required to ensure the data can meaningfully be represented by a layered earth model. We used the TEMIXGL software package for modelling the data from the Fraser Valley.

GEOLOGICAL SETTING

The lower Fraser Valley has a complex geological history (Armstrong, 1957, 1980, 1981; Armstrong and Hicock, 1980; Clague et al., 1983). Quaternary Salish sediments consisting of stream deposits including channel fill, floodplain and overbank deposits, lacustrine silts and clays, bogs, swamps and shallow lake deposits and Fraser River sediments cover Pleistocene sediments. Pleistocene sediments consist of glacial, fluvial, glaciomarine, and marine sediments. Pleistocene sediments are commonly exposed with no Quaternary sediments deposited over them. This complex package of Quaternary and Pleistocene sediments is deposited on Tertiary bedrock which has an irregular surface.

Resistivity variations within this package of sediments are therefore complex, consisting of conductive marine and lacustrine sediments, more resistive sands and gravels and highly resistive bedrock. In some instances, Pleistocene sediments contain saline water, probably trapped 12 000 years ago or more.

ELECTROMAGNETIC DATA

The Geonics EM-47 surveys were carried out between June 20 and July 9, 1994. The locations of the 9 sites are shown in Figure 3. Approximately 100 central loop soundings were made at 9 survey sites during 15 working days; about 6.5 soundings per day on average. Most of the soundings were made using an 80 m square loop, although 20 m and 40 m square loops were used as well. The transmitter current was kept as close as possible to 2.3 A for all soundings. Data were collected at the 3 frequencies of the EM-47 system (UH, VH and H) for each sounding.

Three sets of data were collected at each frequency to provide averaging to increase the signal to noise ratio. Several of the soundings had noise that needed to be removed by editing before interpretation. In general, the Fraser Valley data were more noisy than the data collected over the Oakridges Moraine during the fall of 1993 (Todd et al., 1993). There are more electromagnetic sources due to the close proximity of airports (Vancouver and Abbotsford airports) and marine navigation and communication systems. In addition, several sources of local noise existed, often associated with large equipment such as trucks, graders, rock crushers and rock sorters (for example at Stokes gravel pit).

Figure 4a illustrates typical data for 2 soundings using an 80 m transmitter loop and one (AGA-120) using a 40 m transmitter loop. Log-log plots of normalized voltage (nV/Am^2), which are proportional to the time rate of change of the magnetic field (Fitterman and Stewart, 1986) are provided in Figure 4a.

The normalized voltage (Fig. 4a) of the sounding along the dyke by the Coalport (DHH0S, site 9 in Fig. 3) at borehole 85-5 (Hunter et al., 1993) illustrates an area with little noise at long times (7 ms). The normalized voltages at long times are nearly equal to 10 nV/Am^2 with no scatter. A noise source, repeatable for all 3 UH data sets on all 3 soundings carried out at the site, occurs approximately between 0.08 and 0.2 ms. The overlap of data points between the UH (square boxes) and VH (diamonds) frequency ranges permits the removal of this noise.

The normalized voltage plot of a sounding at the Agassiz experimental farm in the district of Matsqui (AGA-120, site 7 in Fig. 3) is also shown in Figure 4a. The first 3 time windows (UH frequency range) are saturated (the next 2 times are not saturated but are affected by the saturated values; Geonics EM-47 manual (Geonics, 1991)). Saturation occurs when the voltages, measured in millivolts, exceeds a pre-determined cutoff voltage of 5000 mV. This normalized voltage plot, and the normalized voltage plot of a sounding near Vedder Mountain southeast of Abbotsford (VED-100, site 8 in Fig. 3) in Figure 4a, have electromagnetic noise at longer times (greater than 1 ms) for both the VH and H frequency ranges. Note the data points have normalized voltage values below 1 nV/Am^2. Several other soundings within the Fraser Valley data set have even more noise at longer times. These sites tended to be in areas where the resistivities were higher so the magnetic fields decayed more quickly, resulting in lower voltages at longer times. Within the overall survey, the data quality was generally very good between 0.01 ms and 1 ms, and in many cases was good over a wider time window.

Log-log plots of the apparent resistivity computed from the normalized voltage versus time for these same soundings are given in Figure 4b. They illustrate how the normalized voltage is converted into apparent resistivity. These curves provide the fundamental data for interpretation, assuming of course that the ground can be modelled as a multi-layered earth. In the next section, these examples are used to illustrate the editing and interpretation capabilities of the TEMIXGL software and to provide a sampling of the interpretations that these data generate in terms of multi-layer conductive earth models.

EDITING AND INTERPRETATION

The data in Figure 4a and b are typical of the soundings obtained during the survey. They require editing before interpretation can begin. Any saturated values at short times and noisy values (usually less than 1 nV/Am^2) at long times are first removed from the data. Sometimes other specific values (for example the Coalport data for the UH frequency between 0.08 and 0.2 ms) are removed as well. The edited apparent

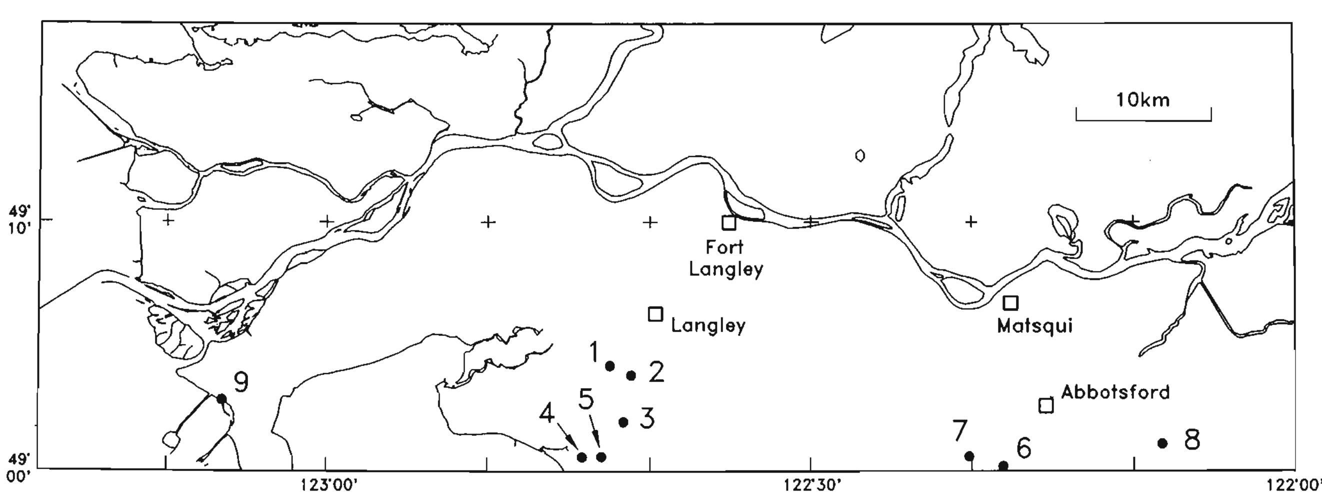


Figure 3. Location map of 1994 EM-47 surveys in the Fraser Valley: (1) Forest Nursery; (2) CRT Site; (3) Stokes Gravel Pit; (4) Farm (8 Ave. & 176 Street); (5) Farm (8 Ave. & 188 St.); (6) Short Street; (7) Agassiz Farm; (8) Vedder Mountain; (9) Coalport Dyke.

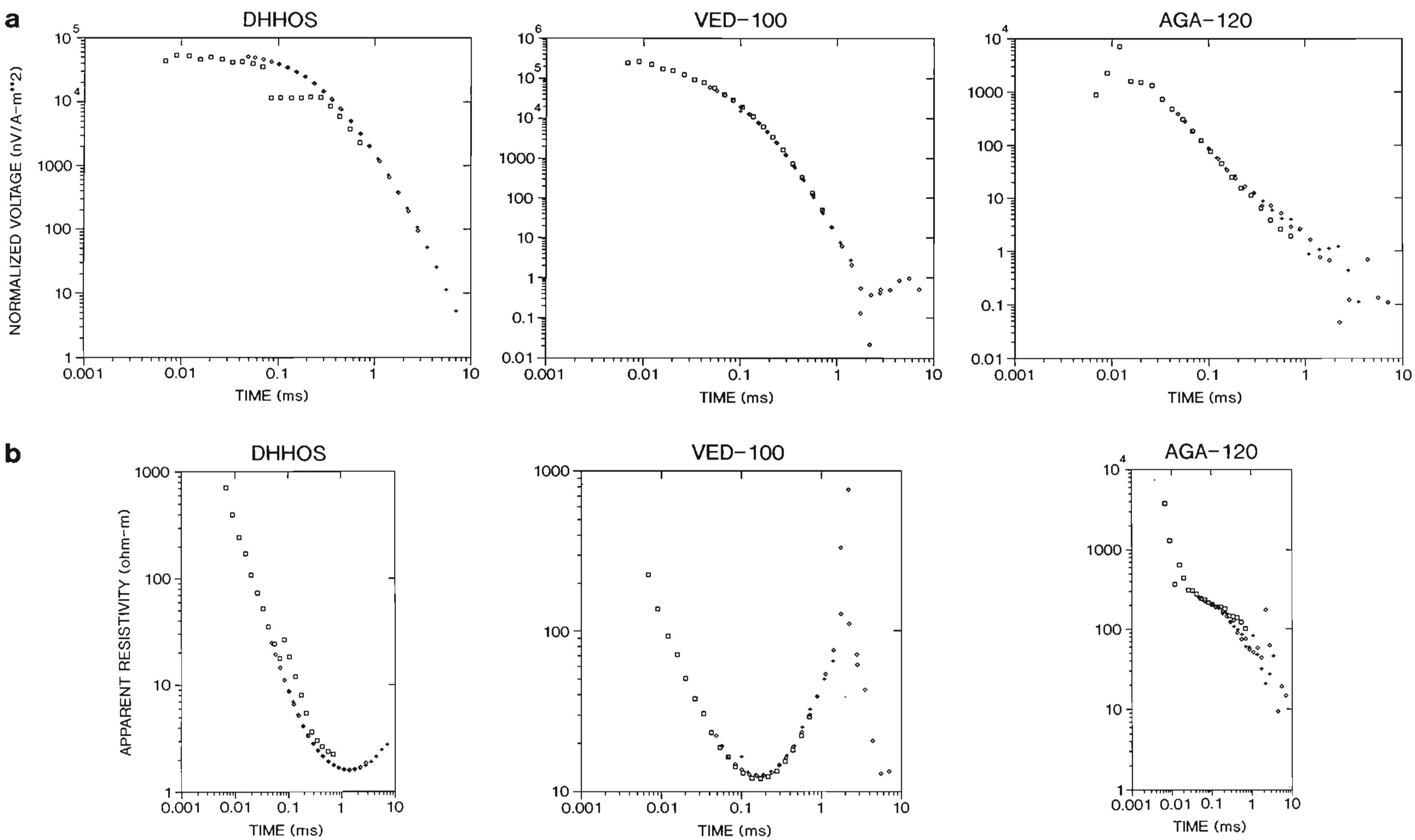


Figure 4. ***a)*** *Log-log plots of unedited normalized voltage (nV/Am^2) versus time for 3 soundings. Note the scales are different for the 3 illustrations: DHHOS - Coalport dyke near drill hole 86-5 (site 9 in Fig. 3); AGA-120 - Agassiz Experimental farm (site 7 in Fig. 3); VED-100 - Vedder Mountain (site 8 in Fig. 3).* ***b)*** *Log-log plots of unedited apparent resistivity (ohm.m) versus time for the 3 soundings in Figure 4a. Again, the scales are different for the 3 illustrations.*

resistivity data for the same soundings in Figure 4b are given in Figure 5. The TEMIXGL software has good editing capabilities using a mouse so the process is fast and efficient. Edited data can be stored for future processing and interpretation. After editing, the apparent resistivity soundings are ready for interpretation.

Interpretation using the TEMIXGL software package is fast because of its interactive capabilities. In this short paper, inter- pretation of these 3 examples is all that space allows. Neverthe- less they illustrate the type of layered conductivity models that can be used to represent the conductivity distribution of the sediments within the lower Fraser Valley and Fraser Delta.

The solid line in each of the 3 apparent resistivity plots in Figure 5 is the apparent resistivity curve computed for the multi-layer conductivity model shown graphically in the same figure. These models were computed using the interactive graphics option provided with TEMIXGL and the TEMIXGL Occum inversion option. Inversion was only performed once the computed apparent resistivity values were reasonably close to the measured data (fitting error less than 75% for example).

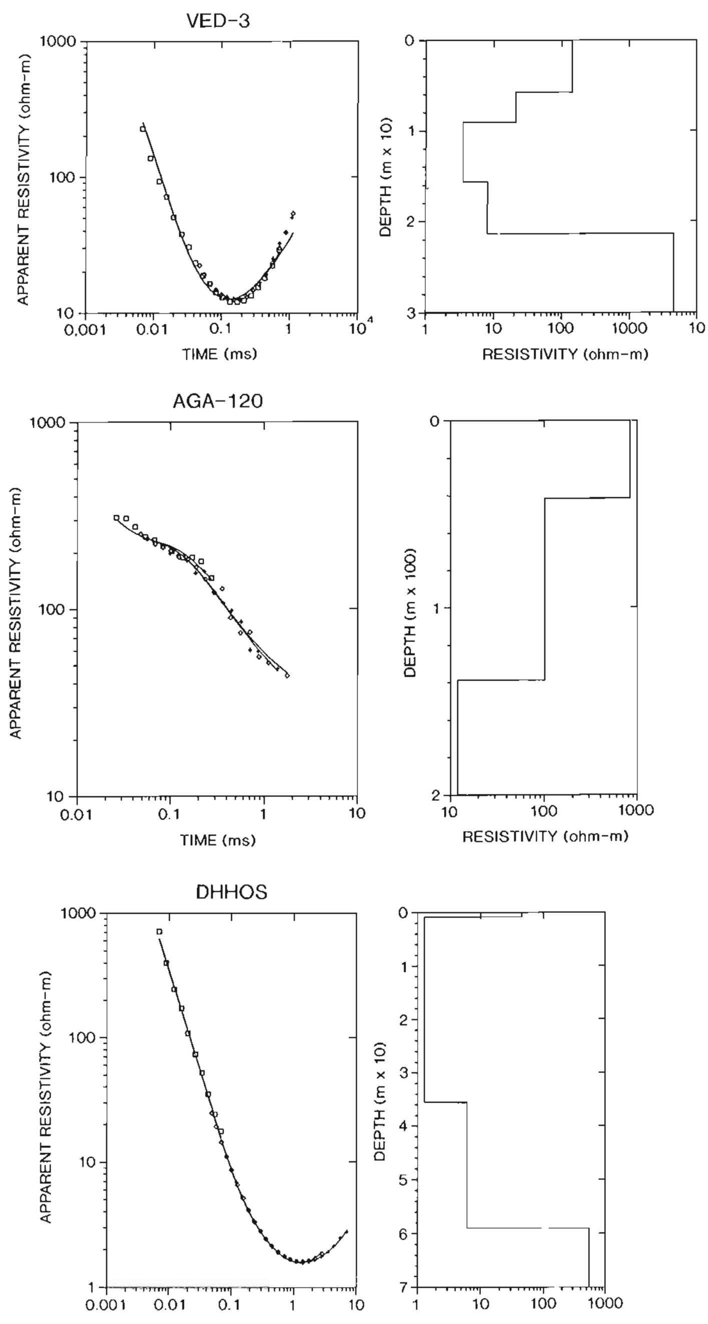


Figure 5.

Log-log plots of edited apparent resistivity (ohm.m) versus time for the 3 soundings in Figure 4a and b (VED-3 is identical to VED-100). The scales are different for the 3 curves. The solid lines correspond to the apparent resistivity computed for the layered earth models shown in the figure next to the data. When the overlapping data from the 3 frequencies are not identical, the TEMIXGL software fits each of the 3 curves separately. This leads to the 2 solid curves for AGA-120 where the UH and VH frequencies overlap around 0.2 ms.

DHHOS

This sounding was made along the dyke by the Coalport near borehole FD86-5. The conductivity log in Figure 6 shows that a layer of salt water sits on top of fresh water. The contact between very conductive water (650 to 950 mS/m) and brackish water (250 to 400 mS/m) is approximately 35 m below the top of the borehole. The brackish water becomes fresher with depth, reaching a conductivity of 20 mS/m around 70 m depth. The change from very conductive salt water to brackish or fresh water usually occurs near the top of the Pleistocene in this region of the Fraser Delta and is therefore a marker for the top of the Pleistocene. The top of the Pleistocene is quite variable and often difficult to see on reflection seismic data because of biogenic gas masking the seismic signal.

A 4-layer model was used to represent the very shallow resistive layer near the surface, the salt water layer, the brackish water layer, and finally the resistive Pleistocene at the bottom. The layer depths and resistivities (inverse of conductivities) were unconstrained in the inversion, which took 3 iterations to reduce the fitting error to approximately 5 per cent (Table 1). The resistivities obtained from the salt water layer and brackish layer were 1.26 ohm.m (793 mS/m) and 6.14 ohm.m (162 mS/m) respectively. These values are consistent with the values from the conductivity log. The thicknesses obtained from the inversion for the salt water layer and the brackish layer were 34.7 m and 23.6 m respectively, again consistent with the values from the conductivity log. It is interesting to note the dipole-dipole resistivity survey carried out about 1.5 km north of this sounding and CSAMT soundings along the Coalport causeway (Nobes et al., 1990) used a 3 layer model (they did not resolve the shallow layer) with resistivity values that were not too different from those obtained from our inversion.

The pattern of resistive-conductive-resistive layers leads to the apparent resistivity behaviour of sounding DHHOS. As an aside, the minimum in this curve moves to longer times as the thickness of the conductive layer increases. The maximum thickness that can be seen using the EM-47 system is about 100 m because the minimum in the apparent resistivity curve is equal to the maximum time of the system (7 ms). Other EM systems that measure the decaying voltage at longer times are available however.

AGA-120

This sounding was made at the Agassiz Experimental Farm in Matsqui District. The surficial geology at this site is Pleistocene Sumas drift (Armstrong, 1980, 1981) overlying Fort Langley Formation sediments. Sumas drift consist of glacial

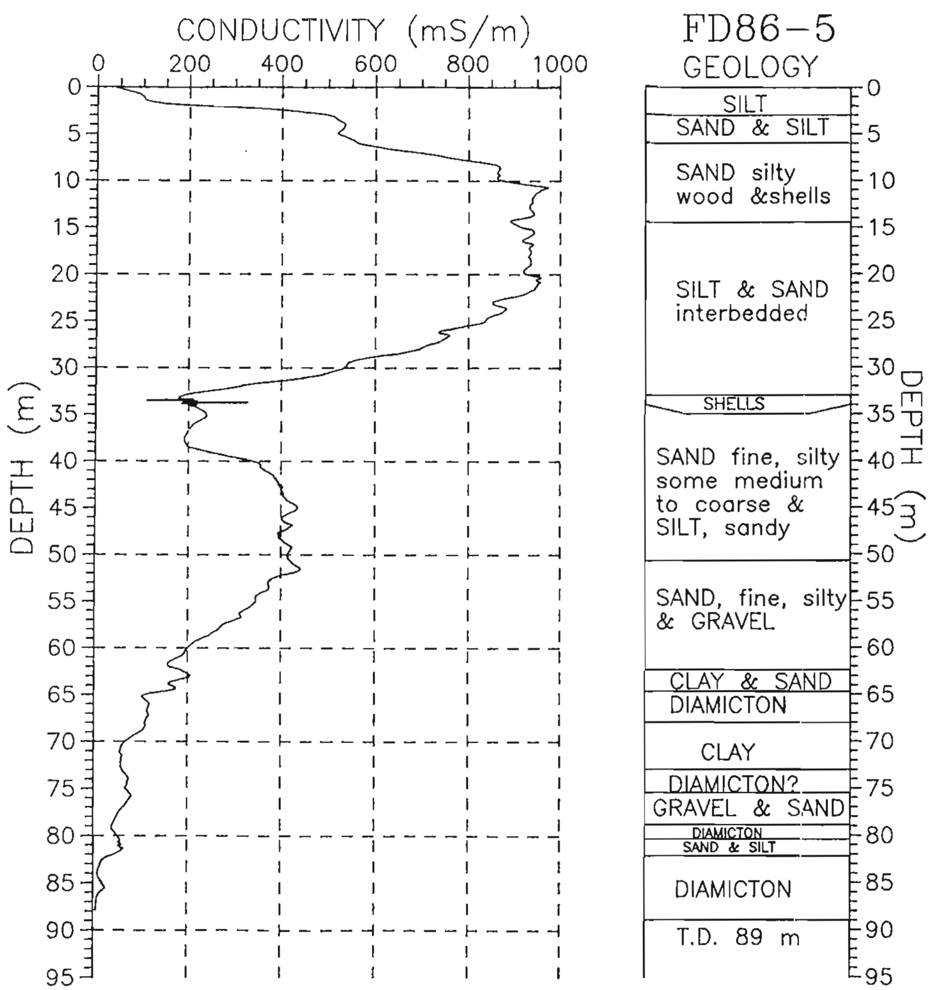


Figure 6.

Conductivity and geological logs for borehole FD86-5 (from Hunter et al., 1993).

Table 1. Error in the layered earth fit for the three examples along with the number of layers and number of iterations used in the inversion. The transmitter loop dimension is included for completeness.

Sounding	Fitting error (%)	Number of layers	Length of side of Tx (m)	Number of iterations for inversion
DHHOS	5.15	4	80	3
AGA-120	9.63	3	40	3
VED-100	18.82	5	80	hand fit

outwash sands and gravels and sands containing clays. The Fort Langley Formation consist of marine and glaciomarine sediments.

The character of this sounding is completely different than DHHOS. A 3-layer model was used to interpret this sounding. Again, an unconstrained inversion was performed after an approximate fit was obtained using interactive forward modelling. The inversion took 3 iterations to produce a fitting error of approximately 9.6 per cent. The noise still contained in the edited data prevents the fitting error to be as low as the fitting error for DHHOS. However, the error is still quite acceptable and the observed fit to the data looks good.

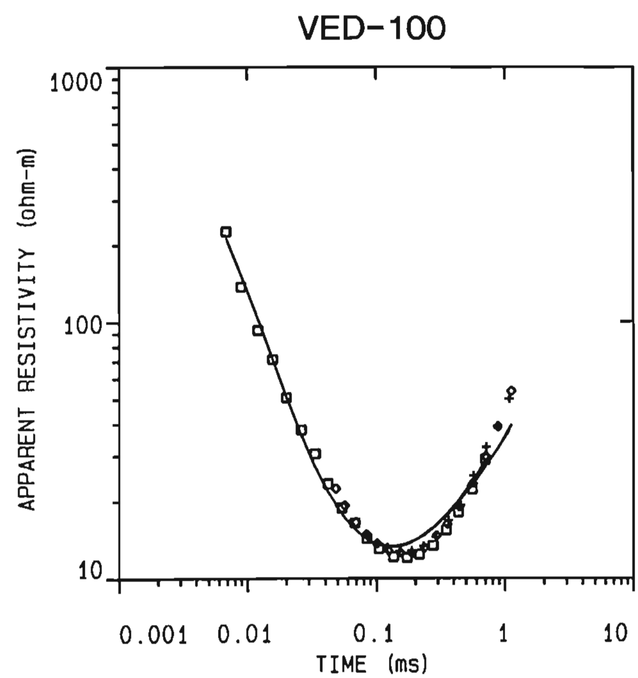
The interpretation is also significantly different than DHHOS. There is no salt water in this area so the resistivities reflect the variations in lithology. The top layer, consisting of 41.4 m of 840 ohm.m material, overlays a 97 m thick layer of 100 ohm.m before reaching more conductive material (12 ohm.m). The resistive material at the surface is Sumas drift consisting of sands and gravels with little or no clay in it. The intermediate layer is most likely Sumas drift but with increased clay content. The underlying layer is probably marine and glaciomarine Fort Langley Formation sediments.

VED-100

The Vedder Mountain sounding was carried out on a farm at the base of Vedder Mountain (note that VED-3 is the same sounding as VED-100). The objective of this sounding was to see if the bedrock (Vedder Mountain crystalline rocks) could be mapped beneath the surficial material on the surface. The shape of the apparent resistivity curve immediately tells us that the model will be resistive-conductive-resistive since its shape is similar to DHHOS. The resistivities and thicknesses are quite different from sounding DHHOS.

The data were fit using a 5-layer model. We found it difficult to fit long times for this apparent resistivity curve, a problem not too uncommon when the bottom layer is highly resistive. This could also be caused by using a layered model interpretation when the resistivity structure is non-layered. The data, fit by hand using the interactive forward model routine, produced a fitting error of approximately 18.8 per cent. Unconstrained inversion of this sounding tended to move the minimum towards shorter times in order to minimize the overall error (which actually was around 19% to 20%).

Another option within the TEMIXGL software package is the smooth model option. The smooth model option uses up to a maximum of 20 layers, with each layer's thickness increasing in a logarithmic progression from shallow to deep. The overall thickness can be determined from the data or it can be fixed to pre-set values. The smooth model uses inversion to fit the data layer by layer. Figure 7 shows the results of a 10 layer smooth model fit to the sounding at VED-100.



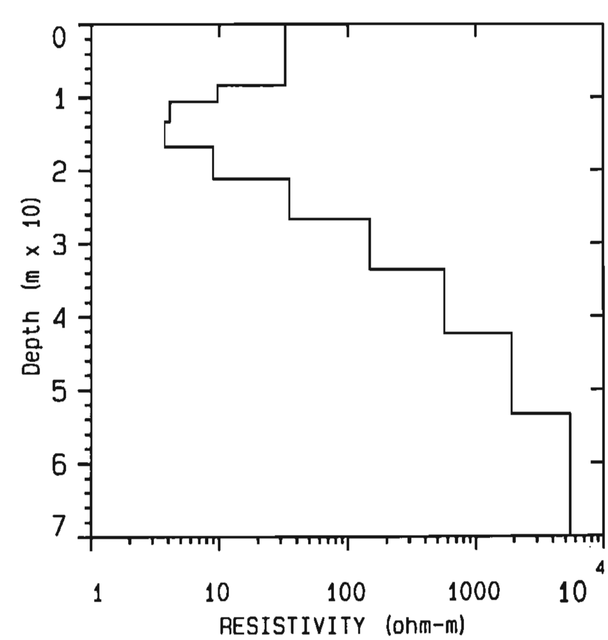


Figure 7. Apparent resistivity curves for sounding VED-100 with the smooth model interpretation using 10 layers. The fitting error for this model was the same as the 5 layer hand fit model discussed earlier (Table 1).

There are quite significant differences between this model and the one in Figure 5. They both show the same pattern of resistive-conductive-resistive.

The surface sediments in this area are Quaternary lacustrine silts and clays (Armstrong, 1981). When they exist, the underlying sediments are Fort Langley Formation marine and glaciomarine sediments. The 140 ohm.m layer near the surface is most likely lacustrine silts and clays overlying conductive (3.5 to 21 ohm.m) marine sediments. The resistive bottom layer (4500 ohm.m) is crystalline bedrock.

SUMMARY

The application of time-domain electromagnetic methods for mapping the surficial geology associated with ground water aquifers in the lower Fraser Valley and Fraser Delta shows considerable promise. The examples presented here illustrate the ability of EM to map a variety of geological environments as well as water quality (salinity). We will continue to learn more about the limitations and strengths of TDEM systems as we continue to analyze the data collected during this survey.

We must point out that no single geophysical method can effectively map all geological environments nor can a single geophysical method necessarily produce the required accuracy even for one environment. A combination of one or more of seismic, electromagnetic, ground penetrating radar and resistivity methods is essential to provide a complete picture of the subsurface. An important step in the Fraser Valley hydrogeology project is to investigate how each of these geophysical methods performs in different environments by comparing their strengths and weaknesses and then investigating which combinations are the best. Future research will address these issues and will provide a more complete interpretation of the data collected during the last year.

ACKNOWLEDGMENTS

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Geological Survey of Canada Project 940010

Shear wave velocities of Holocene sediments, Fraser River delta, British Columbia

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Hunter, J.A., 1995: Shear wave velocities of Holocene sediments, Fraser River delta, British Columbia; in Current Research 1995-A; Geological Survey of Canada, p. 29-32.

Abstract: Shear wave velocity measurements of Holocene deltaic sediments have been compiled for 119 surface and borehole seismic sites in the Fraser River delta to depths of 100 m. The data show a narrow velocity-depth distribution for widely separated sites. A least-squares velocity-depth function has been fitted to the data so that this analytical expression may be used as one input parameter for earthquake ground-motion amplification modelling and cyclic liquefaction studies.

Résumé : Les vitesses des ondes de cisaillement se propageant dans les sédiments deltaïques holocènes ont été enregistrées à 119 sites sismiques situées à la surface ou dans des trous de sondage dans le delta du fleuve Fraser jusqu'à des profondeurs de 100 m. Les données indiquent une distribution vitesse-profondeur étroite pour des sites très éloignés. Une fonction vitesse- profondeur par les moindres carrés a été appliquée aux données de façon à ce que cette expression analytique puisse être utilisée comme paramètre d'entrée pour la modélisation de l'amplification des ébranlements du sol dus à des séismes et pour la réalisation d'études sur la liquéfaction cyclique.

INTRODUCTION

Since 1987, the Geological Survey of Canada (GSC) has been measuring seismic shear wave velocities of the Holocene sediments of the Fraser River delta, British Columbia. The purpose of this work is to provide regional information to geotechnical engineers to use as a guide in estimating liquefaction potential, as well as modelling ground-motion amplification for earthquake studies.

As a result of recent studies by various workers (Robertson et al., 1992; Kayen et al., 1993), shear wave velocities can be utilized in estimation of liquefaction potential of water-saturated cohesionless soils, if the near-surface cyclic shear stress parameters of the "design" earthquake are defined.

Shear wave velocity-depth information is also required for ground-motion amplification studies of the soil column (Byrne and Anderson, 1987; Finn and Nichols, 1988). Large variations between earthquake motions measured on rock and those measured on thick unconsolidated overburden are possible. At the surface of overburden, amplification of ground motion at particular frequencies can occur, and such spectral peaks result in part from the shear-wave velocity distribution within the soil column.

The GSC field program of data acquisition is designed to provide shear wave velocity-depth estimates on a regional scale, by occupying widely spaced field sites within the delta. The work uses three differing shear wave seismic techniques; these are: surface shear wave refraction (reversed profiling in horizontally polarized SH mode), surface-to-borehole SH mode measurements (using a well-locked 3-component geophone system), and the seismic cone penetrometer (witha horizontal geophone installed in the penetrometer and the seismic source on surface). These methods have been compared in detail by Hunter et al. (1991).

THE DATA SET

To date, results from 85 surface refraction sites, 53 seismic cone penetrometer sites, and 6 borehole 3-component sites have been released as GSC Open Files (Hunter et al., 1990, 1992; Woeller et al., 1993a, b, 1994). The locations of these sites in the delta are shown in Figure 1. This data set has been carefully examined and edited in order to compile the velocity-depth distribution for Holocene sediments to a depth of 100 m below surface that is presented in this paper. Some of the boreholes, and several of the surface refraction sites, encountered high velocity Pleistocene sediments at depth; the compilation technique included interpretation of the top of Pleistocene sediments and removing velocities below this level from the data set.

The compiled velocity-depth plot is shown in Figure 2. For surface refraction sites, the plotted values are the depths to the top of "velocity layers" and are irregularly spaced vertically. For both the seismic cone and the borehole 3-component data, the plotted measurements are those made over a vertical interval of 1 m.

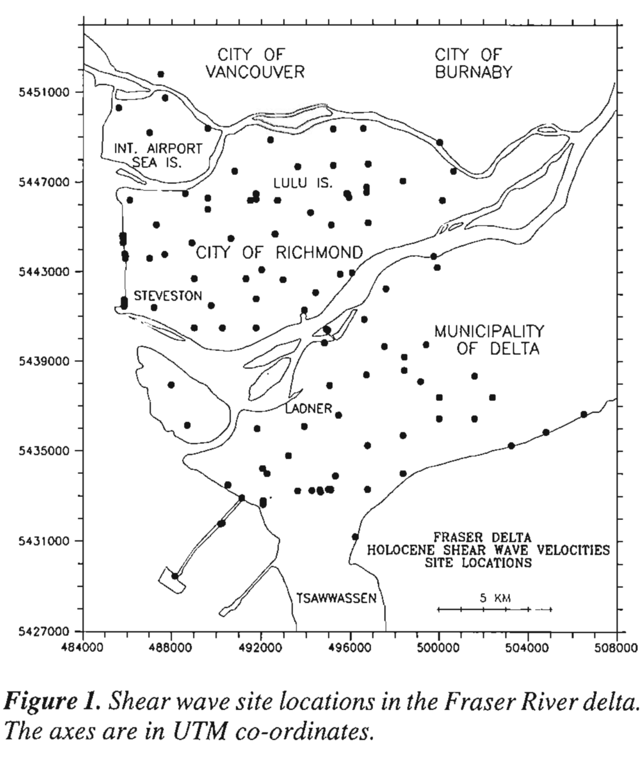

Figure 1. Shear wave site locations in the Fraser River delta. The axes are in UTM co-ordinates.

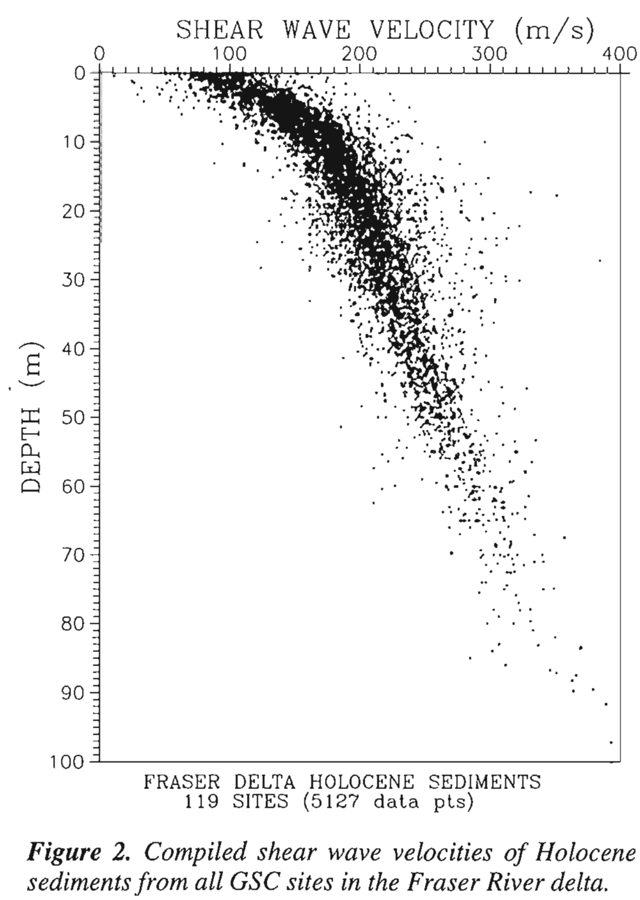

Figure 2. Compiled shear wave velocities of Holocene sediments from all GSC sites in the Fraser River delta.

Although slight variation of velocity-depth determinations with areal distribution of sites were noted, no attempt has been made at present to subdivide the data set within the delta. In future, when more complete coverage of the delta has been attained, such an option will be examined.

In an attempt to compare shear wave velocity data obtained from the Fraser River delta with shear wave measurements made in Holocene sediments elsewhere, the United States Geological Survey (USGS) data from borehole measurements in the San Francisco Bay and Los Angeles basin were examined (Fumal et al., 1981, 1982, 1984; Gibbs et al., 1975, 1976, 1977, 1980). Only those sites having Holocene sediments similar to the Fraser delta sediments (sands, silts and clays) were compiled. These data are shown in Figure 3. Though this is a smaller data set than shown in Figure 2, it is immediately obvious that the USGS data shows considerable variation in comparison to the Fraser River delta compilation; such apparent scatter can be attributed to the wide variation in depositional environments in the Los Angles basin and the San Francisco bay areas relative to the rather uniform deltaic environment of the Fraser River area. The shear wave velocities obtained within the Fraser delta are somewhat unique in that they lie within a narrow range as a function of depth.

DATA ANALYSIS

A polynomial was fitted to the data set shown in Figure 2. Assuming errors in velocity only, Chauvenet's Rejection Criterion (Worthing and Geffner,1959, p. 170) which is tied to the standard deviation of the fit, was successfully applied. Over 4 successive cycles of rejection, only 30 points were removed from the data set of over 5000 measurements. The final data set and the polynomial fit with two standard deviation error curves (96% confidence limits) are shown in Figure 4. The polynomial equation and coefficients are as follows:

$$V_s = A_0 + A_1 * D + A_2 * D^2 + A_3 * D^3 + A_4 * D^4 + A_5 * D^5 + A_6 * D^6$$

where:

D = depth below surface in metres

A_0= +87.7785	A_4= -2.22999E-04
A_1= +13.1673	A_5= +1.49642E-06
A_2= -0.615491	A_6= -3.85033E-09
A_3= +0.01634	

The standard deviation of the fit is 21.5 m/s. The above expression is based on 5097 data points from 119 sites and is valid for Holocene materials at depths of less than 100 m.

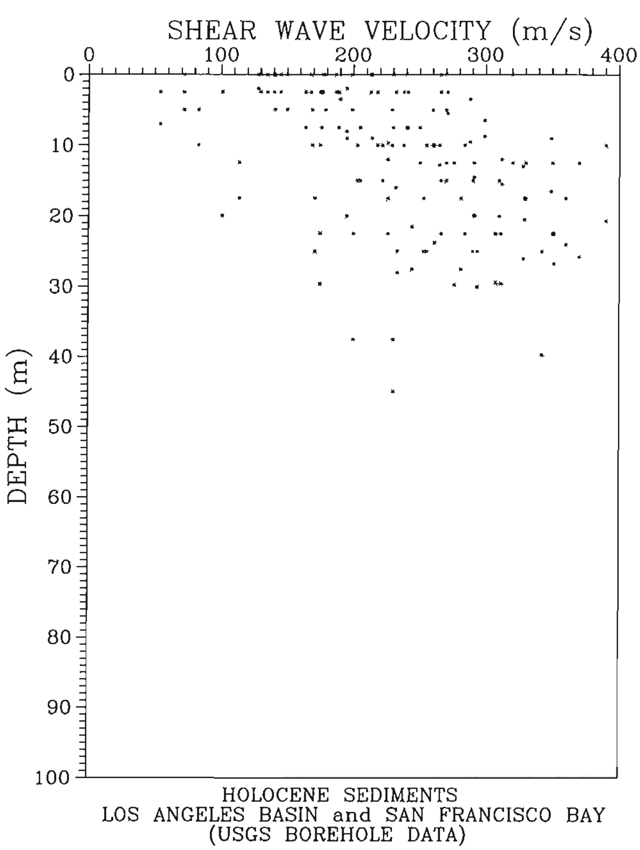

***Figure 3.** Compiled shear wave velocities of Holocene sediments from USGS borehole data for the San Francisco Bay and the Los Angeles basin areas.*

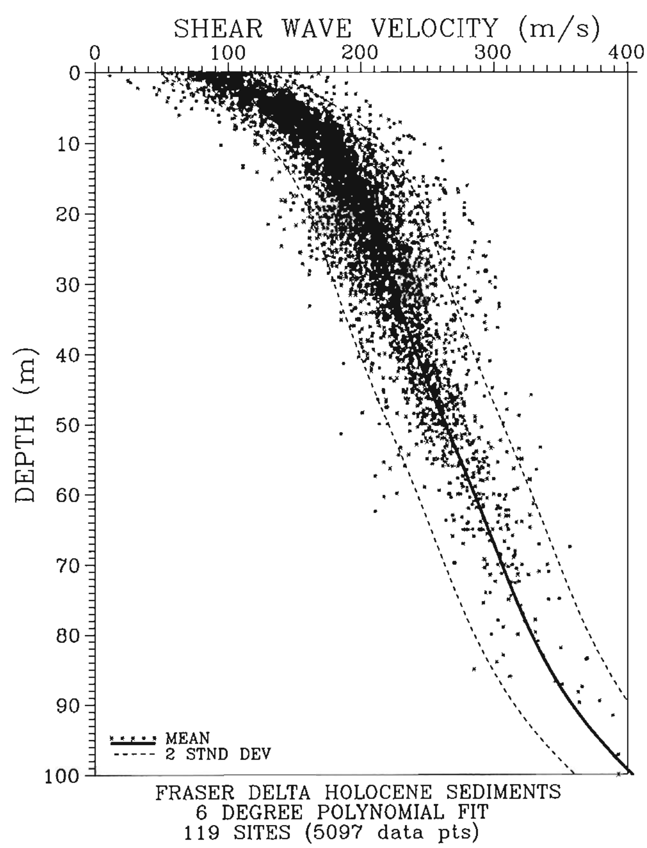

***Figure 4.** Least-squares polynomial fit to the data after Chauvenet rejection.*

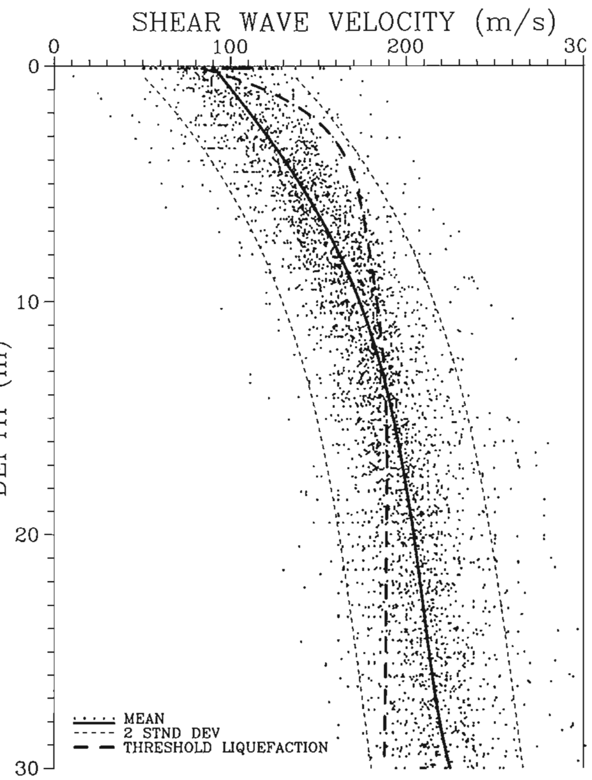

Figure 5. The Robertson/Earthquake Task Force threshold cyclic liquefaction curve superimposed on the Holocene data for the Fraser River delta.

DISCUSSION

Robertson et al. (1992) have suggested an empirical relationship between shear wave velocity and cyclic stress ratio which indicates threshold values for liquefaction of water-saturated non-cohesive soils. Using the cyclic stress ratio-depth relationship recommended for the Fraser delta by the Earthquake Task Force (1991), and the threshold cyclic liquefaction curve of Robertson et al. (1992), a threshold cyclic liquefaction-depth curve can be superimposed on the polynomial velocity fit (Fig. 5) for depths to 30 m. Those data points lying to the left of the threshold curve are susceptible to liquefaction. These results suggest that liquefaction effects might occur to an average depth of 13 m below surface for the design earthquake (magnitude 7.0) given by the Task Force Report. It is interesting to note that the threshold cyclic liquefaction curve falls entirely between the 96 per cent confidence limits. In the depth range 0 to 30 m, approximately 55 per cent of the data points lie to the left of the threshold liquefaction curve.

It is hoped that the compilation of shear wave velocities in the Holocene sediments as given above can be utilized as a guide by practitioners of earthquake modelling within the Fraser River delta. In future, when regional surveys are completed, an areal distribution of velocity-depth variations is planned.

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Geological Survey of Canada Project 890023

Ground penetrating radar survey of the Katz slide, southwestern British Columbia

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Brooks, G.R. and Pilon, J.A., 1995: Ground penetrating radar survey of the Katz slide, southwestern British Columbia; in Current Research 1995-A; Geological Survey of Canada, p. 33-40.

Abstract: The Katz slide near Hope, British Columbia, is the site of two large prehistoric rock avalanches. Much of the debris field of the older of the two landslides is buried beneath floodplain sediments, thus its distribution over the valley bottom is not known exactly. Ground penetrating radar (GPR) was used over this older deposit to test the ability of the equipment to resolve buried, very coarse substratum and to map the extent of the deposit. Seven GPR profiles were obtained with most of the returns containing two distinctive GPR facies. One facies consists of chaotic, discontinuous, hyperbolic reflectors that are interpreted as blocky landslide debris. The other facies contains laterally continuous, horizontal reflectors which represent floodplain deposits. As mapped using GPR data, the toe of the older debris field appears truncated by the Fraser River, suggesting that landslide debris once extended further across the valley bottom.

Résumé : Deux importantes avalanches de pierres se sont produites au cours de la Préhistoire au site du glissement de Katz, près de Hope, en Colombie-Britannique. Le champ de débris du plus ancien des deux glissements est en grande partie enfoui sous des sédiments de plaine d'inondation, et il en résulte que sa distribution dans le fond de la vallée n'est pas connue avec exactitude. Un géoradar a été utilisé au-dessus de ce dépôt plus ancien pour étudier la capacité de cet instrument de cerner un substratum enfoui, très grossier, et pour cartographier l'étendue du dépôt. Sept profils de géoradar ont été obtenus, et la plupart des échos indiquaient deux faciès distincts. L'un d'eux consiste en des réflecteurs chaotiques, discontinus et hyperboliques, que l'on croit être des débris blocailleux de glissement de terrain. L'autre faciès contient des réflecteurs horizontaux, continus latéralement, qui représentent des dépôts de plaine d'inondation. Sa cartographie ayant ainsi été dressée au moyen des données du géoradar, le front du champ de débris plus ancien semble être tronqué par le fleuve Fraser, ce qui semble indiquer que les débris du glissement de terrain se sont déjà étendus plus loin dans le fond de la vallée.

INTRODUCTION

Along the upper Fraser Lowland, southwestern British Columbia, steep valleysides are periodically subject to mass movements. Historically, mass movements have occurred as debris flows (e.g., Evans and Lister, 1984), however, these events are of relatively small magnitude. Rock avalanches caused by catastrophic failure of the valleyside represent threats of much lower frequency but far greater magnitude. The locations of several large prehistoric rock avalanches have been identified within the upper Fraser Lowland (Cheam slide, Katz slide, and Lake of the Woods slide (e.g., Savigny and Clague, 1992)). These sites have been subject to varying degrees of research, but much work remains to be done. Complicating this, however, significant geomorphic change has occurred at some sites making interpretation of events difficult.

As described below, landslide debris at the Katz slide site is partially buried beneath Fraser River floodplain sediments. The site seemed an ideal location to test the suitability of ground penetrating radar (GPR) to resolve buried very coarse landslide debris. Survey results could then be used to map buried landslide debris, the extent of which is not specifically known and otherwise could only be determined by much more costly drilling. The purpose of this paper is to present: 1) the results of the GPR survey and 2) a map showing the distribution of buried landslide debris in the area where the GPR survey was conducted.

Figure 1. Location of the study site within southwestern British Columbia.

STUDY SITE

The Katz slide site is located on the north side of the Fraser Lowland, 10.5 km west of Hope, British Columbia (Fig. 1, 2). The site is crossed by the Lougheed Highway, the mainline CPR tracks, and a natural gas pipeline (Fig. 2). Two blocky debris fields are present which have distinctly different morphological characteristics.

One debris field consists of large, angular blocks that cover an area of about 1.1 km^2 (Thurber Engineering Ltd., 1988; 'younger debris field' in Fig. 2). It is adjacent to the north side of the valley and truncates a paleochannel of the Fraser River. Thurber Engineering Ltd. (1988), Naumann (1990), and Savigny (in press) interpreted this debris field as a rock avalanche deposit. Naumann (1990) interprets a radiocarbon age of 3260 ±70 BP (SFU 669) obtained from organic materials buried within the truncated paleochannel as representing the age of this landslide.

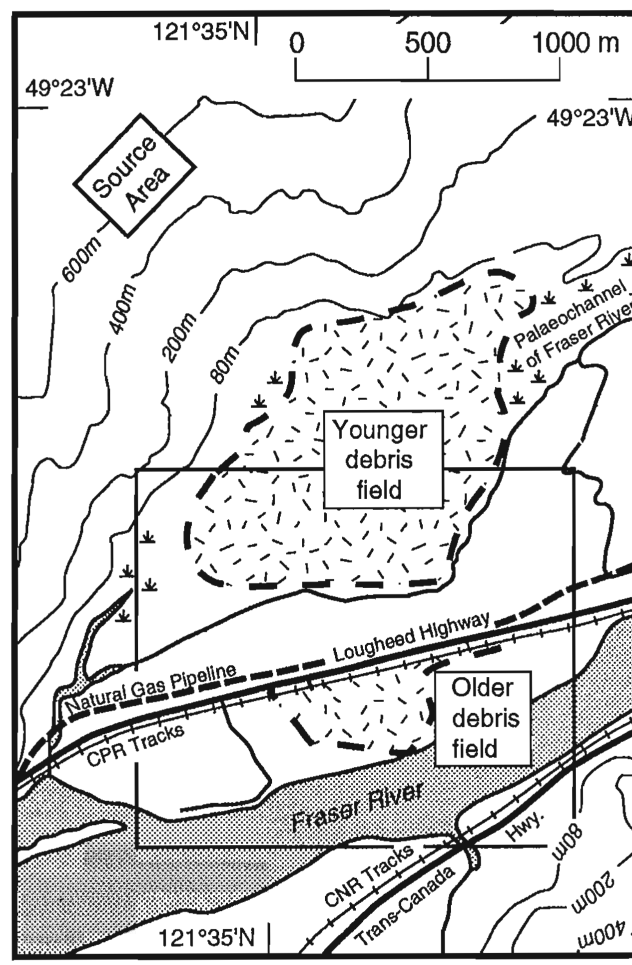

Figure 2. The Katz slide site showing the approximate distribution of the 'younger' and 'older' debris fields, both of which were produced by prehistoric rock avalanches. The large rectangle delineates the area shown in Figures 4 and 9.

The second debris field, which is the subject of the GPR survey, is situated between Lougheed Highway and the Fraser River channel ('older debris field' in Fig. 2). It consists of large blocks that protrude sporadically through contemporary floodplain deposits of the Fraser River (Fig. 3). The specific extent of this debris field is not known because much of the deposit is buried beneath the floodplain.

Thurber Engineering Ltd. (1988) suggested that blocks of the older debris field could be glacial erratics. Naumann (1990) and Savigny (in press), however, interpret the blocks to be the remnant deposit of a second, older rock avalanche. Since the debris field is situated towards the middle of the valley bottom within a Holocene floodplain (Fig. 2), a rock avalanche origin seems to be the most likely explanation.

Naumann (1990) suggests that this older rock avalanche dammed the Fraser River. The presence of large blocks along the northern bank of the Fraser River channel adjacent to the boulder field indicates that the older debris field extends to the river (Fig. 2). However, the majority of the blocks along the river are part of a riprap constructed to protect a sawmill yard from river bank erosion (Fig. 4). The riprap was taken from the older debris field during construction of the sawmill yard in the mid-1980s when blocks in the area now underlying the yard were extensively blasted (R. Skovsjold, pers. comm., 1994).

The source area of both rock avalanches is on the north side of the Fraser Lowland (Fig. 2; Thurber Engineering Ltd., 1988; Naumann, 1990; Savigny, in press). Here, bedrock in

Figure 3.

Photograph of a large block protruding through the Fraser River floodplain within the older debris field (GSC 1994-724B). Other protruding blocks (marked by white arrows) can be seen in the background.

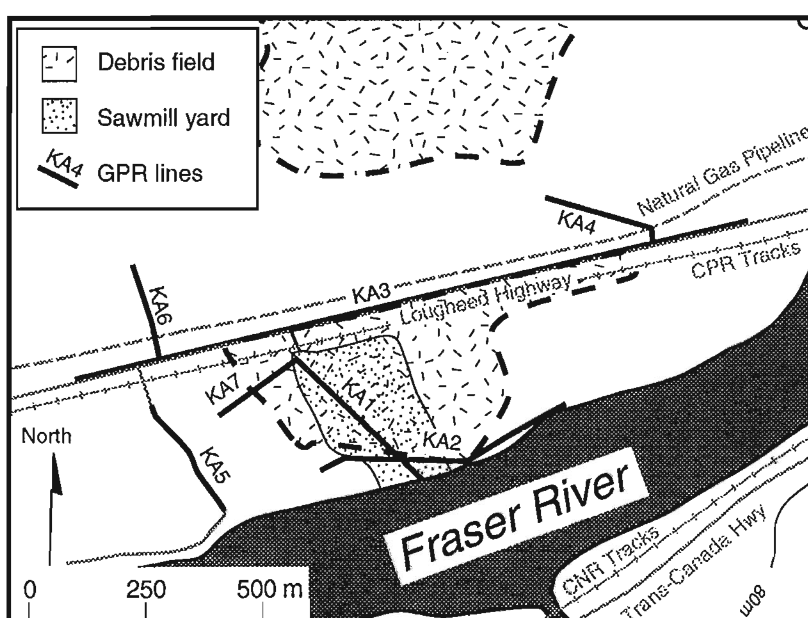

Figure 4.

Map showing the location of the seven GPR lines within the older debris field.

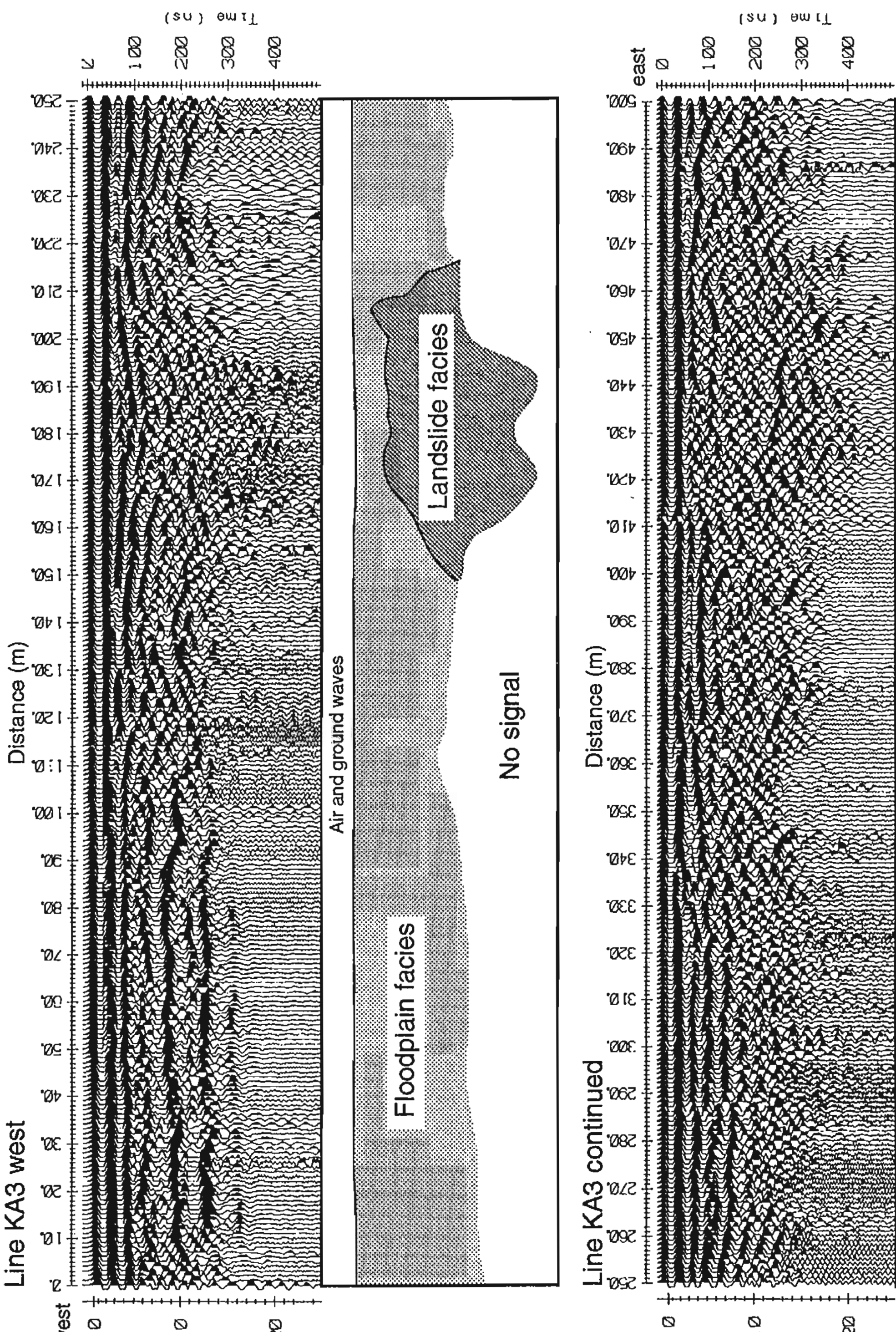

Figure 5. GPR returns of line KA3 between stations 0 and 500 shown in two parts with an interpretative diagram below each. Note onlap of floodplain facies upon landslide facies.

the approximately 700 m high valleyside consists of quartz diorite and diorite which form part of the Spuzzum Pluton (Monger, 1989). A strong north-northeast-trending lineament cuts across the top of the valleyside within the source area (Naumann, 1990; Savigny, in press). Naumann (1990), Savigny (in press), and Savigny and Clague (1992) hypothesize that valleyside instability originates from a process of mountain-top spreading and gravity faulting which results in southeast displacement and downward translation of the valleyside. A large graben in the headscarp area is believed to provide the lateral load on the rock mass. The authors suggest that the two catastrophic failures occurred along exfoliation surfaces thought to be present within the valleyside. Savigny (in press) also mentions possible neotectonic displacement along the lineament cutting across the top of the source area. Thurber Engineering Ltd. (1988) identified a remnant of intact rock that remains perched at the top of the source area and which could produce a third rock avalanche if it were to fail catastrophically.

METHODOLOGY

GPR is an emerging high resolution geophysical technique for studying overburden and bedrock geology. The basic principles of GPR have been described by Davis and Annan (1989) and are not repeated here.

GPR profiles presented in this paper were obtained with a pulseEKKO IV GPR system using a 1000 volt transmitter and a centre frequency of 50 MHz. Transmitter/receiver separation for reflection surveys was 2 m with a station increment of 1 m. A stack of 64 pulses at each station was used to maximize the signal to noise ratio and optimize survey efficiency. The speed of propagation of the radar signal was measured by conducting common midpoint (CMP) surveys with a station increment of 0.5 m. The calculated propagation speed was later used to create the depth scale along each profile. Post-acquisition processing involved application of an automatic gain control factor with a varying amount of high frequency filtering (point averaging of 5 to 10).

RESULTS

Seven GPR lines numbered KA1 to KA7 were measured in the general location of the older debris field (Fig. 4). These were conducted along existing roads and clearings and across a sawmill yard; dense secondary forest to the north of Lougheed Highway restricted the survey to these locations. Selected portions of profiles are depicted in Figures 5 to 8, space restrictions prevent the entire dataset from being shown. All collected GPR records extend over return times ranging from 190 to 430 ns. The velocity of propagation of the GPR signal calculated from two CMP surveys averages 0.098 mns^{-1}, indicating that the effective depth of signal penetration along the profiles ranges from about 10 m to 22 m. The air and ground waves that effectively mask the upper several metres of the profiles are two strong reflectors present within the initial 40 ns of every trace (e.g., Fig. 5).

Our interpretations of the GPR profiles are presented without of any ground truth (boreholes or sections), but are made with knowledge that blocky debris protrudes through the valley bottom either directly along or adjacent to lines KA1, KA2, KA3, and KA7. Interpretations of these returns are extrapolated to lines KA4, KA5, and KA6.

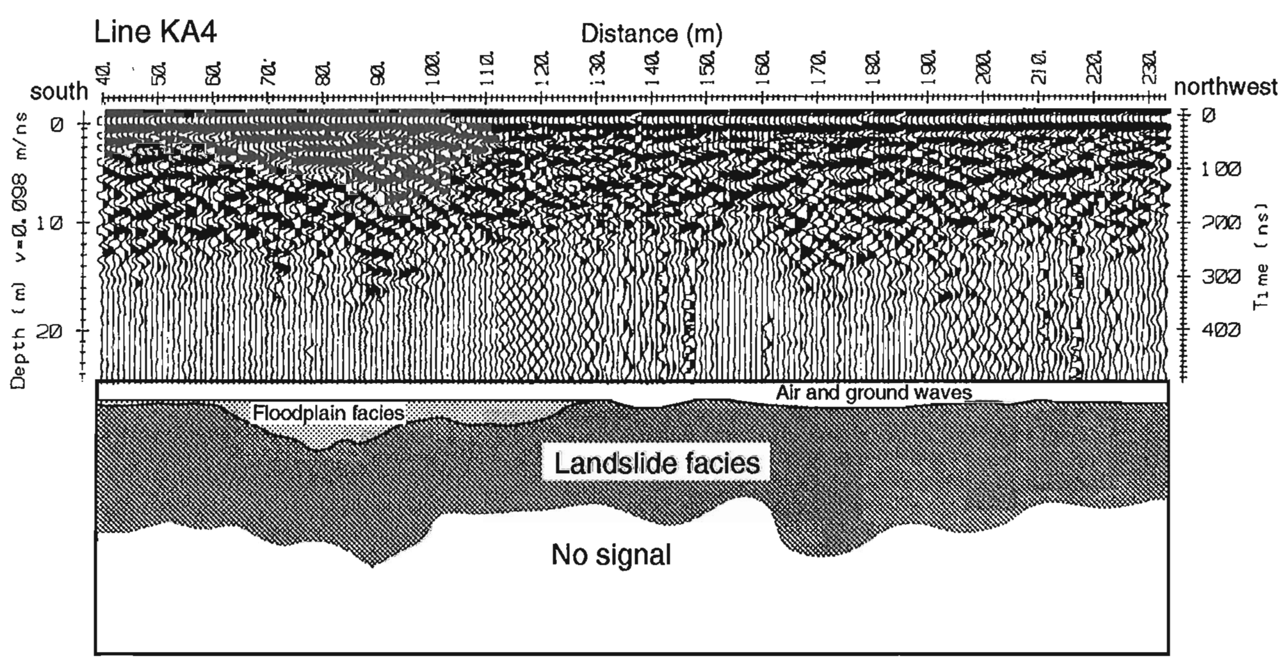

Figure 6. GPR returns of line KA4 between stations 40 and 233 with an interpretative diagram shown below. This part of the line consists of predominately landslide facies, part of which is overlain by a pocket of floodplain facies.

Two distinctive GPR facies are present over much of the profiles and are interpreted to represent landslide and floodplain deposits. Typically, the landslide facies appears as a chaotic assemblage of discontinuous, hyperbolic returns as exemplified in line KA3 between stations 410 to 500 (Fig. 5). Such returns are interpreted as point source reflectors from numerous large blocks under the surface. In marked contrast, the floodplain facies consists of laterally continuous, horizontal reflectors as shown, for example, in line KA3 between stations 0 to 100. The character of the floodplain facies is due to the dielectric contrast within the beds of the alluvial deposits.

The subsurface architecture of the deposits can be seen clearly along sections of the profiles. For example, in line KA3 between stations 260 to 410 (Fig. 5), the onlap of floodplain facies upon the margin of the landslide facies can

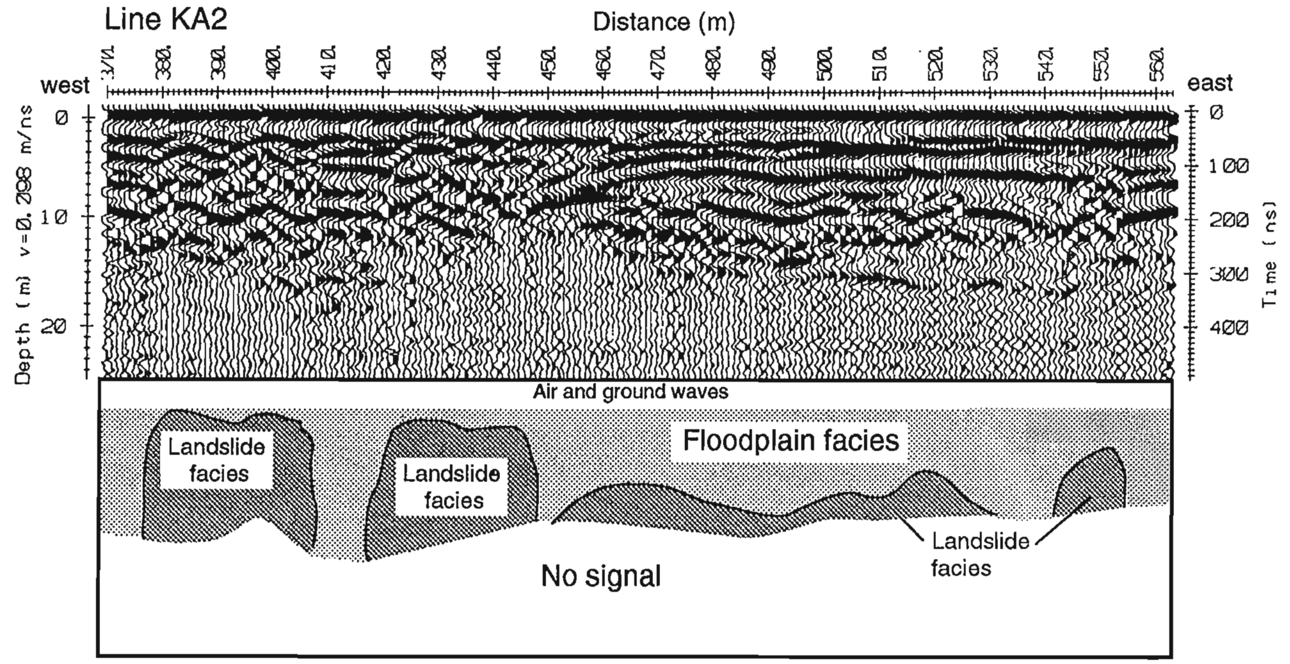

Figure 7. GPR returns of line KA2 between stations 370 and 563 with an interpretative diagram shown below. In this part of the line, landslide facies occurs discontinuously.

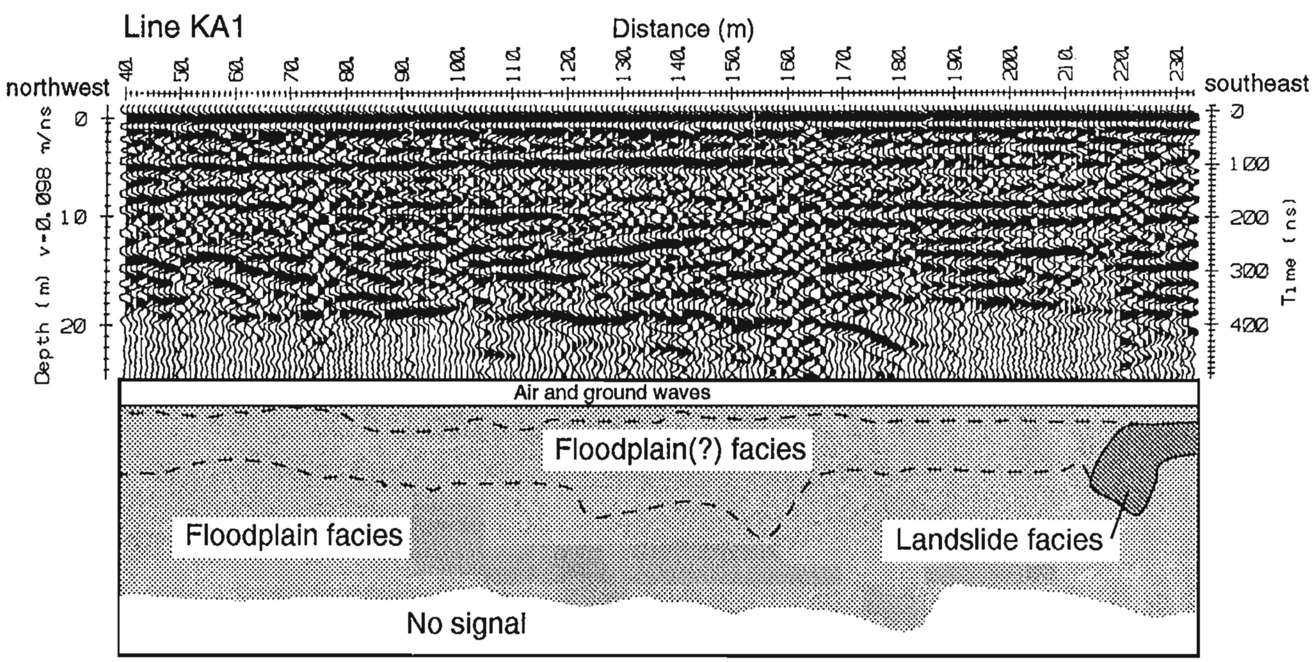

Figure 8. GPR returns of line KA1 between stations 40 and 233 with an interpretative diagram shown below. Note the bed of floodplain(?) facies overlying floodplain facies.

be identified. A thin belt of floodplain facies can be seen to overlie landslide facies in line KA4 between stations 60 and 130 (Fig. 6), a characteristic that is consistent with the surface appearance at the site (Fig. 3).

Interpretation of the GPR profiles, however, is not straightforward everywhere. In line KA2 between stations 380 and 555 (Fig. 7), four separate pockets of poorly defined, side-by-side point reflectors are present and appear to represent landslide facies. Whether all these point sources are in fact large blocks is not clear; some could represent logs or log jams buried within the floodplain. In line KA1 between stations 40 and 215, the upper 8 to 12 m of the returns consist of more or less horizontal, discontinuous reflectors, some of which are very poorly defined (the obvious reflector at 5 m depth may be the water table; Fig. 8). These reflectors do not indicate the presence of large blocks. Conceivably, they could represent returns from floodplain deposits or from a bed of smaller sized landslide debris which may overlie deeper (below 8 to 12 m) floodplain deposits. The presence of the debris field within this part of line KA1 cannot be determined from the GPR returns, thus, the upper 8 to 12 m of the line is shown as floodplain(?) facies in Fig. 8.

The presence of landslide facies has been interpreted in all lines except KA5. Line KA3 is the longest profile (1389 stations) and the only one to completely cross the mapped extent of the debris field (Fig. 4). With respect to the western end of the line, landslide facies is present between stations 150 and 220 and again between stations 260 and 1330 (partially shown in Fig. 5). The latter deposit represents the major occurrence of landslide facies at the site while the former appears to be a small pocket of debris. Lines KA4 and KA6 branch northward directly from and represent extensions of line KA3. Line KA4 consists entirely of landslide facies while in KA6, the facies is only present between stations 0 to 70.

Lines KA1, KA2, and KA7 are all situated within the sawmill yard. Lines KA1 and KA7 both originate from a point in the northwest corner of the yard and extend roughly southeast and southwest, respectively (Fig. 4). Along line KA7, landslide facies is present over the initial 120 stations of the profile. The facies terminates at a point coinciding approximately with the edge of the visible blocks that protrude through the surface of the floodplain. Landslide facies is present in line KA1 between stations 0 to 40 and 215 to 384, but appears absent between stations 40 to 215 (see above).

Line KA2 runs from west to east and roughly parallels the Fraser River, crossing line KA1 at station 171 (Fig. 4). Landslide facies is present within line KA2 between stations 0 and 555, but is discontinuous over the last 100 stations (Fig. 7).

Line KA5 is situated on the western side of the Katz slide site and extends south of line KA3. It consists entirely of floodplain facies and thus lies outside of the debris field, but buried debris could be present below the 15 to 19 m effective depth of GPR signal penetration. Several isolated point source reflectors are found in the returns, but these could be buried logs.

DISCUSSION

The estimated boundary of the older debris field as shown in Fig. 9 is based upon the occurrence of the landslide facies in six of the seven GPR lines and the distribution of blocks protruding through the floodplain. The mapped distribution of the debris field is generally wider than boundaries estimated solely from the presence of protruding blocks, but not significantly. Line KA3 indicates that the debris field occurs continuously over a 1070 m wide section of the valley bottom. Clearly, blocks of the debris field form a major deposit on the valley bottom and are not present in isolated clusters. A gap

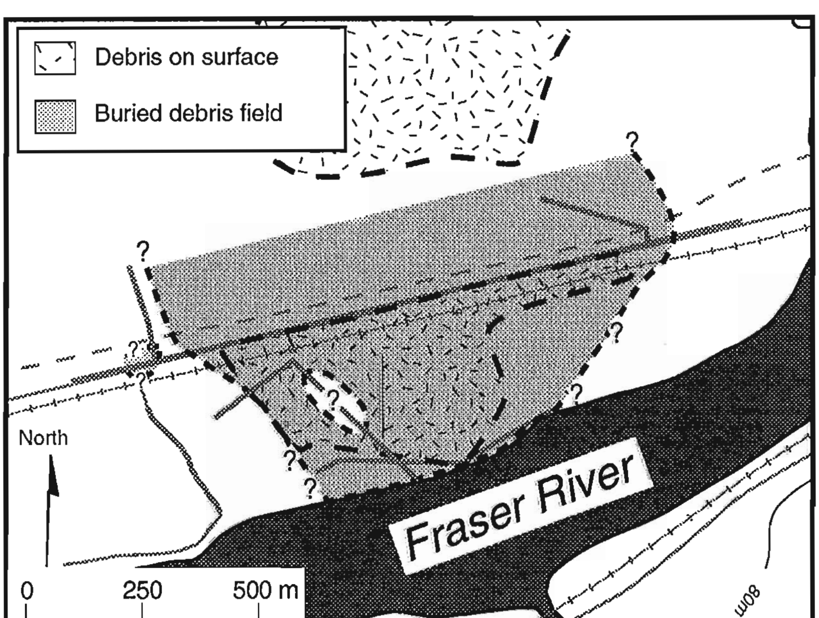

Figure 9.

Map showing the approximate boundary of the buried older debris field in the general area of the GPR survey. This boundary is based upon interpretations of GPR profiles and occurrence of blocks protruding from the floodplain.

in the field, however, appears to be present under the sawmill yard as indicated by line KA1 (Fig. 8, 9). Importantly, the existence of the debris field is confirmed north of Lougheed Highway, but only in the immediate area of line KA4 and at the origin of KA6 adjacent to line KA3. The extent of the deposit north of Lougheed Highway cannot be better defined from our results. To do so would require running additional GPR lines over the northern area of the valley bottom, which could only be done by clearing trails through the dense secondary forest cover.

In Figure 9, the debris field appears to be truncated by the Fraser River channel along the sawmill yard and probably once extended farther across the valley bottom. The original extent of the deposit is unknown. Considering that opposite the Katz slide site the Fraser River is close to the southern valleyside, it seems very likely that the river was at least partially obstructed by the rock avalanche that formed the debris field.

GPR results obtained at the Katz site show that this geophysical technique is very useful for investigating buried, very coarse landslide debris. GPR cross-sections were obtained ranging in depth from 10 m to 22 m. Resolution of the coarse debris with 50 MHz antennas was adequate and generally produced a GPR signature that was distinctive from that of floodplain deposits. Although results adequately delineate buried landslide debris, the GPR equipment was unable to see through the debris. Thus, the thickness of the deposit remains unknown.

CONCLUSIONS

At the Katz site, GPR surveys successfully produced seven profiles delineating the upper 10 to 22 m of the subsurface. Two distinctive GPR facies present in a large portion of the profiles represent landslide and floodplain facies. The landslide facies appears as a chaotic assemblage of discontinuous, hyperbolic reflections produced by large point reflectors (i.e. large buried blocks). Floodplain facies consist of horizontal, laterally continuous reflections that generally are readily distinguishable from the landslide facies.

Based upon the GPR survey, the older of two debris fields at the Katz site forms a major subsurface deposit on the valley bottom that appears to be truncated by the Fraser River. This debris field likely once extended farther across the valley bottom and at least partially obstructed the Fraser River.

ACKNOWLEDGMENTS

We are grateful for the co-operation of the Chawathil and Skawahlook Indian Bands, Sebastion Forest Products, and the Seventh Day Adventist Church in allowing access to their properties. Robert Gerath (Thurber Engineering Ltd., Vancouver) and Rasmus Skovsjold (Emil Anderson Contracting, Hope) provided valuable discussions of the Katz slide site. Sue Pullan reviewed the paper; her comments are appreciated. Field assistance was ably provided by Katherine Bron.

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Geological Survey of Canada projects 830016GB and 920039JP

Cretaceous fossil identifications, Coast Belt, British Columbia

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Haggart, J.W., 1995: Cretaceous fossil identifications, Coast Belt, British Columbia; *in* *Current Research 1995-A; Geological Survey of Canada, p. 41-46.*

Abstract: Cretaceous fossils have been identified in recent collections made from strata in the Mount Waddington (NTS 92N) and Bute Inlet (92K) map areas. In the eastern part of the Coast Belt, ammonites and inoceramid bivalves of Hauterivian age are found in strata assigned to the informal Cloud Drifter formation of Mount Waddington map area. Bivalves of poorly constrained age found in strata correlated with the Taylor Creek Group are assigned a Valanginian to Albian age. Float occurrences of Valanginian fossils from widely spaced localities in the Bute Inlet map area of the western Coast Belt suggest the possible presence of unmetamorphosed, pre-Albian, Cretaceous strata in the region, although significant glacial transport can not be ruled out.

Résumé : Des fossiles du Crétacé ont été identifiés dans de récentes collections provenant de couches dans les régions cartographiques du mont Waddington (SNRC 92N) et de l'inlet Bute (92K). Dans la partie est du Domaine côtier, des ammonites et des inocérames de l'Hauterivien se rencontrent dans des couches de la formation informelle de Cloud Drifter de la région cartographique du mont Waddington. On attribue un âge allant du Valanginien à l'Albien aux bivalves d'âge mal défini trouvés dans des couches correspondant au Groupe de Taylor Creek. Des occurrences dispersées de fossiles du Valanginien de localités très éloignées dans la région cartographique de l'inlet Bute de la partie est du Domaine côtier laissent croire à la présence possible de couches crétacées pré-albiennes, non métamorphisées, dans la région, bien qu'on ne puisse pas écarter un transport glaciaire d'importance.

INTRODUCTION

This report summarizes new age information for fossil collections made by Geological Survey of Canada mapping parties in the Cordillera during the 1993 field season. The report also includes some collections submitted for identification by non-GSC personnel. The collections were made from rocks on both the eastern and western margins of the Coast Belt of British Columbia, in Mount Waddington (92N) and Bute Inlet (92K) map areas (Fig. 1). Ages of fossil collections are based on the biochronological scheme developed for the Cordillera by J.A. Jeletzky (1970), with modification by the author.

Collection locality data are reproduced as reported by the collector, with minor modification where relevant. The format of each reported collection gives the Geological Survey of Canada (GSC) locality number and collector's field number, followed by the relevant geographic locality data for the collection. Stratigraphic assignment is generally that of the collector, either from the original submitted information or from published maps. A brief description of the lithology of the matrix surrounding the fossils, or other lithological information supplied by the collector, is appended after the locality information. A listing of fossils identified at the locality, as well as the age assignment for the collection, completes the fossil determination. The collector's field number includes the collector code: MTB = P. Mustard, Cordilleran Division; UC = Paul Umhoefer, Northern Arizona University.

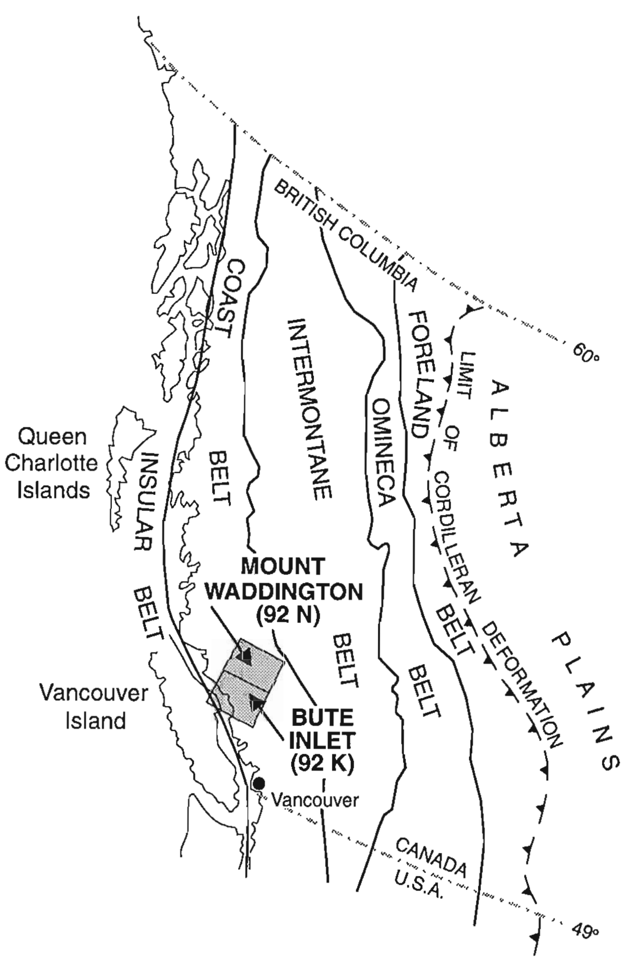

Figure 1. General location map of British Columbia showing areas discussed in text.

IDENTIFICATIONS

Northeast Mount Waddington map area (92N)

Fossil collections made from Mount Waddington map area are concentrated in the northeastern and southeastern parts of the map area, west of the Yalakom Fault. The geology of the northeastern region is discussed in detail by Jeletzky (1968) and Tipper (1969), and more recently by Mustard and van der Heyden (1994). Previous fossil collections from this area have also been described by Jeletzky (1968) and Tipper (1968, 1969). Localities of fossil collections discussed below are plotted on the geological map of Mustard et al. (1994) and form part of the basis for age and unit assignments on that map. Collections are assigned by collectors to either the Cloud Drifter formation of Rusmore and Woodsworth (1989; informal name for map unit 12 of Tipper, 1969) or the Taylor Creek Group.

Cloud Drifter formation

GSC loc.C-101527 (MTB-93-002)
NTS: 92N/15 (Tatla Lake)
UTM: Zone 10, 380623 E; 5741879 N

Lowest outcrop in Sapeye Creek

Massive siltstone

Fossils: *Pleuromya* sp. (*sensu lato*)
Inoceramus cf. *paraketzovi* Efimova (fragments)
Inoceramid prisms
Belemnite, juvenile

Age: A general Hauterivian age is suggested, possibly middle.

GSC loc.C-101528 (MTB-93-004)
NTS: 92N/15 (Tatla Lake)
UTM: Zone 10, 380205 E; 5742116 N

North side of Sapeye Creek; topographically higher than sample C-101527 but many small faults preclude the interpretation of a continuous stratigraphic section

Massive sandy siltstone

Fossils: *Simbirskites (Hollisites)*? sp.

Age: Possibly Hauterivian, middle to late.

GSC loc.C-101529 (MTB-93-006A)
NTS: 92N/15 (Tatla Lake)
UTM: Zone 10, 379649 E; 5742441 N

Sapeye Creek, main canyon about 200 m upstream from falls (= GSC loc.C-176120)

Fine- to medium-grained sandstone

Fossils: *Inoceramus* cf. *paraketzovi* Efimova

Age: A general Hauterivian age is suggested, possibly middle.

GSC loc.C-101532 (MTB-93-030)
NTS: 92N/10 (Razorback Mountain)
UTM: Zone 10, 376874 E; 5733210 N

Main cliffs in creek

Poorly sorted, medium grained sandstone, locally tuffaceous (questionably assigned to Cloud Drifter formation by collector)

Fossils: Ammonite fragment, large, thick-whorled, indeterminate
Nanonavis? sp.
Pectinid bivalves, *Entolium*? sp.
Apiotrigonia cf. *kayana* (Anderson)

Age: Probably Hauterivian-Barremian. Trigoniid bivalves referable to *Apiotrigonia* cf. *kayana* have been described by Poulton (1977) from Hauterivian to Barremian strata of the Manning Park area, British Columbia. *A. kayana* is also known from California (Anderson, 1938), but its stratigraphic level there is uncertain.

GSC loc.C-101534 (MTB-93-047)
NTS: 92N/14 (Bussel Creek)
UTM: Zone 10, 358900 E; 5743404 N

About 30 m above Perkins Peak Mine main Au portal, fossils from outcrop and immediate scree

Siltstone and silty mudstone

Fossils: *Pleuromya* sp.
Thin-shelled bivalves, indeterminate

Age: Indeterminate.

GSC loc.C-101546 (MTB-93-161B)
NTS: 92N/15 (Tatla Lake)
UTM: Zone 10, 379007 E; 5743006 N

Topographically highest outcrop in Sapeye Creek, north side of creek; same lithology as beds lower in creek - appears to be stratigraphically highest part but could be fault repeated

Dark grey massive siltstone

Fossils: *Inoceramus* cf. *paraketzovi* Efimova

Age: A general Hauterivian age is suggested, possibly middle.

GSC loc.C-101547 (MTB-93-162)
NTS: 92N/15 (Tatla Lake)
UTM: Zone 10, 380176 E; 5744207 N

Topographically highest outcrop in east-flowing creek about 1.5 km north of Sapeye Creek

Massive siltstone

Fossils: *Inoceramus* sp. cf. *paraketzovi* Efimova?
Acroteuthis sp.
Belemnite, juvenile, indeterminate
Goniomya sp.
Oxytoma? sp.

Age: A general Hauterivian age is suggested, possibly middle.

GSC loc.C-101548 (MTB-93-171)
NTS: 92N/14 (Bussel Creek)
UTM: Zone 10, 356986 E; 5743290 N

Perkins Peak area, upper cat trail about 1 km southwest of main switchback intersection

Massive metasiltstone

Fossils: *Inoceramus* sp. cf. *paraketzovi* Efimova?
Inoceramid prisms

Age: A general Hauterivian age is suggested, possibly middle.

GSC loc.C-101549 (MTB-93-172)
NTS: 92N/15 (Tatla Lake)
UTM: Zone 10, 368074 E; 5750858 N

Miners Lake fire break road, east end (Klinaklini River)

Massive siltstone

Fossils: *Inoceramus* cf. *paraketzovi* Efimova
Acroteuthis sp
Inoceramid prisms

Age: A general Hauterivian age is suggested, possibly middle.

GSC loc.C-176112 (MTB-93-255)
NTS: 92N/15 (Tatla Lake)
UTM: Zone 10, 370463 E; 5747095 N

Southwest facing slope on north side of east-west-trending gully which runs east into main creek (gully trends toward green repeater hill)

Massive siltstone

Fossils: *Inoceramus* cf. *paraketzovi* Efimova

Age: A general Hauterivian age is suggested, possibly middle.

GSC loc.C-176120 (MTB-93-006B)
NTS: 92N/15 (Tatla Lake)
UTM: Zone 10, 379649 E; 5742441 N

Sapeye Creek, main canyon about 200 m upstream from falls (= GSC loc.C-101529)

Fine- to medium-grained sandstone

Fossils: *Simbirskites (Hollisites)* sp.
Simbirskites sensu stricto

Age: Hauterivian, middle to late.

The best constrained of these collections from the Cloud Drifter formation include ammonites and bivalves of Hauterivian age, consistent with the earlier interpretation of Jeletzky (1968) for the succession of this region.

Taylor Creek Group

Several collections made from exposures just west of Bluff Lake (see Mustard et al., 1994) have been assigned to the Taylor Creek Group, although a precise age assessment for these strata is presently lacking.

GSC loc.C-101535 (MTB-93-060B)
NTS: 92N/15 (Tatla Lake)
UTM: Zone 10, 376145 E; 5734964 N

Clay Creek, lower part

Marine sandstone/mudstone turbidites

Fossils: *Pholadomya distorta* Anderson
Pholadomya russelli Anderson?
Bivalves, juvenile, indeterminate

Age: Probably Valanginian to Albian. The holotype of *Pholadomya distorta* is known from the upper Paskenta Group (upper Valanginian) of northern California, and possibly from the upper Albian as well (Anderson, 1938). *P. russelli*, with which one specimen in the present collection is questionably compared, is also known from the upper Albian of northern California.

The precise stratigraphic assignment of various pholadomyids from northern California has not been established and several of the forms defined as distinct species by Anderson may simply reflect intraspecific variation. However, it seems readily apparent that a zone of abundant pholadomyids succeeds strata with common buchiids in California, and that they characterize the uppermost Valanginian through Albian interval there; it is considered likely that a similar relationship exists in British Columbia.

GSC loc.C-101537 (MTB-93-065)
NTS: 92N/15 (Tatla Lake)
UTM: Zone 10, 376983 E; 5734877 N

West tributary of Cherry Creek, small canyon about 200 m long

Fossils: *Pholadomya* cf. *altumbonata* Anderson
Nemodon breweriana (Gabb)?
Pectinids, *Entolium*? sp.
Crioceratiform ammonite, juvenile, indeterminate

Age: Probably latest Valanginian to Albian - see comments above for collection C-101535.

GSC loc.C-101538 (MTB-93-066)
NTS: 92N/15 (Tatla Lake)
UTM: Zone 10, 377514 E; 5734719 N

West tributary of Cherry Creek, cliffs on both sides of creek

Fossils: *Pholadomya distorta* Anderson?
Pholadomya russelli Anderson
Possible crioceratid ammonite fragment?
Wood fragments

Age: Probably latest Valanginian to Albian - see comments above for collection C-101535.

The numerous bivalves present in these three collections assigned to the Taylor Creek Group were preliminarily compared by the author to the inoceramid genus *Birostrina*, of Albian age (Haggart in Mustard and van der Heyden, 1994; Mustard et al., 1994). This tentative age interpretation formed the basis for assignment of the stratigraphic unit in which they occur to the Taylor Creek Group. Detailed study of the bivalves in all three collections has subsequently shown, however, that they are actually pholadomyids, rather than inoceramids, with a more poorly constrained age range than *Birostrina*, Valanginian to Albian. Although this age range is somewhat broad, encompassing the time of deposition of the Cloud Drifter formation as well as the Taylor Creek Group, the lack of pholadomyids in all the Hauterivian-age Cloud Drifter collections described above suggests that they represent a different temporal level than that unit. In addition, such pholadomyids are uncommon or rare in Valanginian strata studied by the author elsewhere in British Columbia, rocks which are characterized by abundant buchiid bivalves. For these reasons the three collections tentatively assigned to the Taylor Creek Group are considered as likely of post-Hauterivian age. The collector also reports that subsequent detailed petrographic examination of sandstones from this unit indicate it has a different composition and provenance than sandstones of the Cloud Drifter formation and also contains sedimentary chert, a distinctive component of Taylor Creek Group sandstones in areas to the east (P. Mustard, pers. comm., 1994).

Southeast Mount Waddington map area (92N)

Two collections were submitted from the southeastern part of Mount Waddington map area. The stratigraphic assignment of the strata bearing these collections is problematic. They were originally assigned by the collector to either the Silverquick Conglomerate, a nonmarine conglomerate, sandstone, and siltstone sequence which is well-developed to the southeast (Garver et al., 1989; Mahoney et al., 1992), or to a 'nonmarine' facies of the Cloud Drifter formation. The sequence includes intercalated strata interpreted as nonmarine, marginal marine, and shallow marine deposits (P. Umhoefer, pers. comm., 1994).

GSC loc.C-202605 (UC-93-37)
NTS: 92N/1 (Chilko Mountain)
UTM: Zone 10, 416250 E; 5677900 N

South-central Tredcroft Creek

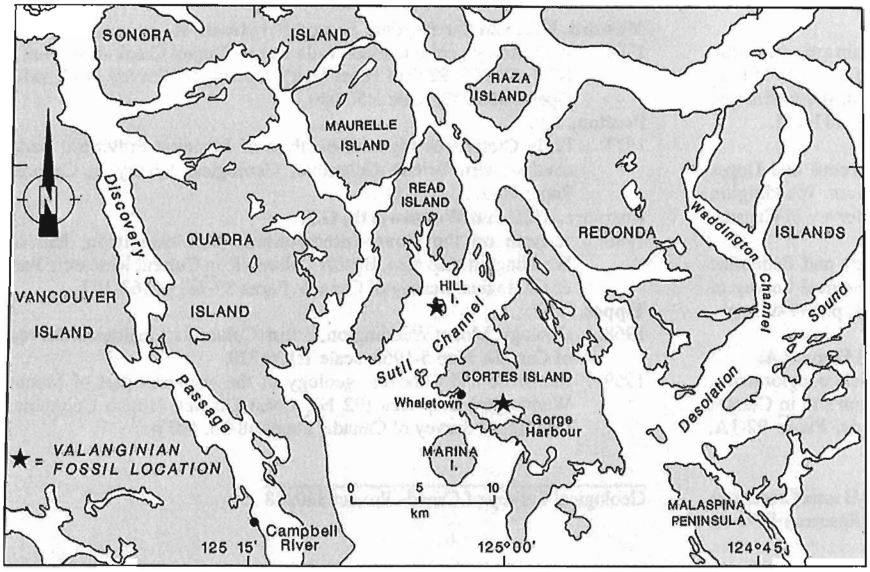

Figure 2.

Map of southern part of Bute Inlet map area showing localities of float occurrences of Cretaceous strata.

Shell-rich, calcareous fine grained sandstone to siltstone

Fossils: *Pleuromya*? sp. (*sensu lato*), juvenile
Abundant small bivalves, several taxa of venerids represented

Age Indeterminate.

GSC loc.C-202607 (UC-93-57)
NTS: 92N/8 (Stikelan Creek)
UTM: Zone 10, 412200 E; 5682900 N

Mount Edwards

Calcareous, fine grained silty mudstone

Fossils: Abundant, small gastropods, indeterminate
Small bivalves, indeterminate, including possible pteriomorphs

Age: Indeterminate.

Both collections consist of a hash of abundant small gastropods and bivalves, including moulds, fragments, and a few whole shells. Some specimens exhibit relatively thick, calcareous shell material adhering to the moulds. The mollusc assemblages, coupled with the presence of common calcareous shell material, suggest a shallow-marine, rather than nonmarine depositional environment. A shallow-water, high-energy, setting is postulated, possibly in tidal channels.

Bute Inlet map area (NTS 92K)

NTS: 92 K/3 (Quadra Island)
UTM: Zone 10, 355500E, 5551700N (approximate)

Cortes Island, north shore of Gorge Harbour, vicinity of the Government Dock, collected by C. Gronau from float among beach gravels

Fine- to medium-grained sandstone

Fossils: *Buchia* sp. cf. *keyserlingi* (Lahusen) - specimens returned at collector's request

Age: Likely Late Valanginian. The specimen shows no evidence of metamorphism.

The identification of *Buchia* sp. cf. *keyserlingi* in float at Gorge Harbour on Cortes Island (Fig. 2) is particularly interesting as no sedimentary strata of Valanginian age are known from this region of the western Coast Belt. Another float occurrence of buchiid fossils in this region (of Middle Valanginian age: *B. pacifica* Jeletzky) is found at Hill Island, approximately 8 km northwest of Gorge Harbour (G.J. Woodsworth, pers. comm., 1993). At that locality, as well, no outcropping Cretaceous strata are known. The fossils at both sites must be either derived locally, from unrecognized Lower Cretaceous rocks in the region, or transported to their site of collection by glacial ice. If the latter, the source is still problematic, as unmetamorphosed Valanginian strata are unknown in the western or central Coast Belt to the east of the fossil localities. Valanginian-age strata are found in the eastern Coast Belt, however the distance to these localities, as well as the direction of glacial transport, suggests the likelihood of their being the source of the collections is unlikely.

ACKNOWLEDGMENTS

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Geological Survey of Canada Project 880038

Early historical and ethnographical accounts of large earthquakes and tsunamis on western Vancouver Island, British Columbia

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Clague, J.J., 1995: Early historical and ethnographical accounts of large earthquakes and tsunamis on western Vancouver Island, British Columbia; <u>*in*</u> *Current Research 1995-A; Geological Survey of Canada, p. 47-50.*

Abstract: Abundant geological evidence has been found for one or more large earthquakes on the Cascadia subduction zone about 300 years ago. The earthquakes produced severe shaking, crustal subsidence, and large tsunamis along the Pacific coast from Vancouver Island to northern California and must have profoundly affected the people living in these areas. Northwest Coast Indian oral traditions record, albeit in an exaggerated way, tsunamis triggered by these rare, plate-boundary events. The oldest known historical earthquake in British Columbia, in February 1793, was recorded by Spanish explorers wintering at Nootka Sound on Vancouver Island and may have occurred at shallow depth in the crust or deeper, within the subducting Juan de Fuca plate.

Résumé : De nombreux indices géologiques ont été découverts d'un ou de plusieurs séismes d'importance qui ont secoué la zone de subduction de Cascadie il y a environ 300 ans. Les séismes ont provoqué de fortes secousses, une subsidence crustale et de gigantesques tsunamis le long de la côte du Pacifique, depuis l'île de Vancouver jusqu'au nord de la Californie. Ils ont dû affecter grandement les habitants de ces régions. Les traditions orales des Amérindiens de la côte du Nord-Ouest font état, quoiqu'avec des exagérations, de tsunamis provoqués par ces événements rares, se produisant à la frontière de plaques. Le plus ancien séisme connu de la période historique en Colombie- Britannique a eu lieu en février 1793 et a été consigné par des explorateurs espagnols passant l'hiver dans le détroit de Nootka, dans l'île de Vancouver. Il est possible que ce séisme se soit produit à une faible profondeur dans la croûte ou encore plus profondément, au sein de la plaque Juan de Fuca en subduction.

INTRODUCTION

Geological evidence has recently been found on the west coast of Vancouver Island for a large earthquake and tsunami less than 400 years ago (Clague and Bobrowsky, 1994). In numerous excavations and tidal bank exposures near Tofino and Ucluelet, a peaty marsh soil is abruptly overlain by a sand sheet, inferred to have been deposited by a tsunami. Foraminifera and vascular plant fossils show that the buried soil was submerged suddenly and was quickly covered by sand (Clague and Bobrowsky, 1994). Clague and Bobrowsky (1994) attributed the submergence to sudden subsidence of the crust during the most recent large earthquake on the northern part of the Cascadia subduction zone. The sand sheet overlying the peaty soil records a tsunami triggered by this earthquake. Similar stratigraphic sequences of about the same age have been reported from numerous estuaries along the outer coasts of Washington and northern Oregon (Atwater, 1987, 1992; Darienzo and Peterson, 1990; Nelson, 1992; Nelson and Personius, in press; Peterson and Darienzo, in press), suggesting that hundreds of kilometres of the subduction zone may have ruptured during one, or a series of great ($M_w \geq 8$) plate-boundary earthquakes less than 400 years ago.

A great earthquake on the Cascadia subduction zone would have profoundly affected the indigenous peoples of southwestern British Columbia. In particular, Indians living on the west coast of Vancouver Island would have experienced severe ground shaking, crustal subsidence, and a powerful tsunami, all on a scale beyond their normal experience. It thus is worth asking the question: are there oral traditions of great earthquakes and tsunamis among the native people of western Vancouver Island? In addition, is there any mention of large earthquakes in the journals of early European explorers in the region?

THE EARTHQUAKE OF FEBRUARY 1793

The historical period in British Columbia is brief; it was only about 200 years ago that Europeans "discovered" what is now southwestern British Columbia, and the first detailed written

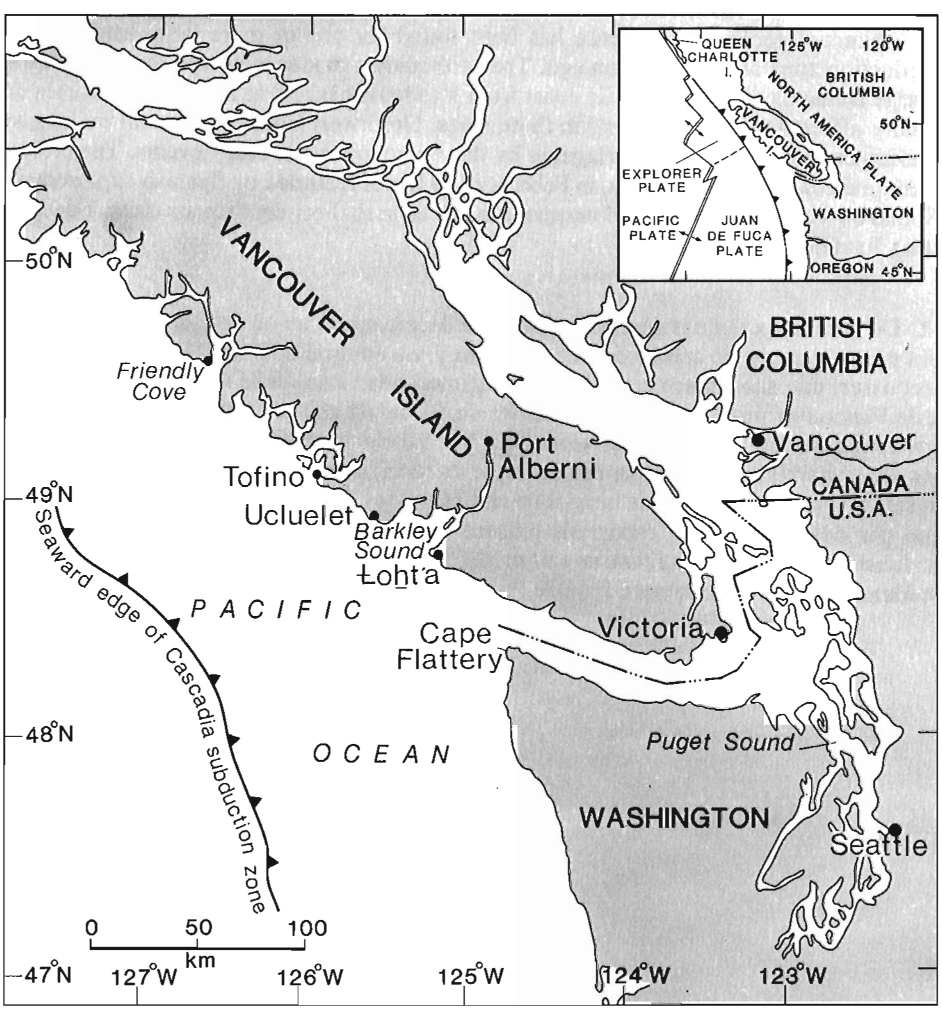

Figure 1. Map showing locations of sites mentioned in the text. Inset map shows lithospheric plates.

records date back only to the middle of the nineteenth century. The first report of an earthquake in southwestern British Columbia comes from Spanish explorers who wintered at Friendly Cove in Nootka Sound (Fig. 1) in 1793. Salvador Fidalgo, the commandant of the Spanish settlement at Friendly Cove, complained to visiting Captain George Vancouver "of having passed a most irksome winter, characterized by incessant rain, punctuated in February by a severe earthquake..." (Gillespie, 1992, p. 159). From this brief account, it is not possible to determine the size or source of the earthquake. However, there have been three, moderate to large, crustal earthquakes on central Vancouver Island during the twentieth century (in 1918, 1946, and 1957), and the 1793 earthquake may have a similar source.

ORAL TRADITIONS

Native oral traditions of tsunamis provide evidence for recent, large earthquakes on the Cascadia subduction zone. An example is the Clayoquot story of a great ebb and flow of the sea (Sproat, 1987, p. 124-125):

> "Generations ago, the Sheshaht, who live now during part of the year in Barkley Sound and the remainder of the year at Alberni, were unacquainted with the head of the inlet. They had two houses on the sound, and used to migrate from one to the other. At that time a most curious phenomenon of nature occurred. The tide ebbed away from the shores of the sound and left it dry, and the sea itself retreated a long distance. This continued for four days, and the Sheshaht made light of the occurrence. There was one, however, Wispohahp, who, with his two brothers, did not do so. After a mature consideration of the circumstance, he thought it likely that this ebb would be succeeded by a floodtide of corresponding height and power. Accordingly, he and his brothers spent three days in the forest collecting material for a rope of cedar inner bark, which, when made, was so large as to fill four boxes. There was a rock near the village, from the base of which sprang a group of bushes, of a sort well known for its toughness. Round these bushes Wispohahp fastened one end of his rope, attaching the other to his canoe. In his canoe were placed all his moveables, his wife, his two brothers, and their wives; and thus prepared they waited for the result. After four days the tide began to flow, and crept slowly up to about halfway between the point of its furthest ebb and the houses. At this point, its pace was suddenly quickened, and it rushed up at fearful speed. The Sheshaht ran to their canoes. Some begged to be attached to Wispohahp's rope; but to this he would not consent, lest it should be broken. Others would have given him, several of their women, but he would not receive them. They were all soon caught be the rising water; and while Wispohahp rode safely at anchor, the Sheshaht, unable to resist its force, drifted in their canoes to distant parts. Finally, the water covered the whole country, except Quossakt, a high mountain near the Toquaht, and Mount Arrowsmith (Cush-cu-chuhl). The Toquaht, another tribe living near the Sheshaht, got into a large canoe (*Eher Kleetsoolh*), and paddled to the summit of Quossakt, where they landed. At the end of four days, the floodtide began to abate..."

A similar story was told to Judge James Swan, a well known student of native lore, in 1868 by a chief of Indians living at Cape Flattery (Fig. 1):

> "...The water suddenly receded leaving Neeah Bay perfectly dry. It was four days reaching its lowest ebb, and then rose again without any waves or breakers, till it had submerged the Cape, and in fact the whole country, excepting the tops of the mountains at Clayoquot. The water on its rise became very warm, and as it came up to the houses, those who had canoes put their effects in them, and floated off with the current, which set very strongly to the north. Some drifted one way, some another; and when the waters assumed their accustomed level, a portion of the tribe found themselves beyond Nootka, where their descendants now reside...Many canoes came down in the trees and were destroyed, and numerous lives were lost. The water was four days regaining its accustomed level." (Swan, 1868; quoted in Heaton and Snavely, 1985, p. 1457).

In their comments on this account, Heaton and Snavely (1985) noted that no tsunami could have the height (ca. 400 m) or duration (four days) of the event reported in the Cape Flattery legend. They thus questioned whether the account might be entirely fictional. However, the description of the water receding and then returning in a strong current, both at Cape Flattery and Port Alberni, is suggestive of a tsunami. Thus, the event probably happened, although it is greatly exaggerated in the oral tradition. In this regard, Swan separated the Cape Flattery account from the majority of legends that he felt were of a more mythical nature. He reported that the event occurred "a long time ago but not at a very remote period"; it thus predated the time of the Indian who reported it to him, but was not of an ancient, mythical era.

There is nothing in the above accounts to indicate the source of the presumed tsunami(s). However, the following account suggests that ground shaking may have accompanied a tsunami that destroyed Lo<u>h</u>t'a (Fig. 1), the principal winter village of the Pachena Bay people, sometime prior to European contact. If so, this tsunami must have been generated by an earthquake off Vancouver Island or Washington, as opposed to a distant, large earthquake elsewhere in the North Pacific Ocean.

> "This story is about the first !Anaql'a or "Pachena Bay" people. It is said that they were a big band at the time of him whose name was <u>H</u>ayoqwis'is, 'Ten-On-Head-On-Beach.' He was the Chief; he was of the Pachena Bay tribe; he owned the Pachena Bay country. Their village site was Lo<u>h</u>t'a; they of Lo<u>h</u>t'a live there. I think they numbered over a hundred persons...There is no one left alive due to what this land does at times. They had no way or time to try to save themselves. I think it was at nighttime that the land shook...They were at Lo<u>h</u>t'a:, and they simply had no time to get hold of canoes, no time to get awake. They sank at once, were all drowned; not one survived...I think a big wave smashed into the

beach. The Pachena Bay people were lost...But they on their part who lived at Ma:lts'a:s, 'House-Up-Against-Hill', the wave did not reach because they were on high ground. Right against a cliff were the houses on high ground at M'a:lsit, 'Coldwater Pool'. Because of that they came out alive. They did not drift out to sea along with the others..." (Arima et al., 1991, p. 230-231).

In conclusion, some of the oral traditions of Northwest Coast Indians seem to record ground shaking and tsunamis that have no precedence in the historical period. These events are probably the result of great earthquakes on the Cascadia subduction zone.

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Geological Survey of Canada Project 870017

Preliminary studies of hydrothermal alteration events at the Island Copper deposit, northern Vancouver Island, British Columbia

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Leitch, C.H.B., Ross, K.V., Fleming, J.A., and Dawson, K.M., 1995: Preliminary studies of hydrothermal alteration events at the Island Copper deposit, northern Vancouver Island, British Columbia; in Current Research 1995-A; Geological Survey of Canada, p. 51-59.

Abstract: Core logging/pit mapping suggest three main stages of intrusion (early, intermediate, late), defined by differences in alteration intensity. Hydrothermal events comprise early biotite-magnetite "hornfels", followed by magnetite-actinolite-plagioclase veining, and then quartz-chalcopyrite veins and fractures with or without "hydrothermal" biotite envelopes. Chlorite-sericite ± clay are likely retrograde overprints as the system cooled and collapsed inwards; epidote may be prograde peripheral, retrograde, or both. Chalcopyrite was introduced in several stages: minor with magnetite-biotite, followed by main-stage disseminations/fracture fills accompanying quartz veins, some with biotite envelopes, and finally with epidote-chlorite and pyrite-chalcopyrite filled fractures. Fluid inclusions in quartz-magnetite veins are highly saline (multiple daughter products); in quartz-chalcopyrite veins saline (halite only); in late veins or reopenings associated with epidote less saline (liquid/vapour only).

Résumé : De nombreux indices géologiques ont été découverts d'un ou de plusieurs séismes d'importance qui ont secoué la zone de subduction de Cascadie il y a environ 300 ans. Les séismes ont provoqué de fortes secousses, une subsidence crustale et de gigantesques tsunamis le long de la côte du Pacifique, depuis l'île de Vancouver jusqu'au nord de la Californie. Ils ont dû affecter grandement les habitants de ces régions. Les traditions orales des Amérindiens de la côte du Nord-Ouest font état, quoiqu'avec des exagérations, de tsunamis provoqués par ces événements rares, se produisant à la frontière de plaques. Le plus ancien séisme connu de la période historique en Colombie-Britannique a eu lieu en février 1793 et a été consigné par des explorateurs espagnols passant l'hiver dans le détroit de Nootka, dans l'île de Vancouver. Il est possible que ce séisme se soit produit à une faible profondeur dans la croûte ou encore plus profondément, au sein de la plaque Juan de Fuca en subduction.

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INTRODUCTION

Island Copper is an island-arc type porphyry Cu-Mo-Au deposit (Perelló et al., 1989; Arancibia and Clark, 1990) operated by BHP Minerals Ltd. on northern Vancouver Island near Port Hardy, British Columbia. It resulted from the intrusion of a series of dyke-like bodies of rhyodacitic quartz-feldspar porphyry of about 180 Ma age into possibly comagmatic high-alumina basalts, basaltic andesites, minor rhyolites and pyroclastic rocks of the Middle Jurassic Bonanza Group (Northcote and Robinson, 1973; Muller, 1977; Nixon et al., 1994). The size of the deposit was initially estimated at 257 million tons of ore at 0.52% Cu and 0.017% Mo (Cargill et al., 1976); the final mined plan envisages a total of 377 million tons at 0.41% Cu and 0.017% Mo at a 0.2% Cu cutoff (Perelló et al., in press). Gold has been produced at an annual rate of 1200-1500 kg (40-50 000 oz) from a head grade of 0.19 g/t. Only about 50% of the gold is recovered in the copper concentrate, which averages 24% Cu, 7 g/t Au and also contains 60 g/t Ag. The molybdenum concentrate contains up to 1400 ppm rhenium, making Island Copper Canada's only producer of this element (Perelló et al., in press).

A joint project with the Mineral Resources Division of the Geological Survey of Canada (GSC-MRD), the British Columbia Geological Survey Branch (BCGSB), BHP Minerals Ltd., and Auckland University (New Zealand) has been initiated to study the deeper levels of the deposit before mine closure and pit flooding, anticipated in 1995 or 1996. Project members involved include: Craig Leitch, Katherina Ross, Ken Dawson, Rod Kirkham, Colin Dunn, and Mel Best of the GSC; Graham Nixon, Jan Hammack, Andre Panteleyev, Victor Koyanagi, Steve Sibbick and Peter Bobrowsky of the BCGSB; John Fleming of BHP Minerals; and Stuart Simmons and Geraint Mathias of Auckland University. The objectives of the GSC-MRD team are to study the geology and alteration of the deep exposures and drill core, to better understand the sequence of intrusive, alteration, and mineralizing events at all levels in the deposit. Methods include U-Pb zircon geochronology, fluid inclusion and stable isotope studies, and lithogeochemistry. Other related studies include biogeochemistry over and around the deposit by Colin Dunn of GSC-MRD and depth of overburden using geophysical methods by Mel Best of GSC-Geophysics and Marine Geoscience Branch. The BCGSB team is updating knowledge of the regional geology, mineral deposits, geochronology, and geochemistry of the surrounding part of northern Vancouver Island. A detailed study of the high-level advanced argillic alteration is the focus of the Auckland University members.

PREVIOUS WORK

Since the beginning of production in 1971, the Island Copper deposit has been the subject of numerous geological studies. Brief summaries of the geology were published by Young and Rugg (1971) and Northcote and Robinson (1973). A more detailed description was published by Cargill et al. (1976), and an updated review of the geology was presented by Fleming (1983). Several theses, including those of Cargill (1975), Fahey (1979), and Perelló (1987), have been completed on the deposit; a PhD study by Arancibia, begun in 1977, remains unfinished. The most recent publications include a comprehensive review by Perelló et al. (in press) and partial results of the Arancibia thesis work (Arancibia and Clark, 1990 and in press).

TIMING OF EVENTS

Porphyry intrusions

All phases of the porphyry are texturally and mineralogically similar, and probably of rhyodacite composition (dacite to rhyolite: O.N. Arancibia, in Perelló, 1987). Unaltered porphyry (O.N. Arancibia, unpublished data; Leitch, unpublished data) consists of approximately 20-30% coarse (0.5-1 cm) bipyramidal quartz phenocrysts, 15-30% 2-5 mm plagioclase laths and <5% chloritized biotite books to 2 mm set in a fine (10 to 30 μm) matrix of quartz and K-feldspar or albite. The plagioclase phenocrysts are oscillatory zoned oligoclase-andesine (An_{30-40}) from rim to core (Leitch, unpublished data).

Distribution of the porphyry phases is shown for section 155 through the mine in Figure 1. Three main intrusive events of quartz-feldspar porphyry are recognized in the present study (cf. Perelló, 1987; Perelló et al., in press; Arancibia and Clark, in press), based on differences in alteration/veining intensity, crosscutting relations and included fragments:

Early phase, characterized by intense magnetite-quartz to quartz-magnetite stockwork and/or flooding by disseminated magnetite, frequently leading to total destruction of texture. Quartz-magnetite veins appear to form a continuum with quartz-pyrite±chalcopyrite-molybdenite-magnetite veins. Increasingly quartz-rich veins crosscut magnetite-quartz veins.

Intermediate phase, cut only by rare quartz-magnetite veins and magnetite fractures, and characterized by a general abundance of planar grey quartz-pyrite±chalcopyrite±molybdenite veins. This phase rarely contains clasts of intensely magnetite-quartz stockworked porphyry. It is generally less intensely altered by clay-sericite-chlorite than the early phase, but the contacts are not always clear, as the porphyries are texturally almost identical. Distinction between the phases is based on the abrupt disappearance of intense quartz-magnetite alteration (including the truncation of veins).

Late phase, completely lacking quartz-magnetite stockwork. This phase contains only minor amounts of disseminated magnetite, and clearly truncates quartz-magnetite veining in the early phase (Fig. 2a, b). It is also observed to cut marginal breccia (see below) developed around the early porphyry. Chalcopyrite is restricted to altered mafic mineral sites. Rare quartz±pyrite veins and crosscutting molybdenite on slips occur. Contacts are fresh, sharp and slightly chilled. Inclusions of earlier intrusive phases are common, but crosscutting relations with intermediate porphyry have not been observed.

Breccias

A "marginal breccia" unit mapped around the margins of the early intrusive at Island Copper by previous authors (e.g. Perelló, 1987) appears to be an inclusive term for several breccia types. These range from crackled and hydrothermally veined or stockworked porphyry (unrotated blocks) to heterolithic breccias including volcanic and porphyry clasts (transported blocks) to hydrothermal breccias composed of rounded, intensely altered clasts (highly milled blocks). Matrix to the breccia is difficult to resolve pending petrographic study, but may include some igneous material in addition to the dominant rock flour (cf. Sillitoe, 1989; Perelló et al., 1989). The milled breccia contains clasts of white quartz, dark magnetite-quartz±hematite, and clay-sericite-pyrite altered rock in a matrix of siderite-quartz-hematite-pyrite±chalcopyrite and rare bornite-chalcocite-covellite.

In exposures of marginal breccia, quartz-magnetite veins are cut off in some clasts but cut through other clasts (cf. Padilla-Garza, 1993), indicating several stages of brecciation that possibly overlap the transition from intrusive breccia to hydrothermal breccia associated with the early porphyry. The marginal breccia was not observed to be associated with the intermediate and late porphyries, although Perelló (1987) stated that some breccias post-date the main mineralizing stage.

An extensive area at the west end of the pit is underlain by what has been termed "pyrophyllite-dumortierite breccia"; it has been reported to include fragments of intermediate porphyry and to be cut by late porphyry (Perelló, 1987). Examination of available drill core and pit exposures of this unit, however, suggest an origin by pyrophyllite-dumortierite alteration of a fragmental volcanic rock or intrusion breccia (or both). The location of this breccia adjacent to and transitional

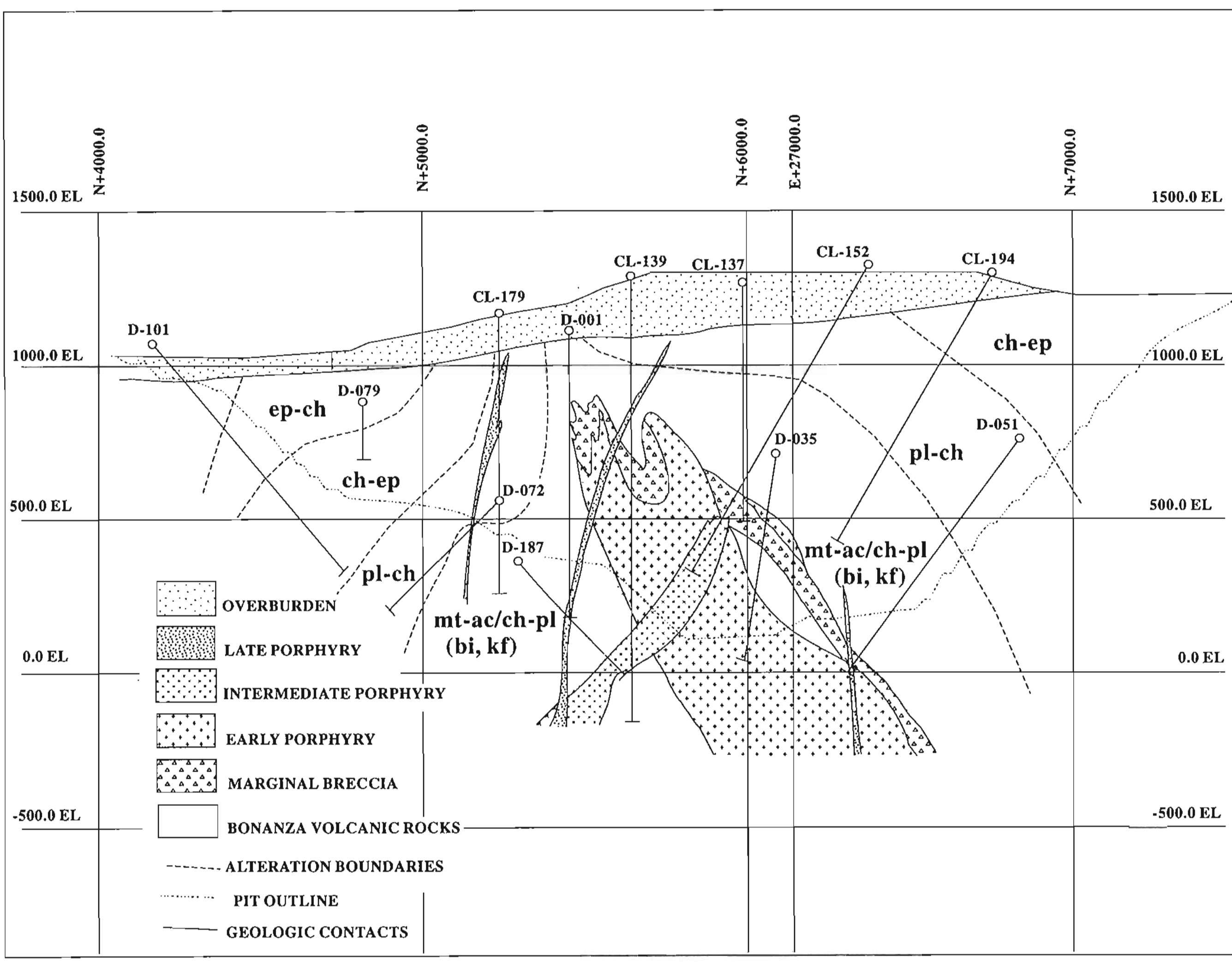


Figure 1. *Cross-section 155W at about 27000 E through the east-central part of the Island Copper deposit to illustrate interpreted geology and alteration zoning. All co-ordinates are in feet (vertical scale same as horizontal scale); mine elevations, used for pit bench designation, are relative to a zero at 1000' below sea level. Abbreviations: ac = actinolite, bi = biotite, ch = chlorite, ep = epidote, kf = K-feldspar, pl = plagioclase, mt = magnetite. Drill hole collars are shown by open circles and identified by number.*

to marginal breccia indicates the possibility that it is an early breccia that has later undergone intense advanced argillic alteration.

Pebble dykes

Late pebble dykes, rarely observed in the pit, generally trend about 325° across the middle of the deposit. These dykes are up to 0.5 m thick and subvertical, and roughly parallel a minor porphyry dyke trend seen in the pit. They have sharp, commonly faulted or sheared contacts. Other examples are observed in drill core, with apparent widths (probably exaggerated in these steep holes) up to several metres. Pebble dykes are divisible into two types: a possibly slightly earlier type that is green, sericitic and strongly pyritic (up to 10% pyrite in the matrix), and a later pink, crumbly and unpyritic type that grades into calcite-zeolite rich fractured or crushed zones. The dykes contain clasts of highly altered and mineralized early porphyry, quartz-magnetite±chalcopyrite veins, and rare clasts of later porphyry in an aplitic-looking matrix (Fig. 2c). The age distinction between the two types is based on the greater degree of alteration, pyritization and lithification of the green type, plus the relation of the pink type to fractured zones. The green type is similar in appearance to some exposures of marginal breccia, implying it may not be much later. Variably pyritic, clay-rich gouge zones are abundant and may in places superficially resemble the pebble dykes, but generally are distinguished by the presence of a less "igneous-looking" matrix (petrography is required to resolve the pebble dyke matrix).

a) Contact (sheared) of intensely magnetite stockworked and flooded early porphyry (dark grey, texture destroyed) with late, pink porphyry (grey, unveined except by calcite-zeolite). Hole E92 at 781 feet, from Bay Lake zone west of the pit.

b) Contact of late porphyry (grey, rare white calcite-zeolite veins) with early porphyry (darker grey, intensely magnetite stockworked) at 469.8' and then of early porphyry with intensely magnetite-actinolite-chlorite altered Bonanza volcanics (black, cut by white calcite-zeolite fractures and grey quartz-sericite-pyrite-clay envelopes). Hole D187, section 155W, centre of pit at 0 level (Fig. 1)

c) Pebble dyke (pink type) containing fragments of quartz vein, magnetite flooded early porphyry, and less altered intermediate or late porphyry in an aplitic matrix (southeast wall of pit, 120 level).

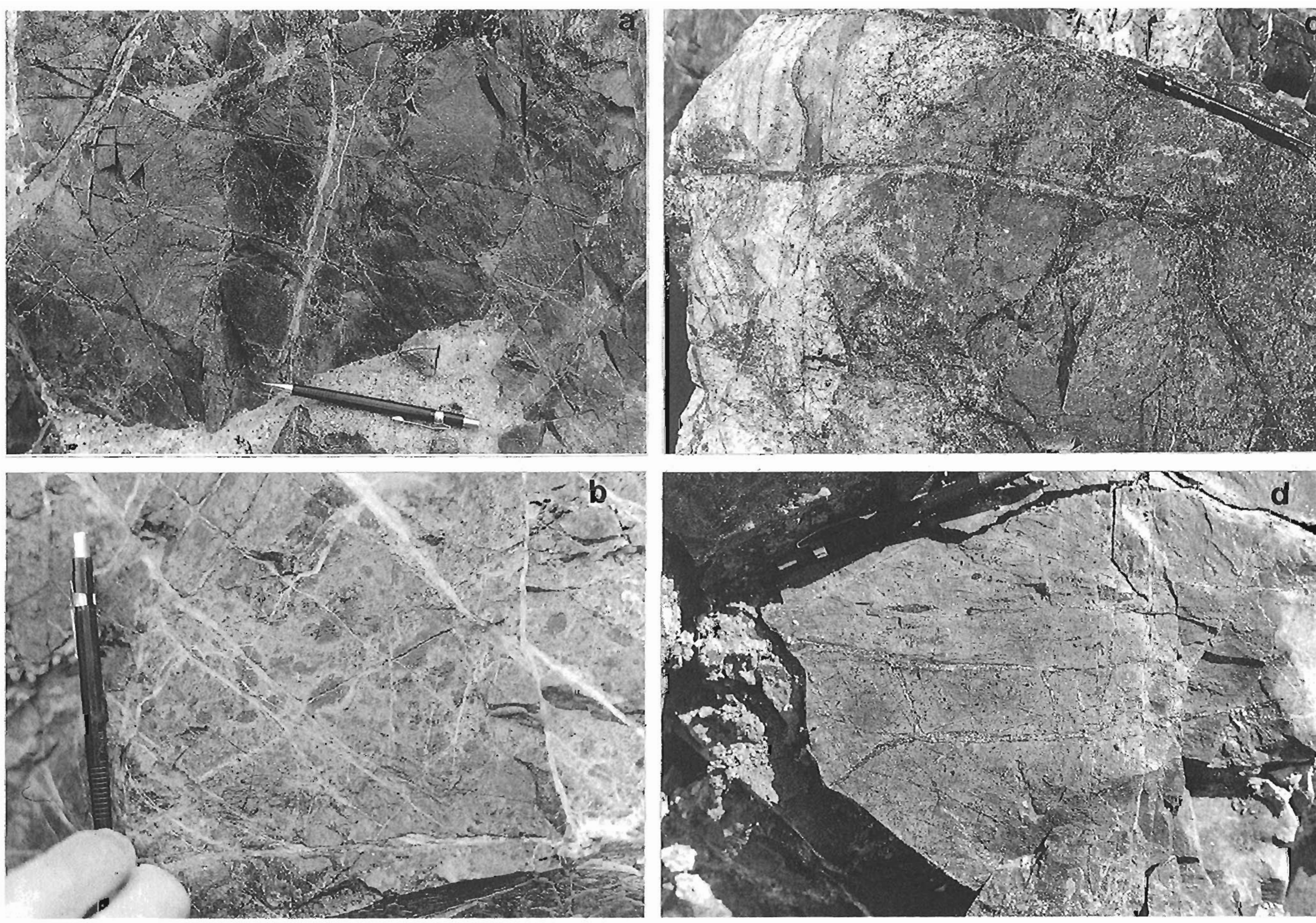
***Figure 2.** Contact relations of intrusive phases at Island Copper.*

Alteration

A concentric pattern of alteration assemblages developed within the Bonanza volcanic rocks and centred on the porphyritic intrusions, has been recognized by previous authors (Cargill et al., 1976; Fleming, 1983; Perelló et al, in press; Arancibia and Clark, in press). However, the temporal relationships between porphyry intrusion, alteration and mineralization are not yet completely understood. The main alteration assemblages within the Bonanza volcanic rocks recognized in this study are, from innermost to outermost (Fig. 1): magnetite-actinolite/chlorite-plagioclase±biotite ±K-feldspar (Fig. 3a); plagioclase-chlorite (Fig. 3b); and chlorite-epidote. The distinction between actinolite and chlorite is almost impossible to make in hand specimen; in many places both may be present. Biotite appears to be partly relict in the inner two assemblages and partly late (see below). The alteration feldspar is generally albite but ranges from -

oligoclase to locally andesine; K-feldspar, likely orthoclase, is also found with increasing alteration intensity closer to the centre of the system or inward in a single fracture envelope (Arancibia and Clark, 1990; Leitch, unpub. data; cf. Leitch, 1981). Pyrite is found throughout all zones. All these alteration types are cut by later, generally structurally controlled, quartz-sericite-clay- pyrite and pyrophyllite-dumortierite alteration assemblages. Plagioclase±chlorite and quartz-sericite-pyrite±clay alteration assemblages are intensely developed locally in both porphyritic intrusions and volcanic rocks. Silicification and magnetite alteration are also moderately to locally intensely developed in the porphyritic intrusive rocks and breccias (Fig. 3c). The distribution of altered intrusive rocks is too variable to show in Figure 1.

The timing of biotite alteration is both significant and contentious. Biotite altered volcanic rock is the most abundant host to copper-gold mineralization. Our observations from pit mapping and drill core logging indicate that a



a) Biotite alteration as remnant "cores" (dark) in Bonanza volcanics cut by dark magnetite-actinolite ±pyrite veinlets with narrow white plagioclase envelopes, encroached on by albitic alteration (pale grey, on left) and chloritic alteration (grey, on right). Late white calcite-zeolite fractures cut all alteration types (east end of pit, 280 level).

b) "Mottled" albite (light grey)-chlorite (dark grey) alteration replacing biotite±magnetite altered Bonanza volcanics, cut by late white calcite-zeolite fractures (east end of pit, 320 level).

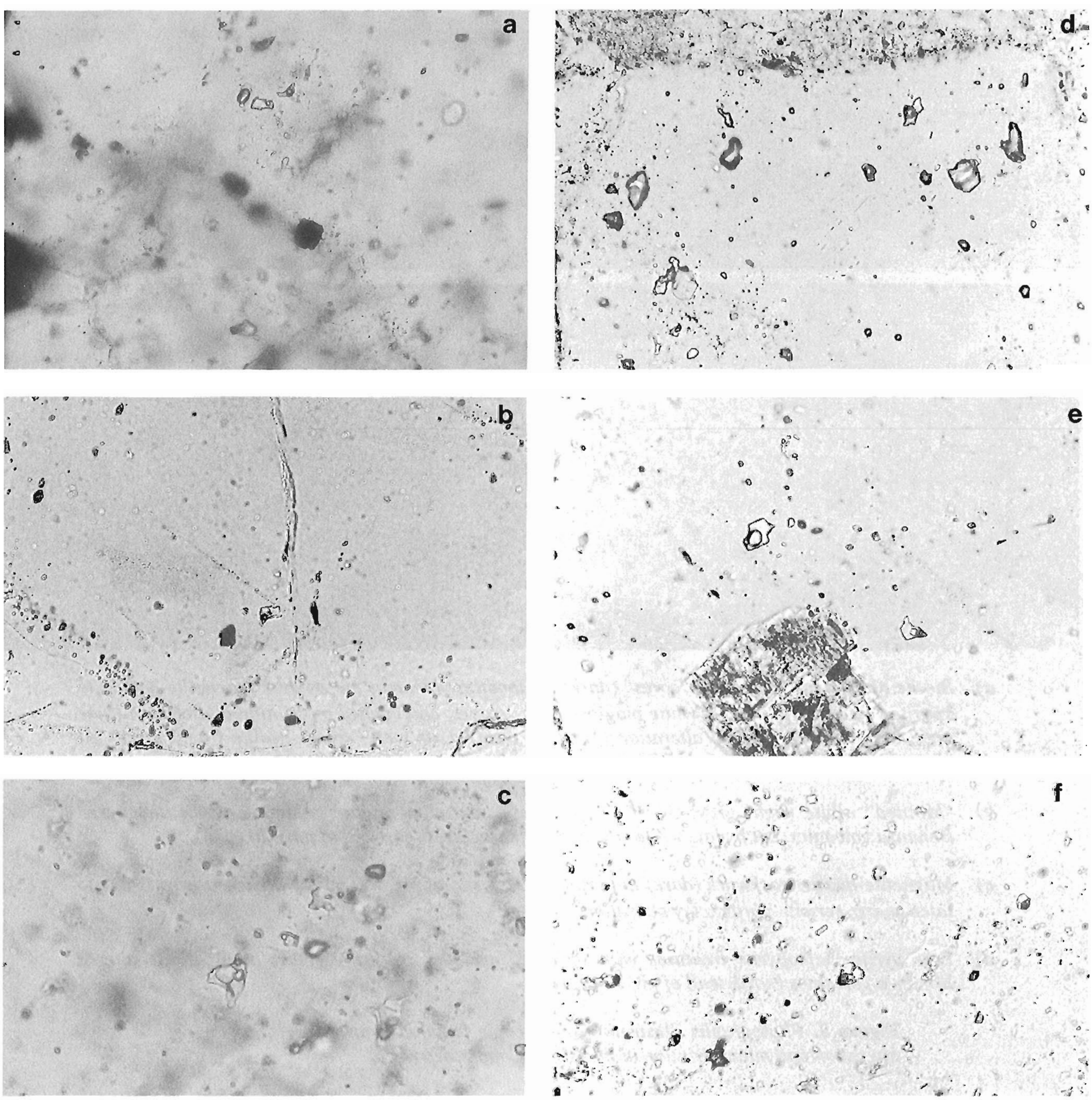
c) Magnetite-quartz stockwork (dark) in porphyry (east end of pit, 160 level). Bleached area is due to later quartz-sericite-pyrite±clay alteration.

d) Late pyrite-chalcopyrite fractures with sericitic envelopes cutting variably biotite-albite altered Bonanza volcanics (north wall of pit, 280 level).

***Figure 3.** Photographs illustrating mesoscopic features bearing on the timing of alteration and mineralization in the Island Copper deposit.*

pervasive biotite alteration is everywhere cut by magnetite-actinolite/chlorite-plagioclase-pyrite-chalcopyrite veinlets. However, in places (rare in the pit; not uncommon in drill core) the magnetite-bearing veinlets are themselves cut by brown biotite-filled fractures and biotite envelopes to quartz-chalcopyrite veins. The biotitic envelopes appear in many places to be retrograded to later sericite. The principal occurrence of the earlier biotite is as widespread remnant "cores" in relatively less fractured areas of the veinlet controlled magnetite-actinolite-plagioclase alteration assemblage (Fig. 3a). Therefore we interpret two biotite episodes: an earlier hornfelsic biotite that predated the magnetite-actinolite/chlorite-plagioclase±chalcopyrite± pyrite alteration, and a later hydrothermal biotite localized along fractures that cut magnetite-bearing alteration assemblages. Biotite-magnetite along with the copper mineralization partly overprints the quartz-amphibole-magnetite "core" assemblage according to Perelló et al. (1989); and biotite-chalcopyrite assemblages that crosscut magnetite have been described by Arancibia and Clark (in press).

Epidote-chlorite hydrothermal alteration assemblages form a peripheral shell that grades outwards to a regional metamorphic assemblage of the same minerals. Quartz-sericite-pyrite±clay alteration assemblages are best developed in the quartz-feldspar porphyry intrusions. It overprints quartz-magnetite stockwork, resulting in a quartz-pyrite stockwork. Locally porphyritic intrusions are reduced to a



mass of clay with rounded quartz crystals and pyrite. The pyrophyllite-dumortierite alteration assemblages that occur in the upper levels of the western end of the pit, apparently overprinting fragmental volcanic rocks and/or breccias (see below), will be the focus of a MSc. thesis by Geraint Mathias at Auckland University.

Mineralization and veining

At the Island Copper deposit, multiple episodes of copper introduction are interpreted from crosscutting relations observed in drill core and pit exposures, but require refining by detailed petrography.

1. The first introduction was near the end of the period of quartz-magnetite veining (e.g. minor chalcopyrite is found with pyrite in magnetite veins in the Bonanza volcanic rocks, particularly in the east end of the pit, north side.
2. The "main-stage" copper introduction involved abundant fine hairline fracture fills and disseminations either accompanying or cutting pervasively biotitized rock – it is not clear which. In places this style of mineralization is accompanied by quartz-pyrite-chalcopyrite veins (±biotite envelopes where they cut volcanics, but not porphyritic intrusive rocks; many of these veins now have sericitic envelopes). This is comparable to the main Cu introduction of Arancibia and Clark (in press).
3. Minor epidote-chlorite±pyrite±chalcopyrite veining may represent either minor introduction or possibly remobilization of copper.
4. Pyrite-chalcopyrite fractures that cut all other veins (Fig. 3d) also possibly represent minor introduction or remobilization of copper.
5. Finally, the minor chalcopyrite present in late calcite-zeolite-gilsonite veins in the north wall of the pit, probably is remobilized copper.

There may have also been several episodes of molybdenum mineralization:

1. The earliest introduction occurs as disseminations, ribbons and parallel fractures in planar, frequently laminated grey-pink quartz veins. These veins are up to 0.3 m in width and occur in sets trending northwest with steep to vertical dips; they can be traced over 30 m.
2. The coarsest molybdenite occurs locally in high grade copper-molybdenum breccias developed in the intermediate porphyry.
3. The most economically significant molybdenite occurs on widespread slips that cut the late pyrite-chalcopyrite fractures.

Sphalerite was observed in rare quartz-?calcite veins peripheral to the main mineralization. It has also been noted in thin sphalerite-rich veins cutting intermediate and late porphyries, giving rise to local zones of over 1% Zn (Perelló, 1987).

Fluid inclusion petrography

A preliminary investigation of fluid inclusions was conducted on 40 thin and polished sections from previous petrographic work done on the Island Copper property. There are at least four types of inclusions present:

- Type 1: One phase or vapour-dominant (no liquid phase visible).
- Type 2: Two-phase aqueous inclusions containing liquid and vapour.
- Type 3: Moderately saline three-phase inclusions containing liquid, vapour and a salt crystal, likely halite.
- Type 4: Highly saline inclusions with multiple daughter products.

Type 1-4 inclusions occur dominantly in vein quartz and in quartz phenocrysts in the porphyries, either isolated or along fracture planes indicating pseudosecondary and secondary origin. No inclusions were observed in recognizable growth zones. Two phase inclusions occur in vein quartz, calcite, K-feldspar and ?zeolite. No temperature or salinity data are available yet for any inclusions.

a) *Type 2 (2-phase) in quartz vein, showing highly variable vapour/liquid ratios from 10 to 70 per cent in a single cluster (hole E96-711').*

b) *Type 3 (3-phase) in quartz-pyrite-chalcopyrite vein, showing vapour bubble, halite crystal, and saline brine (hole E95-387').*

c) *Type 4 (multi-phase) in quartz-magnetite vein, showing vapour bubble, transparent and opaque daughter products, and saline brine (sample 88PPit 010, intensely potassic (biotite-K-feldspar-magnetite-chalcopyrite ±pyrite-molybdenite) altered Bonanza volcanic, from unknown location in pit; note several adjacent vapour-rich inclusions).*

d) *Type 2 (2-phase) in quartz phenocryst from strongly albite-quartz-chlorite-magnetite±pyrite-chalcopyrite altered early porphyry (hole E111-367.5'). Note abundant Type 1 (dark, vapour-rich) examples in the field of view.*

e) *Type 3 (3-phase with halite cube, to right of altered feldspar crystal) and larger Type 2, in quartz phenocryst from clay-sericite-pyrite altered intermediate porphyry cut by calcite-zeolite veins (hole E138-507').*

f. *Abundant Type 4 (multi-phase) inclusions in quartz phenocryst from intensely magnetite-chlorite altered early porphyry cut by calcite-zeolite veins (hole E140-168').*

Figure 4. *Photomicrographs of typical fluid inclusions in quartz from the Island Copper deposit (all in plane polarized light; width of field of view 130 µm except 50 µm in c).*

The vapour-rich inclusions (Fig. 4) are difficult to interpret because of their superficial similarity to decrepitated inclusions that are filled with air. They are variable in size, but tend to be large (over 15 µm) and have rounded, smooth outlines with vapour to liquid ratios 90% or over. There are no visible daughter minerals. These inclusions could contain variable amounts of carbonic (CO_2+CH_4) vapour. Crushing and freezing tests will be necessary to further identify the materials present in these inclusions.

The two phase liquid-vapor inclusions are mainly associated with late quartz, quartz-epidote or chlorite reopenings of main stage quartz veins (Fig. 4a). They are also found in late quartz±calcite±zeolite±K-feldspar veins, or in fractures in quartz phenocrysts in porphyritic intrusive rocks (Fig. 4d). Vapour to liquid ratios are highly variable, from 10 to 90 per cent. These inclusions tend to be small, less than 10 µm in maximum dimension, and are rounded to irregular in shape. No consistent variation in degree of filling with location in the deposit has so far been observed.

The three phase inclusions are associated with the intermediate stage quartz±pyrite±chalcopyrite±magnetite veins, locally with potassic (biotite and/or K-feldspar) alteration envelopes. They contain a liquid phase, a vapour bubble, and a halite cube. These inclusions are the most important from the point of view of the mineralization, but are the least abundant in the veins (Fig. 4b). They also occur in quartz phenocrysts in altered quartz-feldspar porphyry (Fig. 4e). They range in size from 3-12 µm, and are rounded to irregular in shape; liquid to vapour ratios range from 10-40 per cent.

The highly saline, multiphase inclusions are associated with the early stage quartz-magnetite±actinolite/chlorite ±pyrite±chalcopyrite veins. The inclusions are generally in the 5-15 µm size range, with a few up to 30 µm. They have smooth to rounded or irregular shapes (Fig. 4c). The inclusions consist of a liquid phase, a vapor phase, a halite cube, and a variable number of other daughter products. Two colourless, platelet-shaped (one hexagonal), highly birefringent minerals are most common. A red, translucent, hexagonal phase (hematite?) and an opaque phase (magnetite or chalcopyrite?) are less frequently seen. Liquid to vapour ratios range from 10 to 30 per cent. Inclusions of this type were also commonly observed in quartz phenocrysts in intensely quartz-magnetite±actinolite altered quartz-feldspar porphyry (Fig. 4f).

The inclusion populations thus far outlined fit well with the commonly observed progression in porphyry deposits from early high-salinity fluids trapped in veins and phenocrysts to late low-salinity fluids trapped in veins with phyllic alteration envelopes (e.g. Reynolds and Beane, 1985). So-called "blue" quartz veins at Island Copper that contain scattered low-salinity fluid inclusions have been attributed to early quartz that has been recrystallized by later fluids but without affecting the vein envelope mineralogy (J.T. Reynolds, pers. comm., 1994).

Outline of work plan

Geochronology: Three samples of the quartz-feldspar porphyry, representing the early, intermediate and late phases, have been collected for zircon U-Pb dating. These data will complement the zircon U-Pb dating of rhyolitic and andesitic phases of the Bonanza volcanic rocks currently underway on rocks collected by the BCGSB.

Geochemistry: Approximately 50 samples of the main alteration types in volcanic and porphyry and representative, least-altered samples of the three porphyry phases have been submitted for whole-rock and trace element analysis.

Isotope geochemistry: Samples of vein and phenocryst quartz and feldspar, plus vein calcite, magnetite, and hydrous minerals (chlorite, actinolite, biotite) will be analyzed for oxygen, deuterium, and carbon isotopes. Analyses of feldspar lead and sulphide sulphur are also planned.

Petrography: A comprehensive suite of samples from the lower levels of the pit and from drill core was collected this year to examine alteration changes in detail. Data will be presented as five cross sections and two long sections.

Fluid inclusion studies: At the time of the preliminary investigation, the hand sample equivalents of the thin sections were not available, therefore the overall relationships of veins and alteration were somewhat ambiguous. However, a well constrained set of samples was collected during this year's fieldwork to continue the fluid inclusion study. Following detailed petrography of these samples, microthermometry will be completed.

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Geological Survey of Canada Project 740098

Comparison of drape-flown and computationally draped aeromagnetic data in central British Columbia

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Abstract: A recent GSC aeromagnetic survey flown at a 305 m mean terrain clearance in central British Columbia overlaps with an earlier survey flown at a constant barometric altitude of 2438 m. This has allowed the direct comparison of drape-flown and computationally draped magnetic data. The two compare well, with the drape-flown data showing better resolution of features with wavelengths less than ~4 km. The low-pass filtering required in the computational draping methods has likely led to some reduction in power at shorter wavelengths as compared to the flown data.

Résumé : Un levé aéromagnétique récent effectué par la CGC à une hauteur moyenne de 305 m au-dessus du relief dans le centre de la Colombie-Britannique recoupe un levé antérieur réalisé à une altitude barométrique constante de 2 438 m. Cela a permis une comparaison directe entre les données magnétiques recueillies à une hauteur constante par rapport au relief et les données de même altitude corrigées informatiquent pour tenir compte du relief. Les deux méthodes soutiennent la comparaison, mais les données enregistrées à une hauteur constante par rapport au relief permettent d'obtenir une meilleure résolution des éléments dont les longueurs d'onde sont inférieures à environ 4 km. Le filtrage passe-bas nécessaire à la simulation du relief a vraisemblablement entraîné une certaine réduction de puissance aux longueurs d'onde plus courtes comparativement aux données enregistrées à la même hauteur par rapport à la surface du sol.

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INTRODUCTION

In areas of rugged topography, such as the interior of British Columbia, several aeromagnetic surveys have been flown at a constant barometric altitude because fixed wing operations prevented drape flying. As a consequence, mismatches may arise at the survey edges where drape-flown and constant altitude surveys join. Furthermore, resolution is degraded over the deeper valleys and depressions because of the increased distance to the magnetic sources. The loss in resolution and poor fit at survey edges can be partly remedied by computationally draping the constant altitude survey data onto a surface (most often a constant distance above the topography) to mimic a truly drape-flown survey. The continuation of the magnetic field from a horizontal plane to a draping surface results in data with improved resolution, but without the attendant problems inherent in drape-flown data.

Two methods of draping aeromagnetic data that we will consider here are the Taylor series approach (Pilkington and Roest, 1992) and the COMPUDRAPE™ technique (Paterson et al., 1990). Many approaches to draping have used a Taylor series expansion of the observed field (B_{level}) on a horizontal plane given by

$$B_{surf} = \sum_{n=0}^{\infty} \frac{h^n(x,y)}{n!} \frac{\partial^n}{\partial z^n} B_{level}(x,y)$$

where B_{surf} is the field on the draping surface h(x,y). Evjen (1936) originally used this approach on gravity data. Grauch and Campbell (1984) used a single term Taylor series approximation for magnetic data while Cordell and Grauch (1985), using a three term expansion, draped the observed field onto the basement surface to map magnetization zones. Convergence of the series is dependent on the frequency content of the measured field and the continuation distance and has been discussed for a number of source geometries by Roy (1967) and Negi (1967). Excessively large amplitude terms in the series will result from large gradients in the data (usually due to magnetic source body edges) and large downward continuation distances. Taking n^{th} order derivatives also leads to magnification of the high frequency content (noise, gridding artifacts) of the data which may obscure meaningful geological signals. Consequently, it is necessary to apply low-pass filtering to the Taylor series terms.

Simple level to level continuation is the basis for the COMPUDRAPE™ technique (Paterson et al., 1990) which is based on the "Chessboard" approach of Cordell (1985). Initially, the field is upward continued onto a series of level planes, then at each grid point, values are vertically interpolated onto an interim continuation datum, which is simply a vertically displaced (upward) version of the specified draping surface. The continued data are then treated like the field on a level plane and downward continued a constant distance onto the draping surface. The interim datum is used to avoid problems with continuation through magnetic sources since it is likely that some sources will lie above the lowest point on the draped surface. As in the case of the Taylor series approach, a low-pass filter must be applied when the downward continuation is done. COMPUDRAPE™ allows for tuning of the severity of this filter as a function of the continuation distance.

EDGE HILLS SURVEY, BRITISH COLUMBIA

Aeromagnetic surveys from the Edge Hills area of central British Columbia have been used previously to demonstrate the utility of computational draping to bring two surveys, one flown as a loose drape and the other at constant barometric altitude, to the same level (Paterson et al., 1990; Pilkington and Roest, 1992). Figure 1 shows the survey distribution with areas 2A and 3 being drape-flown at a mean terrain clearance (MTC) of 305 m and area 2B flown at a constant barometric altitude of 2438 m. All surveys were flown with east-west flight lines separated by 1000 m and gridded at a 200 m interval. Surveys 2A and 2B flown in 1987 were used in the earlier draping studies before area 3 was flown in 1993-1994. The recent availability of survey area 3 data and the fact that it overlaps with the computationally draped area 2B (Fig. 1) has allowed a direct comparison of the two.

For comparative purposes, two profiles (locations given in Fig. 1) from the constant altitude survey 2B, the drape-flown data of area 3 and the calculated draped data from area 2B were extracted from the appropriate grids. Any degradation of the original data due to gridding and subsequent interpolation for extracting the two specified profiles is assumed to be comparable for the flown and processed data. Figures 2 and 3 show the extracted drape-flown data and constant altitude data along with the computational draping results from the Taylor series and COMPUDRAPE™

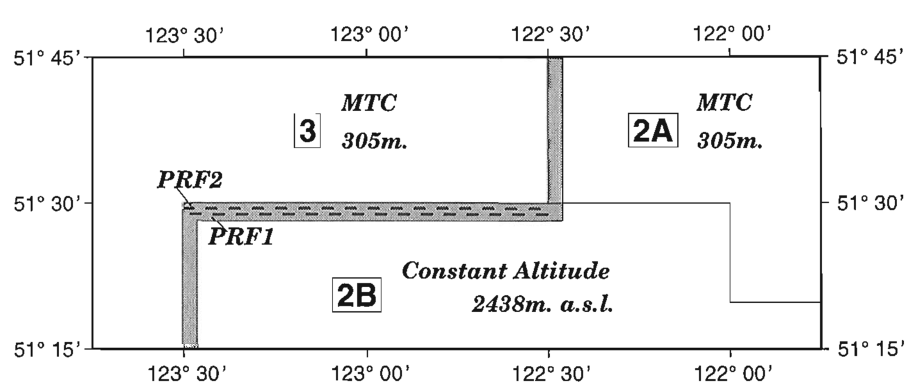

Figure 1.

Edge Hills area survey distribution. Shaded area denotes overlap between surveys. Profiles 1 and 2 are indicated within this zone. MTC – mean terrain clearance.

approaches. Also shown are the two respective draping surfaces. For the drape-flown survey, this is given by the measured barometric altitude and for the calculated data, this is a surface 305 m above the radar altimeter values. Radar altimeter data from survey area 2B were not usable and were generated from digitizing the topography of the area and subtracting this from the barometric altitude. For both profiles, the measured barometric altitude is generally higher than the surface on which the draped field is calculated. The root mean square (RMS) difference between the two is 77.4 and 55.8 m for profiles 1 and 2, respectively. This corresponds to ~10 per cent of the average continuation distance (>650 m) which we consider small enough that the results of draping can be compared directly to the drape-flown data.

The results of draping using both methods show a good correspondence with the flown data for profile 1 (Fig. 2), but less so for profile 2 (Fig. 3). The agreement between the COMPUDRAPE™ and Taylor series approaches is consistently good with RMS differences of 22.6 and 29.2 nT for profiles 1 and 2. Differences are greater between the individual

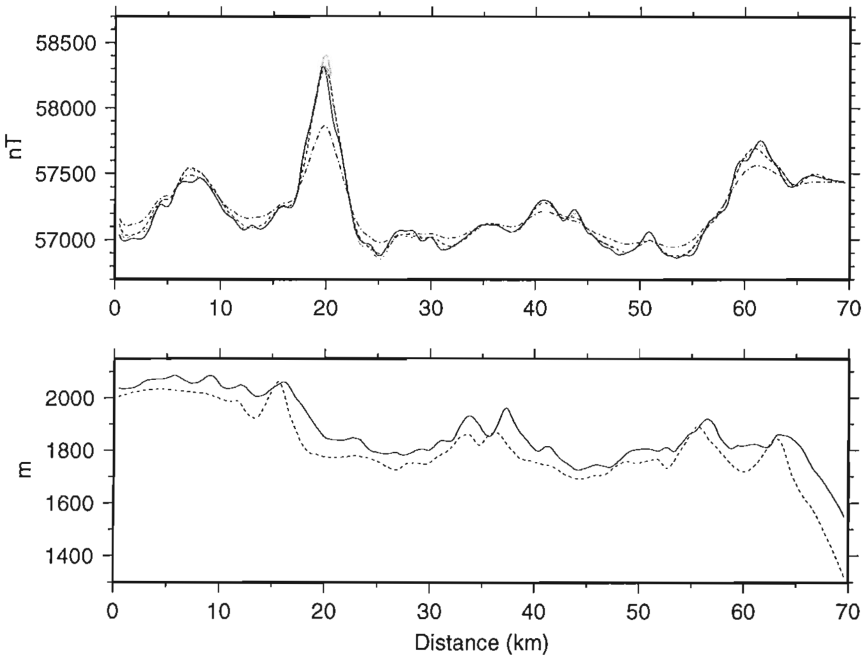

Figure 2.

Profile 1. (Top) Solid line, drape flown data from survey 3. Dashed-dotted line, constant altitude data from survey area 2B. Dashed line, calculated drape using Taylor series method. Dotted line, calculated drape using COMPUDRAPE™. (Bottom) Solid line, barometric altitude for drape-flown data. Dashed line, draping surface for computationally draped data (topographic elevation plus 305 m).

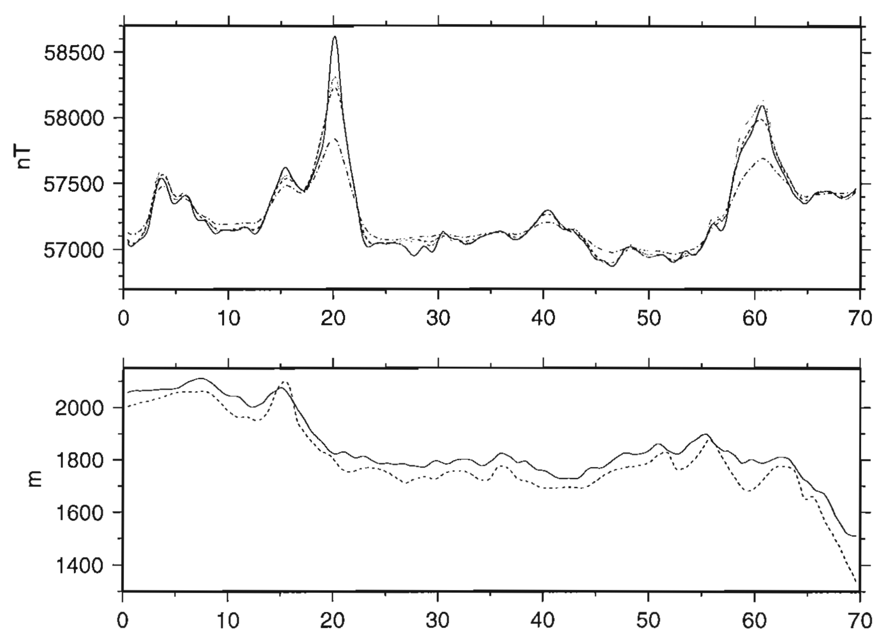

Figure 3.

Profile 2. (See caption for Figure 2.)

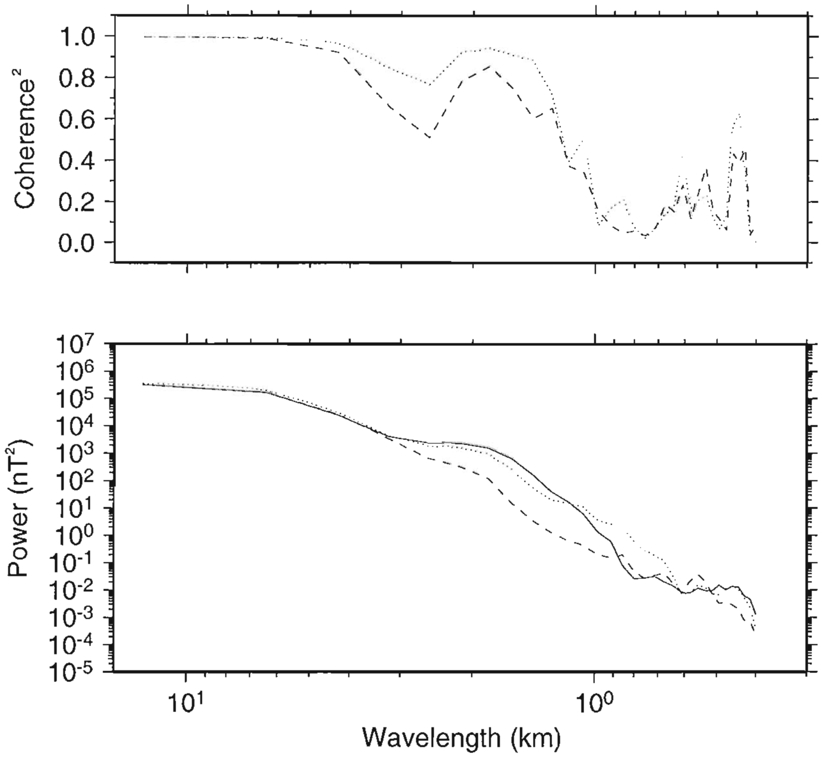

Figure 4.

Profile 1. (Top) Dashed line, coherence between Taylor series method draping and drape-flown data. Dotted line, coherence between COMPUDRAPE™ and drape- flown data. (Bottom) Solid line, power spectrum of drape-flown data. Dashed line, power spectrum of Taylor series draped data. Dotted line, power spectrum of COMPUDRAPE™ data.

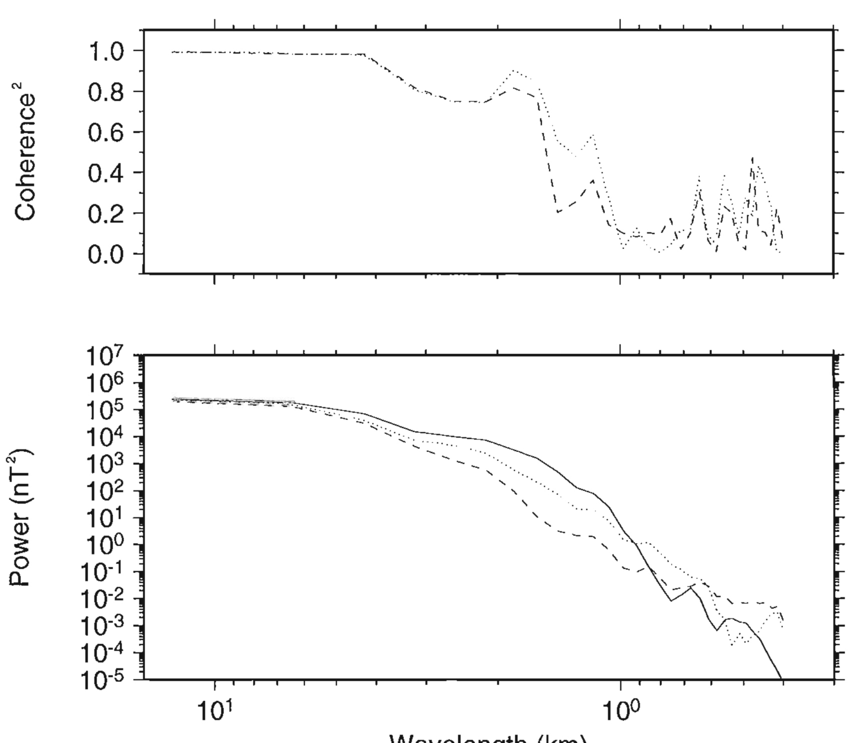

Figure 5.

Profile 2. (See caption for Figure 4.)

draping methods and the flown data: for profile 1, RMS differences are 31.7 nT for COMPUDRAPE™ and 36.2 nT for the Taylor series. For profile 2, they are 46.7 and 52.4 nT, respectively. Anomaly amplitudes are generally higher for the drape-flown data (e.g., at 15 km, 20 km on profile 2) which also shows better resolution of some of the shorter wavelength features e.g., at 60 km on profile 1. The comparison as a function of wavelength is better seen in Figures 4 and 5 which show the coherences between the flown and calculated data and also their respective power spectra. For both profiles, the coherence is essentially zero for wavelengths smaller than 1 km. Since area 2B was flown at a constant altitude of 2438 m, more than 950 m above the average topography, we expect the smallest wavelength anomalies due to magnetic sources detectable at such a distance to be in the 1-2 km range. Hence it is likely that wavelengths less than 1-2 km in the computationally draped data are mainly magnified instrument noise and gridding artifacts rather than geologically meaningful signals.

Both profiles show maximum coherences for wavelengths >4 km with values of close to unity. As mentioned earlier, both draping techniques require some low-pass filtering to stabilize the downward continuation onto the draping surface. The power spectra of Figures 4 and 5 suggest that such filtering may have caused some reduction in power at shorter wavelengths as compared to the flown data.

CONCLUSIONS

Profiles of observed and calculated draped magnetic data compare favourably and demonstrate that computational draping of constant altitude surveys is a viable approach to improving the utility of such data. Draping guarantees the continuity of trends across survey boundaries and ensures the data have a resolving power as close as possible to drape-flown aeromagnetic surveys.

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Geological Survey of Canada Project 930024

Tertiary volcanic stratigraphy of the Clisbako River area, central British Columbia[1]

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Metcalfe, P. and Hickson, C.J., 1995: Tertiary volcanic stratigraphy of the Clisbako River area, central British Columbia; in *Current Research 1995-A; Geological Survey of Canada, p. 67-73.*

Abstract: Eocene volcanic rocks exposed in the Clisbako River area of central British Columbia host potentially economic epithermal alteration and mineralization. Two major assemblages were identified by mapping and petrography. Weakly plagioclase+augite-phyric intermediate volcanics are exposed throughout the area. This assemblage comprises flow breccias grading laterally and upsection into colonnade-forming and flow-laminated lavas.

The second assemblage is identified by euhedral biotite phenocrysts and comprises rhyolite flow domes and pyroclastic units. The latter contain accidental blocks of plagioclase+augite-phyric dacite. In the central area rare exposures of quartz porphyry contain relict biotite phenocrysts.

Eocene volcanic rocks form a circular highland area, 50 km in diameter. Biotite-phyric pyroclastic rocks occurring in the centre and southeast are interpreted as the products of a large pyroclastic eruption. The area is interpreted as the erosional remnant of a large caldera, partially filled with basaltic lavas of the Chilcotin Group.

Résumé : Des roches volcaniques de l'Éocène qui affleurent dans la région de la rivière Clisbako, dans le centre de la Colombie-Britannique, contiennent des indices d'altération et de minéralisation épithermales présentant un potentiel économique. Deux assemblages majeurs ont été reconnus par cartographie et pétrographie. Des roches volcaniques intermédiaires contenant quelques phénocristaux de plagioclase+augite sont exposées dans toute la région. Cet assemblage comprend des brèches de coulée se transformant latéralement et vers le haut de la coupe en laves en colonnades et à litage de flux.

Le deuxième assemblage est identifié par des phénocristaux euédriques de biotite et comprend des dômes de rhyolite et des unités pyroclastiques. Ces dernières contiennent par endroits des blocs de dacite à phénocristaux de plagioclase+augite. Dans la région centrale, de rares affleurements de porphyre quartzique contiennent des vestiges de phénocristaux de biotite.

Les roches volcaniques de l'Éocène forment une zone circulaire de hautes terres de 50 km de diamètre. Les roches pyroclastiques à phénocristaux de biotite qui se rencontrent dans le centre et le sud-est de la zone sont vraisemblablement le produit d'une importante éruption pyroclastique. La zone pourrait être le vestige d'érosion d'une vaste caldeira, partiellement remplie de laves basaltiques du Groupe de Chilcotin.

[1] Contribution to Canada-British Columbia Agreement on Mineral Development (1991-1995), a subsidiary agreement under the Canada-British Columbia Economic and Regional Development Agreement.

INTRODUCTION

This paper summarizes current results of reconnaissance geological mapping, sampling, and petrological analysis of Tertiary volcanic rocks in the Clisbako River area of central British Columbia. This project, to identify and correlate Tertiary volcanic rocks in the Clisbako area, is part of the Canada-British Columbia Agreement on Mineral Development (1991-1995) (van der Heyden et al., 1993, 1995).

A preliminary study of Tertiary volcanic stratigraphy in the Clisbako area of central British Columbia (Fig. 1) began in September 1993. A second period of fieldwork was conducted in June and July 1994. The purpose was to determine the stratigraphic succession and petrological relationships of Early Tertiary felsic volcanic rocks that host epithermal mineralization discovered on the BAEZ and CLISBAKO claim groups (MINFILE numbers 093C-015 and 093C-016, respectively), near the headwaters of the Clisbako River.

The study area is part of the Chilcotin Plateau and comprises four 1:50 000 map sheets (93B/12, B/13, C/9, C/16), bounded by latitudes 52°30′N and 53°00′N and by longitudes 123°30′W and 124°30′W (Fig. 2). Relief is gentle and the area is forested. Outcrop is not abundant. The area is accessible by means of numerous logging roads, west from Quesnel and northwest from Alexis Creek.

Findings from this study area will be compared with information from Tertiary volcanic rocks exposed to the north in the Nechako River area (93F) (Green and Diakow, 1993; Diakow et al., 1993; Diakow and Webster, 1994) and to the south in the Taseko Lakes area (92O) (Hickson et al., 1991; Hickson, 1992, 1993; Hickson and Higman, 1993). A final report for the project, including petrographic and geochemical data, will be produced.

GEOLOGY OF THE CLISBAKO AREA

Older volcanic rocks

The basement to the Clisbako area consists of intermediate fragmental rocks, which differ from other lithologies examined in that a schistose foliation is present. Outcrop scarcity precluded comprehensive structural mapping. Tipper (1959, 1969) assigned this assemblage to the Lower Jurassic Hazelton Group, based on lithological similarity. These older volcanics were not examined in detail.

Clisbako volcanic assemblage

Basement rocks are overlain by a succession of intermediate to felsic volcanic rocks that are the subject of the present study. The approximate area underlain by the assemblage is

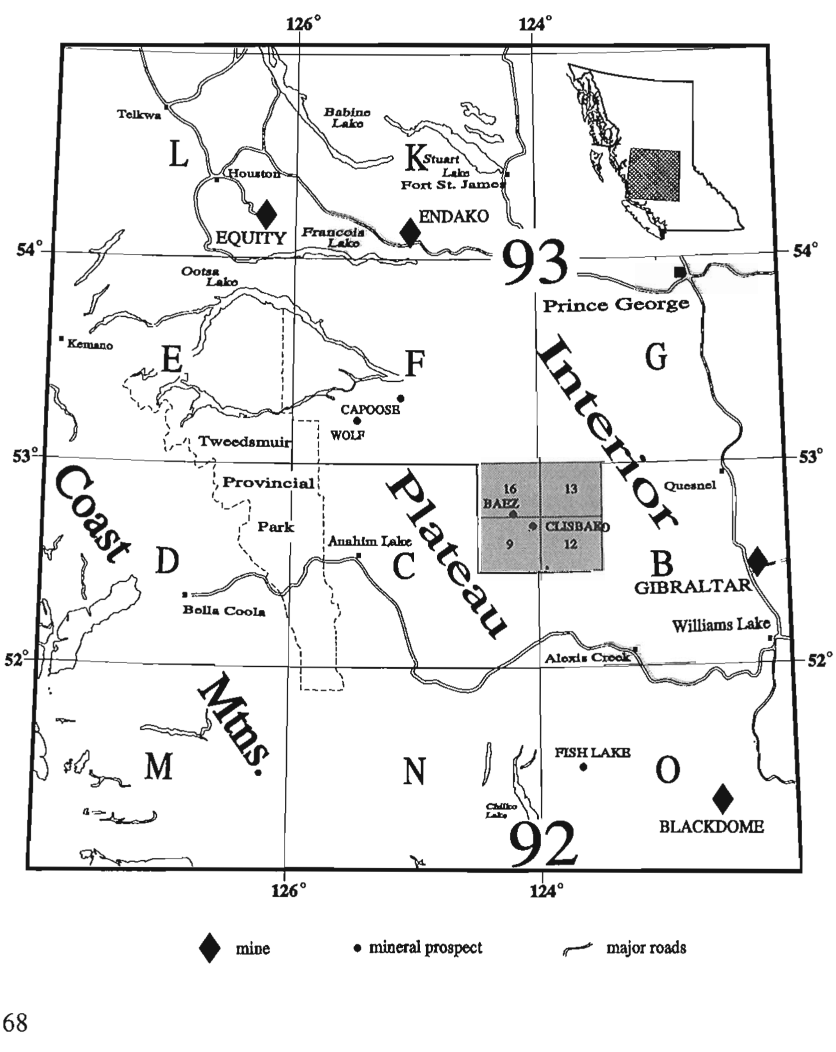

Figure 1.

Location of the study area.

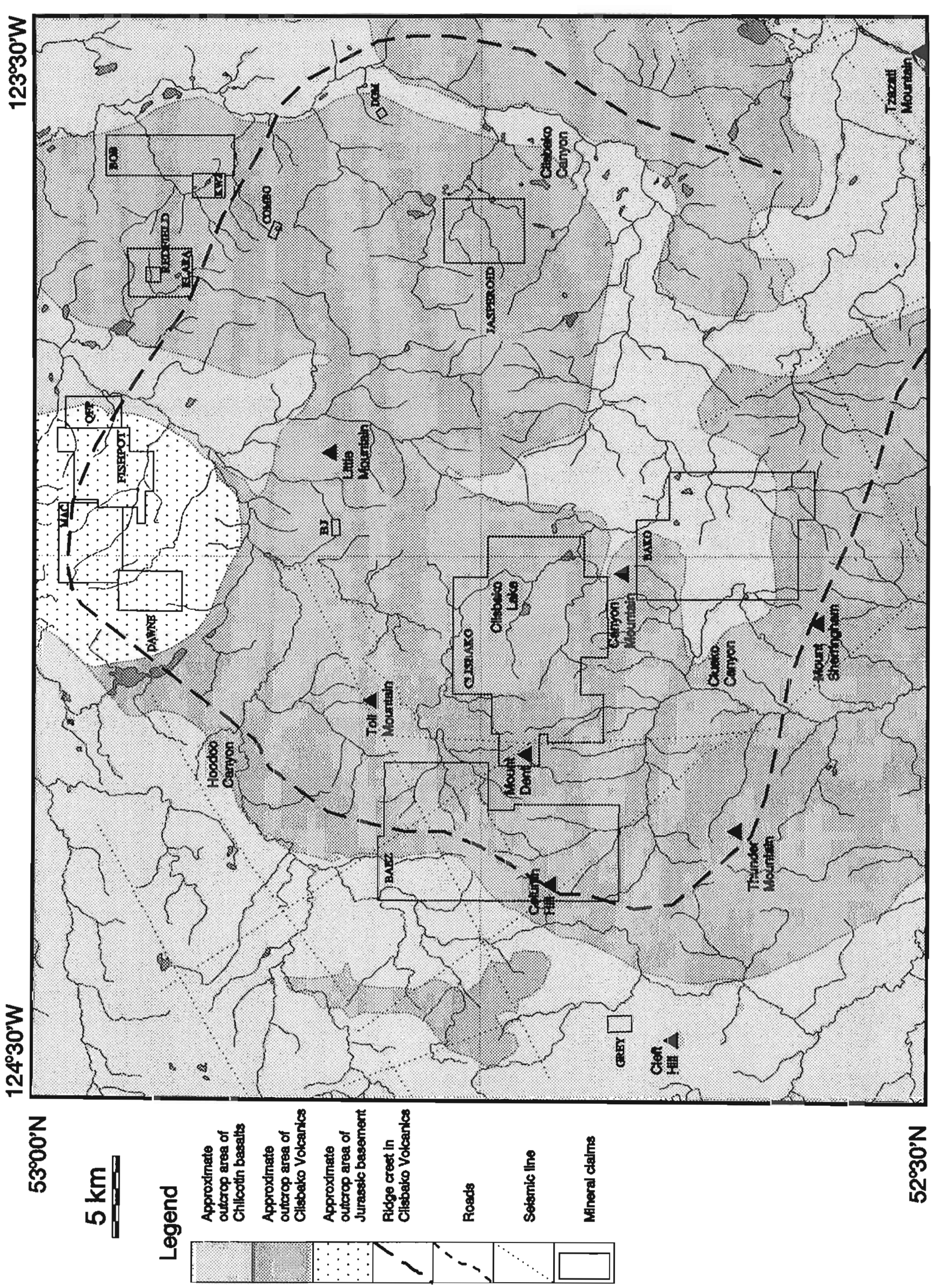

Figure 2. *Approximate area underlain by Eocene and Miocene volcanic rocks (this study; Tipper, 1959, 1969). The area was glaciated and outcrop is not abundant in lowland areas. Eocene volcanic rocks rest upon a deformed basement of presumed Jurassic age. Eroded remnants of Eocene volcanic rocks are partially covered by valley-filling Chilcotin Group basalts.*

shown in Figure 2. Eocene rocks underlie all higher ground in the study area and were identified by Tipper (1959, 1969) as part of the Ootsa Lake Group. Hydraulic brecciation, epithermal alteration, and mineralization are locally abundant in this assemblage.

Eocene felsic and intermediate volcanic units outcroping in the study area include four lithological assemblages, identified on the basis of fieldwork and petrography. Contacts are rarely exposed and the precise stratigraphic relationship between the assemblages has not yet been determined. The relationships observed at seven sections in the study area are shown in Figure 3.

Plagioclase+augite-bearing assemblage

The most commonly exposed assemblage is weakly to moderately porphyritic, containing plagioclase phenocrysts and/or augite phenocrysts. Both phenocryst phases are usually less than 3 mm in size and make up, at most, 15 per cent of the rock. Where both occur, plagioclase is early. Plagioclase is subhedral to euhedral and commonly exhibits strong oscillatory zoning. Epidote alteration is common but rarely pervasive. Augite phenocrysts are euhedral to subhedral, often altered to amphibole and, less commonly than plagioclase, exhibit oscillatory zoning.

Several sections are shown in Figure 3. The lowest part of the plagioclase+augite assemblage is exposed mainly in the western part of the study area. Basal units are a thick sequence of flow breccias containing glassy, flow-banded, aphyric, and porphyritic blocks up to 2 m in size in a red-, yellow-, or cream-weathering matrix. The flow breccias are intercalated with discontinuous lobes of coherent dacite, exhibiting flow banding and flow folding. These are interpreted as non brecciated flow lobes within the breccia; their chemical composition is nearly identical to that of unaltered blocks.

Areas underlain by flow breccias are of moderate topography and poor in outcrop, usually near contacts with the more resistant overlying units. The flow breccia sequence has a total thickness of at least 100 m; breccias include blocks with glassy aphanitic groundmasses, plagioclase-phyric and plagioclase-augite-phyric blocks, and blocks exhibiting well developed flow banding and flow folding.

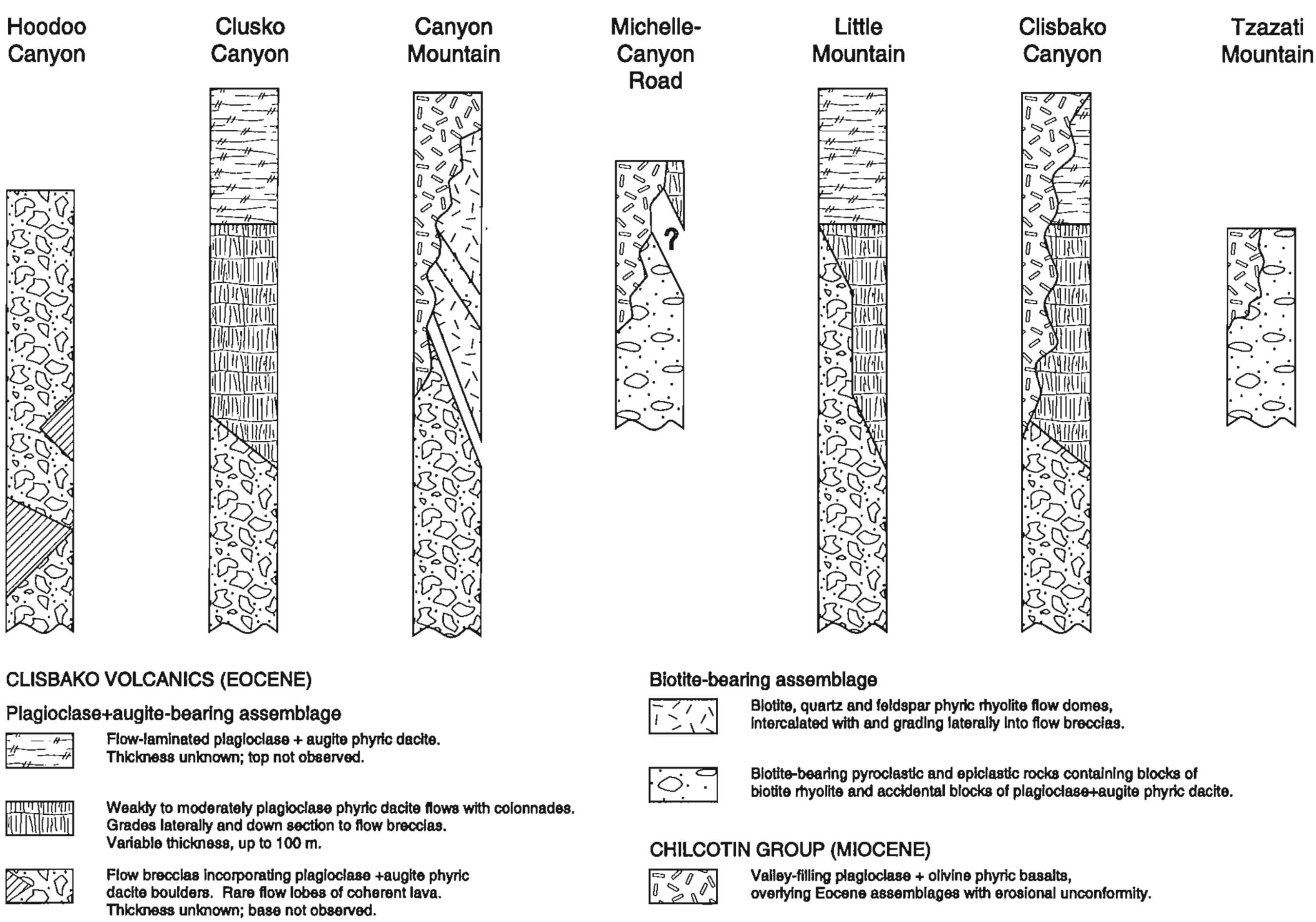

Figure 3. Stratigraphic relationships of Eocene volcanic assemblages present in the study area. The biotite-bearing assemblage overlies at least some of the plagioclase+augite-bearing assemblage, but eruption of the latter may have continued after the biotite-bearing eruptions.

The flow breccias pass laterally into and are overlain by black glassy flows very similar in lithology to blocks included in the flow breccias. These flows are mainly of dacitic composition, aphyric or with plagioclase±augite phenocrysts, and are interpreted as proximal, nonbrecciated equivalents of the flow breccias.

Lava flows are usually structureless but locally exhibit flow banding and flow folding. This lithology is resistant to erosion and forms rare cliff exposures with spectacular colonnades as much as 30 m high. The total thickness of this part of the succession could not be measured, due to lack of continuous outcrop; it probably varies greatly throughout the area. Significant exposures of colonnade-forming lavas occur at Little Mountain, Clisbako Canyon, Clusko Canyon, and 8 km west of Mount Dent, at Column Hill (Fig. 2).

The colonnades are overlain by a series of intermediate (mainly dacitic) lava flows. Several of the ridges, particularly near the periphery of the Eocene outcrop area, are cored or capped by this assemblage. These lavas are also mainly of dacitic composition, although their distribution in the field suggests a more fluid and mafic composition (Metcalfe and Hickson, 1994). The rocks are poorly vesicular; the vesicles are usually less than 2 mm and are irregular in shape. Flow banding is rare and flows are typically cut by a penetrative fabric subparallel to their flow bases, which is deformed in some exposures in a manner similar to flow folding. The fabric is interpreted as a flow lamination caused by the streaking out of vesicles. Flow-laminated flows may therefore be lateral equivalents of colonnade-forming flows but appear to overlie the latter in all parts of the area. Alternatively, flow-laminated exposures may represent the upper portions of thick (>100 m) dacite flows, while the colonnades represent portions of the flow centre. A contact between the lithologies was found in Clusko Canyon (Fig. 2), but was not accessible. The top of a flow-laminated unit was not observed.

Biotite-bearing assemblage

Canyon Mountain, in the south-central portion of the area (Fig. 2, 3), is underlain by a flow dome complex consisting of biotite-porphyritic rhyolite flows and flow breccias intercalated with pyroclastic and epiclastic fragmental units of similar composition. Similar fragmental units occur 11 km northeast of Canyon Mountain and in that location exhibit well developed reverse grading, interpreted as the result of a pyroclastic flow. Both here and at Canyon Mountain, the fragmental units include blocks of plagioclase+augite-phyric dacite.

The biotite-bearing assemblage is poorly exposed, but occurs in one other significant exposure on Tzazati Mountain, in the extreme southeastern part of the area (Fig. 2). The outcrop is separated from those in the central portion of the area by a topographic low, filled with basalts of the younger Chilcotin Group (Fig. 2). The topographic low probably formed by erosion of less resistant fragmental rocks. The topography and distribution of fragmental rocks may indicate collapse of the volcano to the southeast, followed by eruption of a small resurgent dome at Canyon Mountain.

The relative ages of biotite-bearing and plagioclase+augite-bearing assemblages are unknown. Blocks of the latter assemblage occur in biotite-bearing fragmental rocks. However, northeast of Canyon Mountain, near the junction of the Michelle-Canyon and Michelle-Baezaeko forest service roads, felsic pyroclastic rocks underlie at least one unit of black felsic flow breccias. In the roadside exposure to the east of Canyon Mountain, strata appear to dip moderately to the east, beneath a hill capped by a dacite colonnade. Fragmental rocks may fill paleovalleys, but it is also possible that eruption of plagioclase+augite-bearing lavas persisted after eruption of the less commonly occurring biotite-bearing assemblage.

Quartz porphyry

Small exposures of a white- to buff-weathering felsic rock whose contacts were obscured at all localities examined are found south of Clisbako Lake (Fig. 2). The unit may be intrusive or extrusive. The unaltered rock contains 5-10 per cent subhedral biotite phenocrysts (1-2 mm) and 20-25 per cent euhedral to subhedral quartz phenocrysts (2-5 mm) in an extremely fine-grained groundmass or matrix. Quartz phenocrysts enclose relict mica. Minor hornblende phenocrysts are also present.

Quartz porphyry was not observed elsewhere in the area, although smaller amounts of quartz occur in the biotite-bearing assemblage. The occurrence of relict mica phenocrysts suggests that this lithology is part of the biotite-bearing assemblage, but the abundance of modal quartz is unusual.

Amygdaloidal lava

Minor exposures of amygdaloidal lava are found south of Thunder Mountain, east of Clisbako Canyon, and in several other parts of the study area. They have not yet been examined in detail. The lithology is green as a result of pervasive chlorite-epidote alteration and contains as much as 10 per cent amygdules, 0.5-3 cm in size. The amygdules contain fine-grained silica and, rarely, carbonate. The flows appear to overlie the other Eocene lithologies in the Clisbako area and probably represent the latest stages of volcanic activity in this period.

Chilcotin Group basaltic rocks

Relatively well exposed basaltic lava flows occur in most valleys in the study area. These lavas are distinctively fresh in appearance and commonly show well developed colonnades and flow tops. They were assigned by Tipper (1959, 1969) to the Miocene Chilcotin Group and were also described by Mathews (1989). Tipper noted the presence of a number of small cinder cones in valleys and on ridge crests, which are probable source vents for the lavas. These cones could be more abundant than suggested by Tipper's mapping.

One such cone (Cleft Hill) lies southwest of the Eocene outcrop area (Fig. 2) and was examined during the course of 1994 mapping. It has a central crater which is open to the north. Rocks exposed on the crater wall comprise interlayered agglutinate and vesicular basalt blocks. The cone appears to

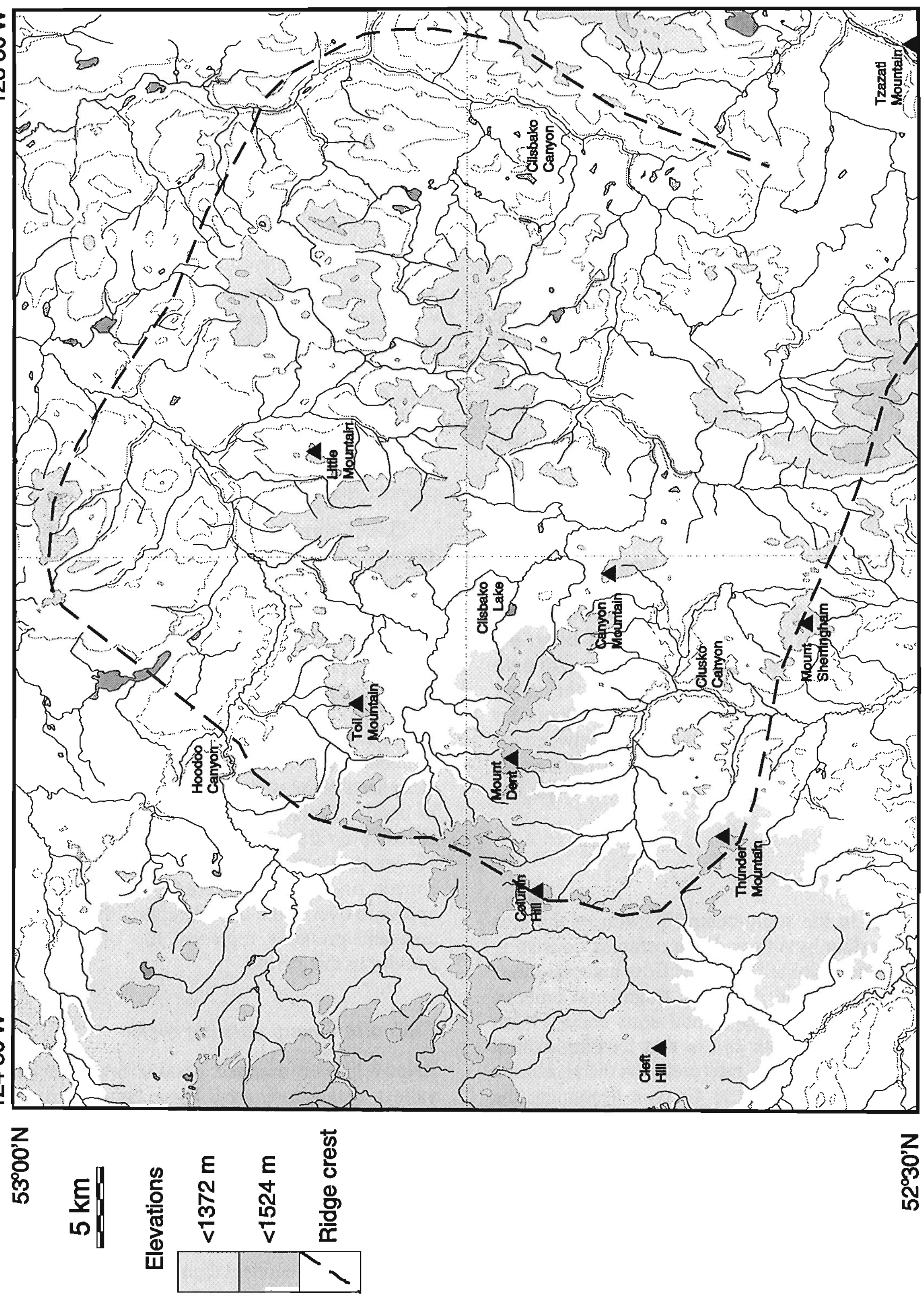

Figure 4. *Topography of the study area. Ridges cored by Eocene rocks form a circular pattern, interpreted as the remnant of an Eocene volcano. Lowland areas are filled with younger Chilcotin Group basalts.*

have a very low proportion of scoriaceous material, compared to the 7200 ka Nazko cone (Souther et al., 1987) in the northeastern part of the area.

SUMMARY

The study area contains three volcanic assemblages, the oldest a deformed volcanic basement of presumed Jurassic age and the youngest, valley-filling basaltic flows of unknown age identified as Chilcotin Group. The oldest undeformed units in the area are felsic to intermediate volcanic flows and fragmental deposits of Eocene age, which host hydrothermal alteration and mineralization.

Eocene volcanic rocks are subdivided into two minor and two major lithological types, on the basis of fieldwork, chemical analysis and petrographic examination. The most commonly exposed assemblage is a sequence of flat-lying to moderately inclined flows and related flow breccias, most commonly dacites but ranging to rhyolitic and andesitic compositions. Lithologies include flow breccias near the base, passing upsection to colonnade-forming and flow-laminated lava flows. Phenocrysts are rare and comprise plagioclase and augite, both phases showing oscillatory zoning. Plagioclase is the first-formed phenocryst, in all cases observed.

The second assemblage is characterized by the presence of biotite as a phenocryst phase. Biotite rhyolite flow domes and associated pyroclastic rocks are exposed in the central part of the area and fragmental rocks are exposed at the southeastern edge. Quartz porphyry occurrences in the central part of the area may be related to this assemblage. Rare occurrences of pervasively altered intermediate flows near the periphery of the Eocene outcrop area are interpreted as the latest products of Eocene volcanism.

The outcrop area of felsic volcanic rocks and overlying mafic assemblage forms an approximately circular highland area with a diameter of approximately 50 km (Fig. 4). It is eroded to the north, exposing the basement and the central part of the area in a topographic low, partially filled by younger basaltic lavas of the Chilcotin Group. The area described could have been the site of a large composite volcano that underwent caldera formation and subsequent erosion prior to eruption of Chilcotin Group basalts. Thus, the currently subeconomic mineralization of the Clisbako area could be related to caldera formation and exploration targets in the area may be masked by Chilcotin Group lavas.

ACKNOWLEDGMENTS

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Geological Survey of Canada Project 930025

Current status of the Interior Plateau Program, Canada-British Columbia Agreement on Mineral Development (1991-1995)[1]

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Abstract: This report provides an update for geoscience projects carried out by the Geological Survey of Canada (GSC) under the Interior Plateau Program, a multidisciplinary initiative that is part of the Canada-British Columbia Agreement on Mineral Development (1991-1995). The Interior Plateau is a favourable region for mineral deposits, but exploration has been hampered by an extensive blanket of plateau basalt and glacial drift. The GSC has completed regional airborne magnetic and radiometric surveys, bedrock and surficial geological mapping, till geochemistry, and a biogeochemical survey, as part of a regional program to assess the potential for buried mineralization. The current program concludes in 1994-1995.

Résumé : Ce rapport fournit une mise à jour des projets en sciences de la Terre réalisés par la Commission géologique du Canada (CGC) dans le cadre du programme du plateau de l'Intérieur, initiative multidisciplinaire faisant partie de l'Entente Canada-Colombie-Britannique sur l'exploitation minérale (1991-1995). Le plateau de l'Intérieur est une région propice aux gisements minéraux, mais l'exploration a été entravée par la présence d'une épaisse couverture de basalte des plateaux et de débris glaciaires. La CGC a complété des levés radiométriques et magnétiques aériens régionaux, de même que la cartographie du socle et des roches superficielles, l'analyse géochimique des tills et une étude biogéochimique, dans le cadre d'un programme régional pour évaluer le potentiel en minéraux enfouis. Le présent programme se terminera en 1994-1995.

[1] Contribution to Canada-British Columbia Agreement on Mineral Development (1991-1995), a subsidiary agreement under the Canada-British Columbia Economic and Regional Development Agreement.
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INTRODUCTION

The Interior Plateau Program is a multidisciplinary geoscience initiative, funded federally and provincially under the guidelines of the Canada - British Columbia Agreement on Mineral Development 1991-1995 (MDA). The Interior Plateau study area straddles the southern part of the geographic centre of British Columbia (Fig. 1). It is characterized by forested, hilly to low-lying mountainous topography in which potentially economic mineralized bedrock is largely obscured by barren Miocene and younger lava flows and extensive glacial overburden. This, combined with a partially obsolete geological database and a lack of modern geophysical and geochemical coverage, are factors that have been a severe hinderance to exploration. Economic mineral deposits are present adjacent to the study area [e.g., Endako (Mo), Gibraltar (Cu,Mo), Fish Lake (Cu,Au), Equity Silver (Cu,Ag), Silver Queen (Au,Ag), Blackdome (Au,Ag)]. Extrapolation of structural trends, plutonic suites and stratigraphy suggests potential for similar, undiscovered economic deposits in the study region. Despite its high mineral potential and reasonable road access, however, the region remained underexplored and poorly understood prior to the current MDA program.

In 1991 the Geological Survey of Canada (GSC) and the British Columbia Geological Survey Branch (BCGS) initiated a number of integrated, multidisciplinary projects throughout the Interior Plateau region, including airborne geophysical studies, bedrock and surficial geology mapping at 1:50 000 scale, till geochemistry, lake sediment geochemistry, biogeochemistry, and mineral deposit studies (van der Heyden et al., 1993; Diakow and van der Heyden, 1993; Matysek and van der Heyden, 1994). The main objectives were to upgrade the geological, geochemical and geophysical databases for the region and to develop new exploration techniques tailored for drift covered areas. Integration of the new geoscientific information has already aided mineral exploration, supported mineral potential assessment and land-use decisions in the region (Commission on Resources and Environment, 1994), and provided new insights into the tectonic evolution of southwestern and central British Columbia.

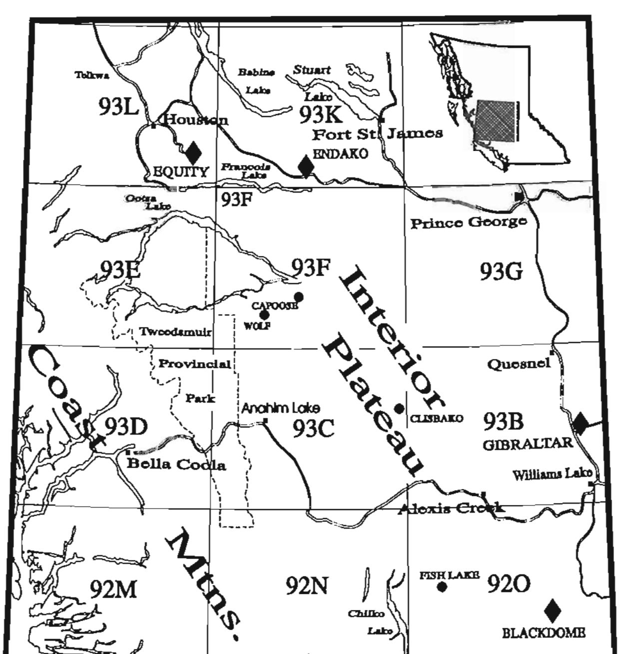

Figure 1. The Interior Plateau region, showing major producing mines (diamonds) and other significant mineral deposits (dots).

This report provides an overview of the status of GSC projects up to 1995, the final year of the current agreement. Current results of BCGS projects are presented separately in British Columbia Ministry of Energy, Mines and Petroleum Resources Paper 1995-1. Preliminary results of the combined GSC and BCGS projects were presented at a June 1994 workshop in Vancouver. A two day field trip, run in conjunction with the workshop, took participants to two mineral deposits in central British Columbia (Clisbako and Wolf Lake). Numerous GSC and BCGS reports and maps for the Interior Plateau program have been published (see selected Bibliography), and additional reports and maps will be released when completed.

REGIONAL AEROMAGNETIC SURVEY

A regional aeromagnetic survey covered 59 adjoining 1:50 000 map areas in the Interior Plateau (Fig. 2) was co-ordinated and monitored by the GSC's Aeromagnetic Surveys Section. The airborne survey was contracted to Geonex Aerodat Ltd., flown in the summer of 1993, and completed in June 1994. Data and high-resolution total field maps at 1:100 000 scale were released in November 1994 (Geological Survey of Canada, 1994). Further processing and interpretation, by GSC Aeromagnetic Section geophysicists and GSC and BCGS geologists participating in the Interior Plateau project, are necessary to minimize effects of glacial drift and extensive Miocene and younger lava flows that mask patterns caused by older strata and deeper structures. Preliminary interpretation indicates that the data set and maps live up to expectations in defining hidden geological contacts and faults, and will enable mapping of geological features across areas of sparse outcrop, thus identifying areas with high potential for mineral exploration.

AIRBORNE MULTIPARAMETER GEOPHYSICAL/GEOCHEMICAL SURVEYS

Airborne (GSC Skyvan and contract helicopter) multiparameter geophysical surveys (gamma ray spectrometry, total field magnetic, and VLF-EM), centred on the Fish Lake (92O) porphyry copper-gold and the Clisbako River (93B,C) epithermal precious metal camps (Fig. 2), were completed in 1994. Ground orientation and follow-up studies, an integral project component conducted in collaboration with other MDA projects, included ground spectrometry, mineralogical determinations, and multimedia geochemical analyses.

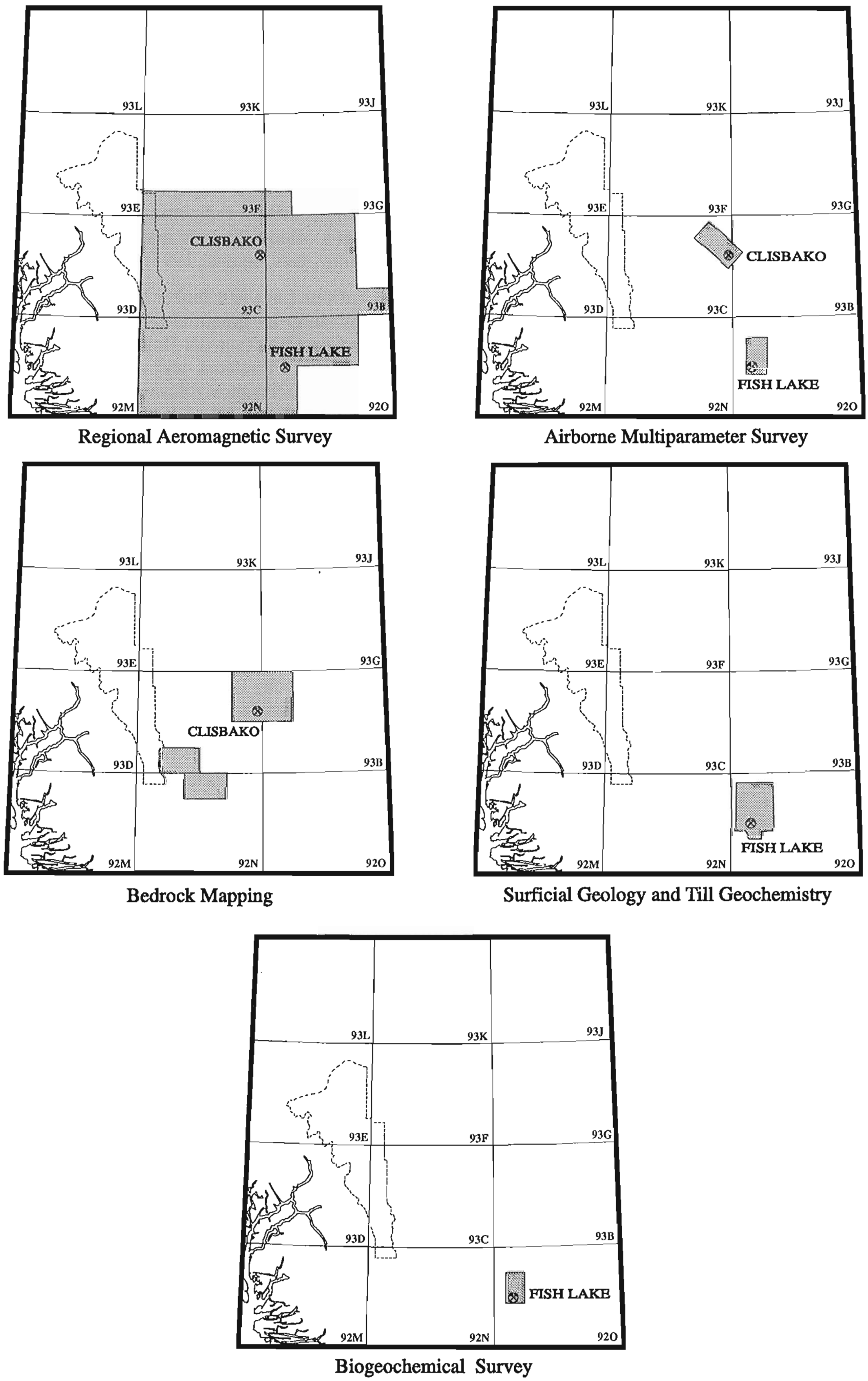

Figure 2. *Location of GSC projects in the Interior Plateau region.*

Gamma ray spectrometry has delineated anomalous radioelement concentrations (K, U, Th) in bedrock and surficial deposits at both Clisbako River and Fish Lake. These concentrations are attributed to mineralizing solutions that have altered the bedrock and have been reworked into the surficial sediments. The patterns of radioelement enrichment and depletion, in concert with magnetic and VLF-EM signatures, have already proven useful as a guide for mineral exploration (industry land tenure decisions, exploration and diamond drill targeting) and as a significant aid to bedrock and surficial mapping. The results indicate excellent potential for successful application of the gamma ray spectrometric technique elsewhere in the Interior Plateau region.

The airborne survey results for the Clisbako River and Fish Lake areas are available as digital data accompanied by "SurView", a viewing/display program, combined with a series of 1:150 000 scale colour contour maps and profiles of radioelement, magnetic and VLF-EM data (Shives and Carson, 1994; Shives and Rebelledo, 1994).

BEDROCK MAPPING

Geology of the Charlotte Lake-Junker Lake and Bussel Creek-Tatla Lake map areas

Geological mapping of the Charlotte Lake-Junker Lake (east-half) (NTS 93C/3,4 (east half)) and Bussel Creek (east-half)-Tatla Lake (NTS 92N/14 (east half),15) map areas (Fig. 2) was completed in 1994. This study has increased an understanding of the geology, tectonic evolution, and mineral potential of these areas. They contain known Cu-Mo (e.g., CQ showing, Minfile 093C-001,004) and Au-Ag mineralization (Perkins Peak prospect, Minfile 092N-010,011,012), and encompass a well preserved wilderness area ('Charlotte Alplands') which is currently included in a provincial land-use proposal as a sensitive development area (Commission on Resources and Environment, 1994). Preliminary results of this study were published in 1993 and 1994 (van der Heyden et al., 1993, 1994a; Mustard and van der Heyden, 1994); geological maps were released in 1994 (van der Heyden et al., 1994b; Mustard et al., 1994).

The Tatla Lake-Bussel Creek map contains the northern terminations of the East Waddington Thrust Belt (EWTB) and Tyaughton basin (van der Heyden et al., 1994a; Mustard and van der Heyden, 1994; Mustard et al., 1994). Foliated plutons and volcanics of the Jura-Cretaceous Coast Belt arc are thrust to the northeast over an Upper Triassic arc succession correlated with Stikine terrane. The arc rocks, in turn, are thrust over Cretaceous marine and nonmarine strata of Tyaughton basin [see Haggart (1995) for a report on Cretaceous fossil identifications for the Tatla Lake-Bussel Creek area], which were in part derived from a westerly, mixed volcanic-plutonic source. At the northern limit of the East Waddington Thrust Belt, the Coast Belt is thrust directly on the mid-Cretaceous nonmarine Silverquick Formation; marine strata of Tyaughton basin are absent, and the East Waddington Thrust Belt is intruded by the ca. 67-63 Ma Klinaklini and McClinchy plutons. The area northeast of Yalakom fault is underlain by the Eocene Tatla Lake metamorphic core complex (Friedman and Armstrong, 1988).

The northern end of Tyaughton basin, between the Yalakom and Tchaikazan faults, appears to be underlain by volcanic and sedimentary rocks that are intruded by at least two Late Triassic plutons, the 205 Ma Sapeye Creek pluton and the 212 Ma Niut pluton (Mustard et al., 1994). Preliminary examination of radiolaria from chert in this succession suggests a Middle Triassic age for these supracrustal rocks (F. Cordey, pers. comm., 1994).

Charlotte Lake and Junker Lake map areas are underlain predominantly by granitoid and metamorphic rocks of the eastern Coast Belt (van der Heyden et al., 1994b). The oldest rocks consist of screens and pendants of greenschist and amphibolite facies metavolcanic schist and gneiss, enclosed in variably foliated Mt. Ada diorite and quartz diorite and tonalite of the Wilderness Mountain pluton. Metarhyolite from the metavolcanic succession has yielded a ca. 190 Ma U-Pb date. The Mt. Ada diorite and Wilderness Mountain pluton have yielded U-Pb and Pb-Pb dates ranging from ca. 168 Ma to ca. 142 Ma. Steeply dipping foliations and ductile shear zones in the plutonic and metamorphic complex trend northeasterly, orthogonal to the regional trend of the Coast Belt.

Granodiorite to quartz monzonite of the Late Cretaceous McClinchy pluton intrudes the Wilderness Mountain pluton east of McClinchy Lake; elsewhere these two units are separated by a possible northerly extension of the Tchaikazan fault. Relatively unmetamorphosed Late Cretaceous volcaniclastic rocks of the Powell Creek formation are in fault contact with the older rocks north of Charlotte Lake. West of Charlotte Lake the Wilderness Mountain pluton is intruded by Early Tertiary granite, granodiorite and quartz monzonite of the Trumpeter pluton. The contact zone is marked by chalcocite, chalcopyrite, and bornite bearing quartz veins in an area that includes the Ada Cu showing (Minfile No. 093C-005). Abundant northeasterly, northerly, and northwesterly trending shear zones are present in the study area. Several of these appear to have localized hydrothermal activity related to the Tertiary intrusions. The Copper Queen Cu and C/DK Cu-Mo showing (Minfile No. 093C-001 and 093C-004) near McClinchy Lake may represent a fault-controlled, high structural level of a classic altered mineralized porphyry system hosted by the Wilderness Mountain pluton.

Eocene volcanic stratigraphy of the Clisbako River area

Reconnaissance mapping of the Clisbako River area (NTS 93C/9,16 and 93B/12,13) (Fig. 2) was completed in 1994. The area is characterized by a broad field of Eocene continental volcanic rocks and associated plutons, many of which appear to have been generated by crustal extension (Metcalfe and Hickson, 1994). Several former producing mines and scattered prospects throughout the Interior Plateau represent epithermal-type precious metal mineralization related to this important magmatic and tectonic episode (e.g., Blackdome, Wolf, etc.). The focus of this study is on elucidating the Eocene stratigraphy and geochemistry, and on providing regional correlations with Eocene volcanic rocks to the northwest and to the southeast of the study area.

The Eocene volcanic rocks host locally pervasive and potentially economic epithermal alteration and mineralization. They include two distinct, felsic to intermediate assemblages, which underlie a circular highland area, approximately 50 km in diameter and open to the southeast. Isolated outcrops of biotite-bearing pyroclastic rock in the centre and at the southeastern edge of the area are interpreted as the products of a large pyroclastic eruption. These Eocene volcanic rocks are interpreted as the erosional remnants of a large caldera that is partially filled with younger basaltic lavas of the Chilcotin Group (Metcalfe and Hickson, 1995). Open file maps and final reports will be released when completed.

SURFICIAL GEOLOGY AND TILL GEOCHEMISTRY

A surficial sediment sampling program in the Mount Tatlow (NTS 92O/5) and Elkin Creek (NTS 92O/12) map areas (Fig. 2) was completed in 1993. This project is part of a larger co-operative program with the BCGS that addresses regional surficial geochemistry and Pleistocene glacial stratigraphy, ice flow pattern indicators, and till lithologies in the Interior Plateau region. In 1992 the BCGS completed regional till sampling in the eastern sector of Anahim Lake map area (93C/1,8,16,19) (Giles and Kerr, 1993; Proudfoot, 1993). Final results of the GSC survey in 92O/5 and 92O/12 were released in 1994 (Plouffe and Ballantyne, 1994).

Till and glaciofluvial sediment samples were collected on exposures adjacent to roads, creeks, and rivers. The silt plus clay size fractions (<63 µm or -230 mesh) were analyzed for a suite of elements by ICP-ES and INA in commercial laboratories; a complete list of analyzed elements can be found in Plouffe and Ballantyne (1994) or van der Heyden et al. (1993). Computer generated proportional symbol maps for metals of economic interest are presented in Plouffe and Ballantyne (1994); the report includes brief notes on methodology, analytical quality control data, a basic interpretation of the anomalies, and a summary of the glacial history, with geochemical results and sample coordinates provided as ASCII files. The geochemical maps can serve to establish background concentrations of metals in the unconsolidated sediments, to determine baseline data for environmental assessments, and to define anomalous areas for further follow-up surveys.

During the 1994 field season, small wood fragments were collected from an organic rich layer intersected in drill core (Fish Lake Cu-Au deposit, Taseko Mines) underneath glacial till. These will be dated using ^{14}C accelerator mass spectrometry, and should provide information about the chronology of the last non-glacial interval in the Interior Plateau. Results of this work will include more detailed information about the style of glaciation of the area.

BIOGEOCHEMICAL SURVEY

A helicopter-assisted reconnaissance-level tree top geochemistry program in the Fish Lake area (92O/5,6,11,12,13,14) (Fig. 2) was completed in 1993. Sample collection was contracted to Pacific Phytometric Consultants (Surrey, B.C.), and involved the collection of Lodgepole pine (*Pinus contorta*) tops from 276 sites within an area of 1625 km^2 of wooded and drift-covered terrain. Sampling was undertaken during a three day period along 715 km of grid lines at a spacing of 2.5 km between sites (Dunn et al., 1994). Biogeochemical surveys at this scale (1 site per 6 km^2) provide a method of rapidly screening an area for regional geochemical trends. These trends can provide a focus for more detailed ground studies designed to pinpoint mineralized zones.

The woody tissues that comprise the tree tops were analyzed by ICP-ES and INA for approximately 50 elements. In general, element concentrations in tree tops from the survey area are low. However, there are regional trends in geochemical patterns and there are close spatial relationships among several elements characteristic of Au mineralization (i.e. Au, As, Cs, W, and Hg).

Follow-up ground biogeochemical studies in the Cone Hill area, north of Fish Lake, have involved the collection of outer bark from Lodgepole pine. Weak north-trending biogeochemical Au anomalies occur near IP and soil Au anomalies of similar trend. At Newton Hill (north of Scum Lake) there is a coincidence of Au and Se enrichment in Douglas-fir bark. Further analytical work in 1994, using the cones from the tree tops collected previously, has indicated Au enrichment near Scum Lake and Vedan Mountain. These biogeochemical data can be of use in detecting zones of mineralization covered by glacial drift, and may reflect metals that have migrated vertically through fractures in plateau basalts. The data should be viewed in conjunction with other data sets (e.g., gamma-ray spectrometry and till geochemistry) in planning exploration follow-up. Detailed results of the tree top survey, with element distribution maps and ASCII data files, were released in 1994 (Dunn et al., 1994).

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Geological Survey of Canada Project 920075

Durchbewegt sulphides, piercement structures, and gabbro dyke displacement in the vent complex of the Sullivan Pb-Zn deposit, British Columbia

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Paakki, J.J., Lydon, J.W., and Del Bel Belluz, N., 1995: Durchbewegt sulphides, piercement structures, and gabbro dyke displacement in the vent complex of the Sullivan Pb-Zn deposit, British Columbia; in Current Research 1995-A; Geological Survey of Canada, p. 81-90.

Abstract: Offsets of gabbro dykes indicate that relative movement between hanging wall and footwall of the Sullivan deposit was multi-directional and resulted in net movement of less than 150 m. These movements produced semi-concordant zones of high strain within the vent complex sulphides, which are marked by durchbewegung structure near the base of the sulphide body and by zones of foliated galena-rich sulphide within the interior of the deposit. Footwall cusps formed during compressional reactivation of north-trending fault structures that mark the margins of a pre-ore graben structure. Hanging wall piercement cusps and associated discordant durchbewegt sulphide are confined to the eastern margin of this structure. Faults of this graben structure focussed hydrothermal flow during the main ore-forming event. At the peripheries of the deposit, sulphides have been mechanically remobilized beyond their original extent.

Résumé : Le décalage des dykes de gabbro indique que le déplacement relatif entre le toit et le mur du gisement de Sullivan était multidirectionnel et qu'il s'est traduit par un déplacement net de moins de 150 m. Ces déplacements ont produit des zones semi-concordantes très déformées dans les complexes sulfurés d'évents qui sont marqués par une structure durchbewegung (broyée) près de la base de l'amas sulfuré et par des zones de sulfures feuilletés riches en galène à l'intérieur du gisement. Les croissants de l'éponte inférieure se sont formés durant la réactivation par compression des structures faillées à direction nord qui marquent les bords d'une structure en graben antérieure à la minéralisation. Les croissants perçant l'éponte supérieure et les sulfures durchbewegt discordantes associées se limitent à la marge est de cette structure. Les failles de cette structure en graben ont concentré l'écoulement hydrothermal durant l'épisode principal de minéralisation. En périphérie du gisement, les sulfures ont été mécaniquement remobilisés au-delà de leur étendue initiale.

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INTRODUCTION

Durchbewegung and piercement structures are common deformation features in many metamorphosed and deformed massive sulphide deposits (e.g., Vokes, 1969, 1973; Maiden et al., 1986; Gilligan and Marshall, 1987; Klemd et al., 1987; Marshall and Gilligan, 1989). Durchbewegung structure is a record of the internal deformation of sulphide deposits, whereas piercement structures are irregularities at sulphide - wall rock interfaces (e.g., Marshall and Gilligan, 1989).

Durchbewegung, a German term introduced by Vokes (1969), refers to the progressive tectonic process of detachment, dismemberment, fragmentation, rotation and milling of competent wall rock or vein material in a flowing mass of ductile sulphide (Fig. 1a). Under the elevated temperatures and pressures of metamorphism, the strength of galena, chalcopyrite, pyrrhotite and sphalerite are very much less than either pyrite or quartzose rocks (Marshall and Gilligan, 1989). Durchbewegt sulphides (i.e., sulphide rock exhibiting durchbewegung structure) commonly consists of clasts pyrite and silicate rocks in a matrix of less competent sulphides and is indicative of extreme deformation in sulphide rocks (Vokes, 1973). Durchbewegung structure may be developed in zones of intense ductile shearing or folding and where incompetent material invades competent rock undergoing brittle failure (Marshall, 1988).

Piercement structures include cusps, piercement cusps and piercement veins (Fig. 1b). Following the terminology of Marshall and Gilligan (1989), cusps are pinched-in folds that occur at the interface between rocks of highly contrasting competency. Cusps consistently point into the more competent material and are separated by oppositely directed lobes. Layering in the material that forms the lobe maintains its integrity around the cusp (Fig. 1b(i)). Piercement cusps are vein-shaped bodies projecting from cusps and may transgress layering (Fig. 1b(ii)). Piercement veins are vein-shaped bodies of incompetent material that transgress layering of competent material but have no spatial association with cusps (Fig. 1b(iii)). Cusps and piercement cusps commonly occur at interfaces between silicates and massive sulphides, and form during layer-parallel to layer-oblique compression (Sokoutis, 1987; Marshall, 1988). Cusps progressively develop into piercement cusps with increased shortening. Piercement veins, however, are products of extension and are formed by the flow of sulphides into dilatent fractures of competent wall rocks (Marshall, 1988).

Sulphides of the Sullivan deposit have been affected by post-metamorphic horizontal compressive stresses (Campbell et al., 1980) which can be divided into three main phases of deformation (McClay, 1983). The bulk of the deformation in sulphide rock documented by McClay (1983) consists of east-verging isoclinal folds and west-dipping, low-angle thrust faults, which resulted from the relative movement of hanging wall over footwall along décollements in the sulphide body. McClay's (1983) observations were based on the bedded ore sequence and the transition zone (see below), where prominent compositional layering facilitates the recognition of fold and fault patterns. The overall lack of distinctive compositional layering within the massive sulphides of the vent complex makes the ready recognition of megascopic fold and fault patterns more difficult than in the bedded ore sequence.

This paper documents the distribution of durchbewegung and piercement structures in the vent complex sulphides of the Sullivan deposit. Recognition of these structures provides a basis for interpretation of the larger scale deformational and structural features of the orebody. Displacements of pre-deformational crosscutting gabbroic dykes provide reliable kinematic indications of orebody translation. Observations and interpretations presented here are based on drill core

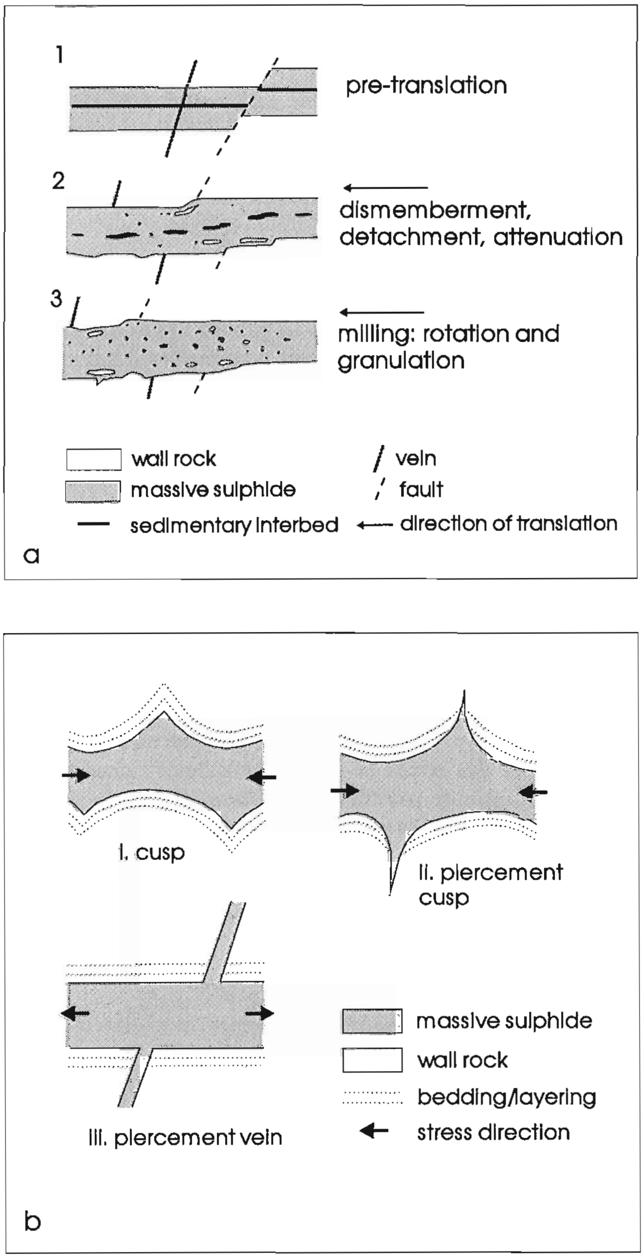


Figure 1. *Schematic illustrations of: (**a**) the development of durchbewegung structure, and (**b**) piercement structure terminology (after Marshall and Gilligan, 1989).*

logging, underground examinations and compilation of Cominco's (Cominco Metals Limited) mine sections and maps.

GENERAL GEOLOGY OF THE VENT COMPLEX

The overall architecture of the Sullivan deposit has been described by Hamilton et al. (1982). In brief, the deposit can be divided into the massive ores of the vent complex in the west and a bedded ore sequence to the east. Vent complex sulphides are grossly layered, and consist of an upper economic ore zone of interlayered galena-sphalerite-pyrrhotite massive sulphides and lithic units; a middle economic zone of crudely layered (foliated) galena-pyrrhotite-sphalerite; and a lower uneconomic zone of massive pyrrhotite-rich sulphide (Fig. 2). Layering and foliation in sulphides is generally conformable to bedding in the host rocks. Massive sulphides are underlain by tourmalinized semi-comformable (footwall conglomerate) and discordant ("chaotic breccia") fragmental sedimentary rocks, which are crosscut by pyrrhotite-rich stringer sulphides. Within the central part of the vent complex, hanging wall and ore sequences are replaced by pyrite-chlorite-carbonate-albite alteration assemblages (Hamilton et al., 1982; Shaw et al., 1993).

Vent complex sulphides grade into bedded ores within a structurally complex zone of intense faulting, folding and rapid changes in thickness known as the "transition zone" (Hamilton et al., 1982; McClay, 1983). Stratiform hanging wall sulphides are strongly developed above this zone (Fig. 2). To the west and south, the sulphide body attenuates through a series of abrupt thickness changes. It is in these peripheral zones of thinner sulphide that durchbewegung structure is most prominently developed. Towards the north, sulphides are cut off by the Kimberley Fault.

The Sullivan deposit is underlain by a gabbroic sill complex from which apophyses cut across the sulphide deposit (Hamilton et al., 1982; Turner and Leitch, 1992). Gabbro intrusions predate metamorphic and deformation events (McClay, 1983).

DISTRIBUTION AND DEVELOPMENT OF DURCHBEWEGT SULPHIDES

The term durchbewegt sulphide here refers to all sulphide assemblages exhibiting durchbewegung structure. Forty-one drill holes, that intersected the complete sulphide interval were logged to assess the gross-scale distribution of durchbewegt sulphide in vent complex sulphides.

Durchbewegt sulphides in the vent complex consist of variable proportions of randomly oriented, rounded to angular clasts of altered and unaltered sedimentary rock, carbonate, quartz, sphalerite, and minor pyrite and gabbro within a sulphide matrix consisting of pyrrhotite and galena (Fig. 3A-D). Sphalerite-dominant matrix is uncommon. Clasts vary in size from less than a millimetre to several metres in diameter, with the larger dimensions being restricted to clasts of sedimentary rock (see Fig. 5, 9). Rounded to angular clasts of pyrite and rare clasts of magnetite occur in durchbewegt sulphides near the southern end of the transition zone and in the open pit area, whereas these minerals form massive layers in the adjacent bedded ore sequence to the east.

Vent complex sulphides with the highest relative proportion of durchbewegt sulphide occur in the thinner sulphides of the west fringe and the open pit area in the south (Fig. 4a). The predominance of durchbewegt sulphide in these flanking areas implies that most of the sulphides at the periphery of the deposit have been tectonically transported.

In these flanking areas, clasts in durchbewegt sulphides consist mainly of unaltered sedimentary rock, and carbonate and quartz vein material. The example from the west fringe

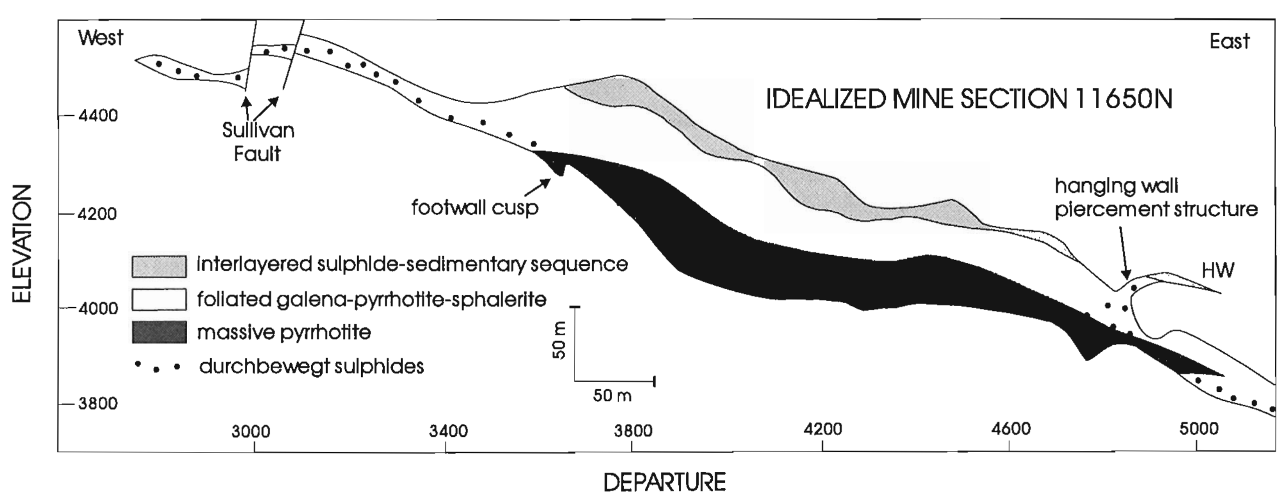
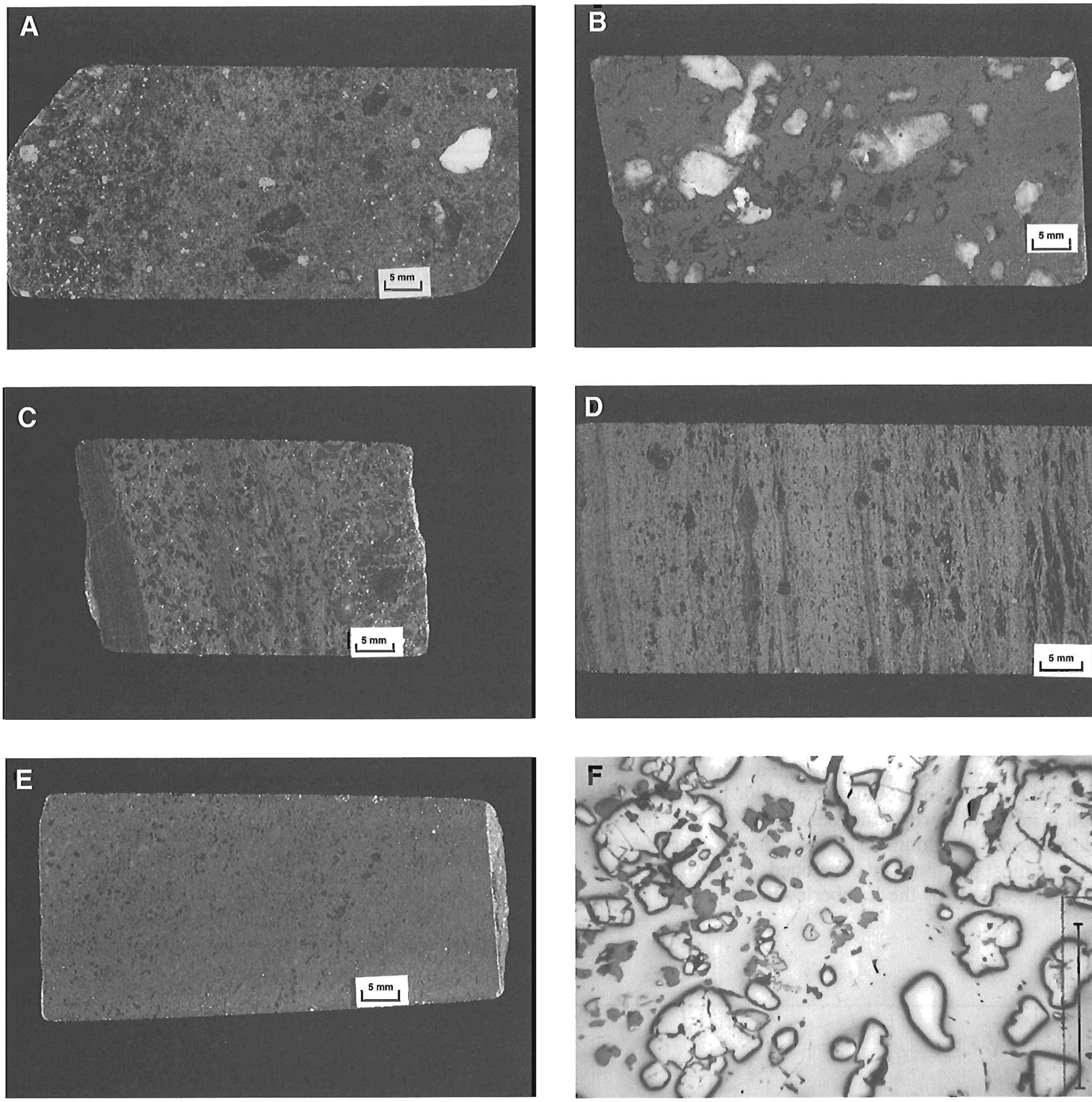


Figure 2. Idealized geological cross-section of the Sullivan vent complex illustrating the distribution of durchbewegt sulphides and piercement structures. HW denotes stratiform hanging wall sulphides. Mine scale is in feet.



A) Polished drill core from the open pit area. Quartz (bright white), argillite (black), pyrite (off-white) and sphalerite (medium grey) clasts in a pyrrhotite matrix.

B) Polished drill core from the transition zone. Quartz clasts (white) enclosed in a matrix of galena and pyrrhotite.

C) Polished drill core from the west fringe area. Carbonate clasts (grey to black) in a weakly foliated, galena-dominant matrix (steel grey). Massive dark grey area to the left of photograph is sphalerite.

D) Polished drill core from the west fringe area. Chloritic argillite (black), sphalerite (medium grey) and garnet (well rounded) clasts in a pyrrhotite matrix.

E) Polished drill core from the basal part of massive pyrrhotite zone. Granulated carbonate (dark grey) in a pyrrhotite-dominant matrix.

F) Photomicrograph of durchbewegt sulphide at the margin of the pyrite core. Reflected light (plane polarized light). Pyrite (high relief mineral), sphalerite (dark grey) and pyrrhotite (off-white) clasts set in a matrix of galena. Vertical bar is 1mm.

Figure 3. *Photographs of typical durchbewegt sulphides in the Sullivan vent complex.*

illustrated in Figure 5 suggests that the lithic clasts were derived from sedimentary rock layers within the sulphide body rather than from the footwall or hanging wall. This may indicate that sulphides of the west fringe areas originally contained interbedded sedimentary rock layers similar to the bedded ore sequences in the eastern part of the deposit. However, without detailed study of the rock face to gather supporting evidence, tabular clasts in sulphides cannot be unequivocally identified as being derived from bedded interlayers because they may also be wall rock detachments.

The greatest thickness of durchbewegt sulphide occurs at the base of the pyrrhotite zone in the north-central part of the vent complex (Fig. 4b). Here, the prominent clast type is carbonate (Fig. 3E), though Hamilton et al. (1982) also

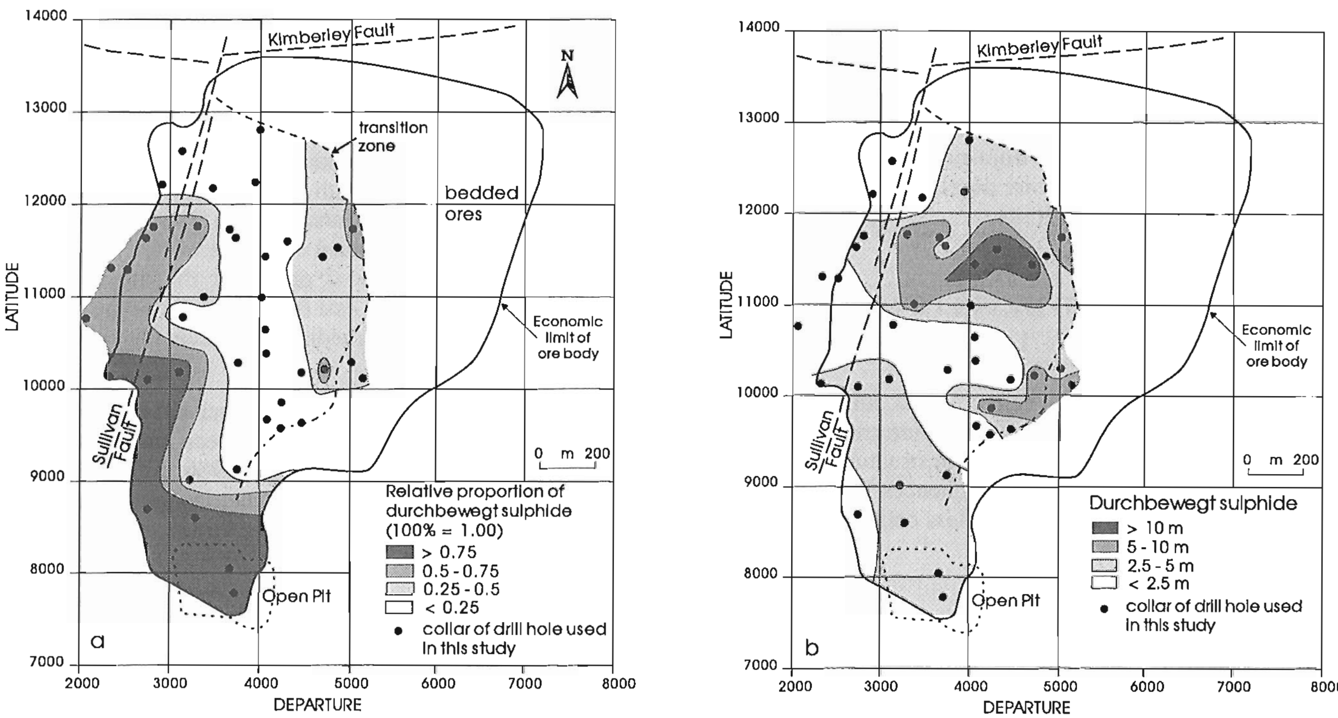


*Figure 4. Plan map of the Sullivan deposit showing the distribution of durchbewegt sulphide in the vent complex; (**a**) the proportion of durchbewegt sulphide relative to the entire sulphide interval, and (**b**) thickness of durchbewegt sulphide intersected in drill core. Mine grid is in feet.*

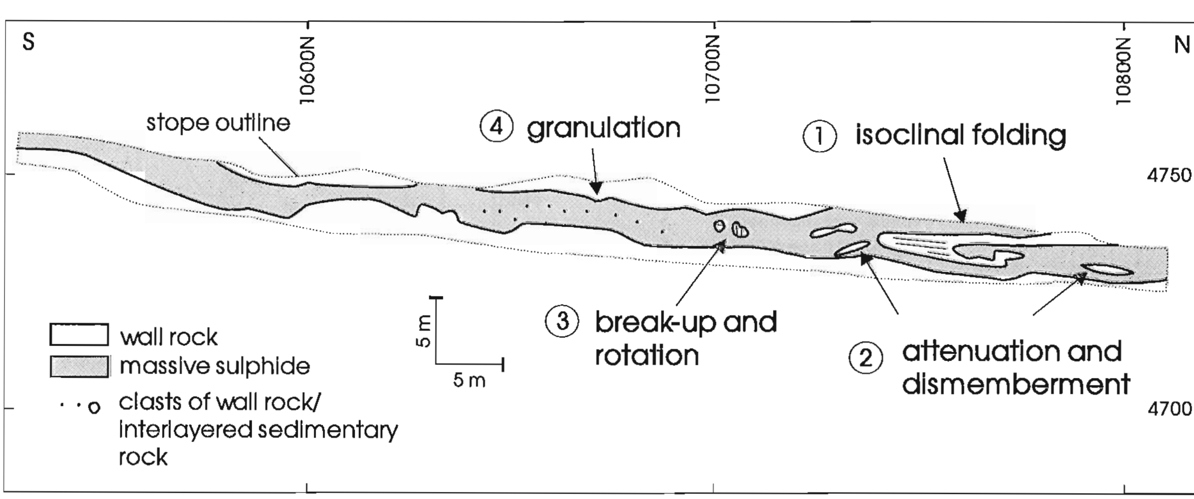


Figure 5. Underground map of O-2-30 #9 stope showing the progressive development of durchbewegung structure in the west fringe area (after a sketch in unpublished Sullivan mine records). Mine scale is in feet.

described fragments of quartz veins and sedimentary rocks. Tourmalinite, chlorite-altered sedimentary rock, and vein quartz clasts are relatively minor clast types. Rare large rafts of sedimentary rock in R10-30 Stope, near the base of the pyrrhotite zone, are assumed to be footwall detachments.

Discordant zones of durchbewegung structure accompany piercement cusps along the transition zone. Here, interlayered sedimentary and wall rock were incorporated as clasts during sulphide transgressions into hanging wall sequences (see Fig. 9).

Durchbewegt sulphides also occur at the base and lateral margins of the pyrite-chlorite-carbonate alteration zone. At the margins of this alteration zone, pyrite clasts are enclosed in a galena-dominant matrix (Fig. 3F), whereas at its base, pyrrhotitic durchbewegt sulphide contains rare clasts of tourmalinite.

DISTRIBUTION AND DEVELOPMENT OF PIERCEMENT STRUCTURES

Piercement structures in the vent complex sulphides include cusps and piercement cusp structures. Piercement veins have not been recognized by us. Piercement structures project into both footwall and hanging wall sequences (Fig. 2, 6, 8, 9), though hanging wall structures are less common.

Footwall cusps

Footwall cusps are included in the definition of "keel" structures described by Hamilton et al. (1982) who used the term to describe downward projections of sulphides at the base of the sulphide zone, including open folds or flexures in sulphide rock along the footwall contact. Only footwall projections with a cuspate form (i.e., pointed) are included in this study.

Underground access to the vent complex sulphides is now extremely limited, preventing a direct re-examination of the cusps. Mine sections and stope maps show that cusps vary from sharp downward projections to irregularly-shaped cuspate forms with amplitudes up to several tens of metres (see Fig. 8). Hamilton et al. (1982) reported that internal layering of sulphides within keel structures is generally conformable to their contacts with wall rock, which is consistent with the criteria of Marshall and Gilligan (1989) for diagnosis of cusps. Axial planes of most cusps are westerly dipping to near vertical, whereas axial planes of the most westerly occurring cusps dip to the east (see Fig. 6, 8). Some sulphide sequences overlying or immediately adjacent to cusps may be contorted.

Cusp structures form during layer-parallel or layer-oblique shortening of the interface between materials of contrasting competency (Maiden et al., 1986; Sokoutis, 1987; Marshall and Gilligan, 1989). Symmetric cusps form under layer-parallel compression (Sokoutis, 1987), whereas asymmetric cusps result from layer-oblique compression

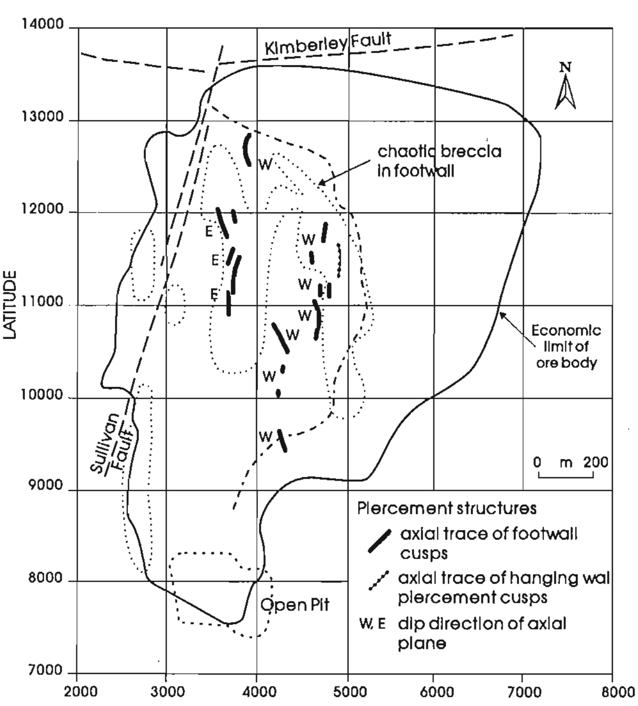


Figure 6. Plan map of the Sullivan deposit showing the distribution of piercement structure and discordant footwall chaotic breccia zones in the vent complex. This map is based on compilation of Cominco's mine sections and Hamilton et al. (1982). Mine grid is in feet.

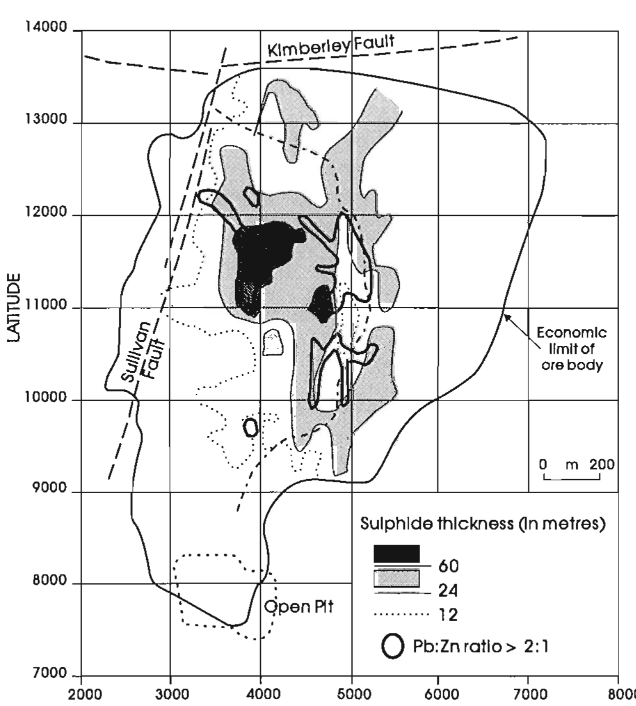


Figure 7. Plan map of the Sullivan deposit showing thickness variations of sulphides (exclusive of hanging wall sulphides) and areas of high Pb:Zn ratio in the vent complex (simplified from Hamilton et al., 1982). Mine grid is in feet.

(e.g., Maiden et al., 1986). McClay (1983) concluded that the bulk of the internal deformation of the Sullivan deposit resulted from a continuum of east-verging folding and thrust faulting events. The form and orientation of most footwall cusps are consistent with this type of deformation. Footwall cusp development begins from open folds or flexures and progresses into definitive cusp structures with increased shortening. Sokoutis (1987) suggested that the initial process of cusp development begins at irregularities along the interface between materials of contrasting strengths. As discussed below, it is significant to the understanding of the architecture of the Sullivan deposit, that footwall cusps are localized along prominent north-trending structures (see Fig. 6, 7).

Hanging Wall Piercement Cusp Structures

Within the vent complex, hanging wall sulphide piercement structures are restricted to the easternmost part of the transition zone (Fig. 2, 6). An interpretive section along 11550N shows the complexity of the piercement cusp structure (Fig. 9). The east-verging piercement cusp structure consists of several crosscutting, vein-like sulphide projections or flame structures. Numerous clasts of altered sedimentary rock, quartz and carbonate are incorporated in the transgressive sulphides and in underlying sulphide rock of the main ore layer, forming typical durchbewegung structures. Some discordant flame structures terminate upwards as layer-conformable sulphide bodies that form an apparent continuum with stratiform hanging wall sulphides.

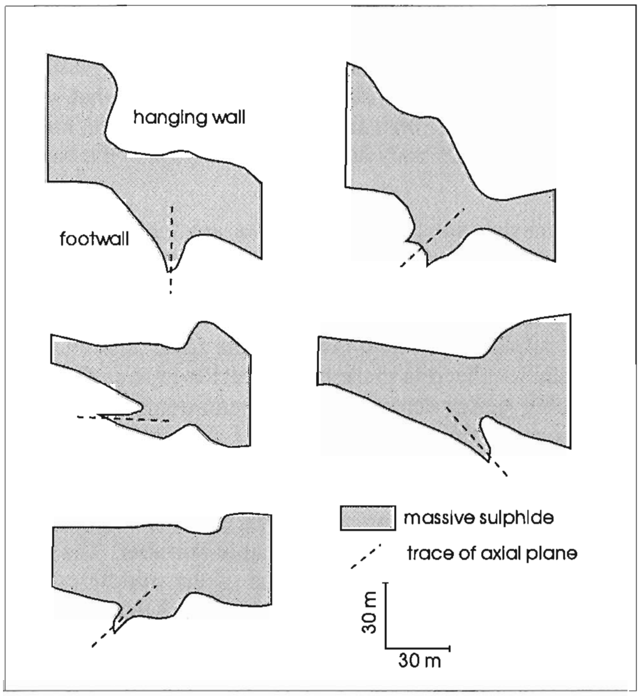


Figure 8. Schematic illustrations of footwall cusps. All sketches are looking north.

The development of hanging wall piercement cusps preferentially in the transition zone can be attributed to several factors. The transition zone is unique in that it not only marks the change from dominantly massive vent complex sulphides in the west to thinner, bedded sulphides in the east, but also marks the outer boundary of highly competent footwall tourmalinite alteration. The changes in sulphide thickness coupled with the juxtaposition of highly contrasting footwall sequences rendered the transition zone more susceptible to imposed deformation than the remainder of the Sullivan orebody (McClay, 1983). Concentration of piercement cusp structures in the transition zone is also aided by its relatively high concentrations of galena (see Fig. 7). The galena-rich sulphides provided the high ductility and extreme contrasts in mechanical strength between sulphides and wall rock needed for piercement.

GABBRO DYKE DISPLACEMENTS

Gabbro dykes are generally steeply dipping and consistently trend in a north-westerly direction. They pre-date deformational events (McClay, 1983) and crosscut the sulphide stratigraphic sequence (Hamilton et al., 1982; Turner and Leitch, 1992). Within the vent complex, they are most abundant along a zone that transects the pyrite-chlorite-carbonate alteration zone but also occur in the northern part of the vent complex near the transition zone (Fig. 10). Most dykes terminate at both the footwall and hanging wall contacts of sulphide rock, their continuation through the sulphide zone being marked by a trail of dismembered gabbro dyke segments.

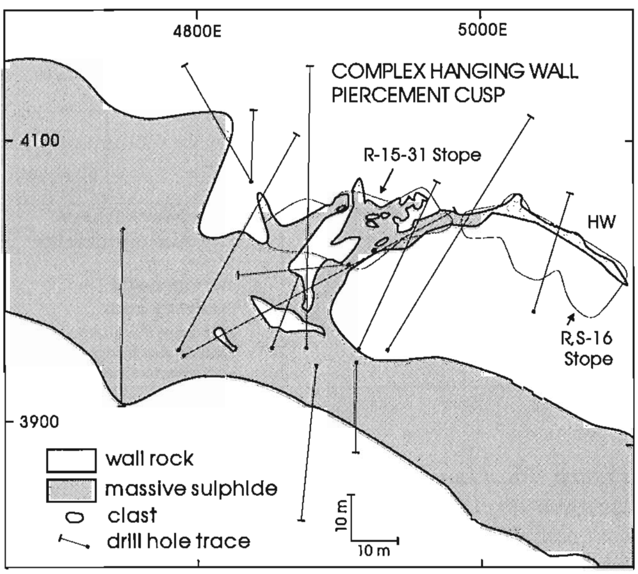


Figure 9. Interpretive cross-section of complex hanging wall piercement cusp structure (mine section 11550N), based on drill core and underground stope maps. HW denotes stratiform hanging wall sulphides. Mine scale is in feet.

Figure 10 shows the vertical projection of gabbro occurrences in the sulphide body and immediately adjacent hanging wall and footwall, as recorded on drill hole logs and stope maps. In the northern part of the vent complex the maximum aggregate displacement of dykes is approximately 150 m southwards. In the western part of the vent complex, hanging wall translation is less than 100 m to the northeast. Here, gabbro dyke segments in the sulphide zone have been transported for greater distances than offsets of the dykes, indicating that there has been a mass flow of sulphides to the northeast relative to hanging wall. To the south and east of the pyrite-chlorite-carbonate alteration zone, displacement is minimal but locally hanging wall has been translated up to 50 m to the southwest. These disparate directions of apparent net hanging wall translation over footwall sequences indicates that there were zones of alternating extension and shortening of the hanging wall with respect to the footwall, which in a general sense has resulted in a clockwise rotation of the hanging wall with respect to footwall.

McClay (1983) suggested a hanging wall displacement of some 300 m to the northeast based on the assumption that the centre of the hanging wall massive albitite alteration correlates with the centre of the pyrite core in the sulphide zone

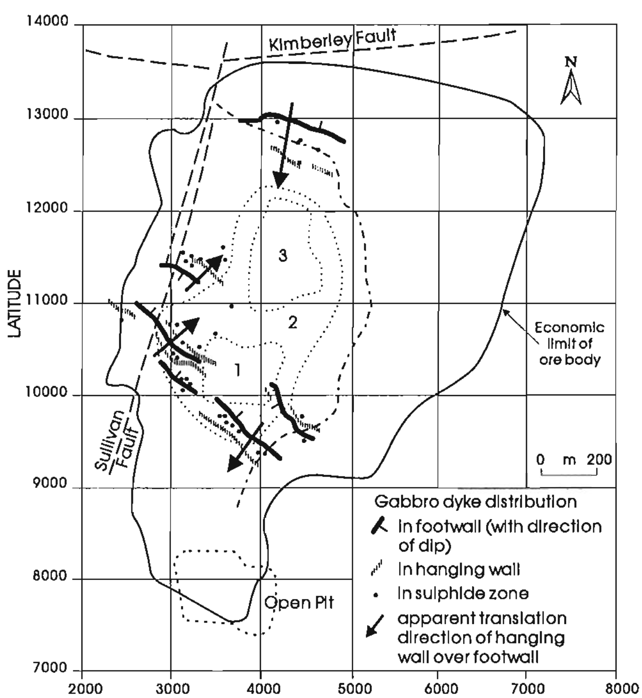


Figure 10. Plan map of the Sullivan deposit showing apparent displacement of crosscutting gabbro dykes in the vent complex. This map is based on compilation of gabbro intersections in drill core and underground stope maps. Gabbro dykes in the footwall include only those in direct contact with the sulphide body. Alteration zones shown are from Hamilton et al. (1982): 1) pyrite-chlorite-carbonate in the sulphide zone (pyrite core); 2) albite-chlorite-pyrite-carbonate in the hanging wall; 3) massive albitite in the hanging wall. Mine grid is in feet.

(Fig. 10). However displacement of gabbro dykes, which provide a more reliable kinematic indicator of translation, indicates a maximum net hanging wall translation of less than 150 m, which in the area of the massive albitite core is towards the southwest quadrant. This would suggest that the pre-displacement upward continuation of the pyrite core is in the albite-chlorite-pyrite-carbonate rock of the hanging wall rather than the massive albitite as suggested by McClay (1983).

DISCUSSION

Durchbewegt sulphides in the Sullivan vent complex are largely confined to the basal part of the pyrrhotite zone and the attenuated massive sulphides in the west fringe and the southern open pit area. This distribution indicates that the base of the vent complex massive sulphides is a high strain zone, representing a zone of décollement for the translation of the hanging wall over the footwall. Macroscopic durchbewegung structure indicates that high strain occurred within a few metres of a sulphide-wall rock interface, from which most lithic clasts have been derived. With increasing distance from wall rock, lithic clasts are progressively comminuted (by milling) to sub-millimetre size, and the highly strained sulphide loses its megascopic durchbewegung structure.

The absence of durchbewegung structure away from the basal zone does not necessarily imply the absence of high strain zones. Although preserved interlayered sedimentary and sulphide rock sequences in upper parts of the vent complex ores reflect a lower strain domain, the middle economic zone of crudely layered sulphides (Fig. 2) is highly foliated and is obviously highly strained. This indicates that semi-concordant zones of high strain also occur within the middle part of the sulphide body and are not restricted to the base of the deposit.

Galena-rich sulphide assemblages are the most ductile sulphide rocks in the Sullivan deposit. Galena constitutes more than 50 per cent of the sulphide minerals over intervals of up to several metres within the middle zone of crudely layered sulphide, and semi-conformable zones of high strain have been localized in these highly ductile layers. The most convincing demonstration of these semi-conformable high strain zones is the en echelon train of dismembered gabbro dyke segments between the hanging wall and footwall intercepts of the dyke. Another example is the truncation of steeply inclined contours of chemical concentrations and ratios of the sulphide rock along semi-conformable surfaces. The most spectacular of these are at the base of the middle crudely layered economic sulphide zone, where high grade foliated ore frequently is in abrupt contact with massive barren pyrrhotite. Interbedded sulphide-sedimentary rock sequences in the upper part of the vent complex have been deformed by folding and likely bedding-parallel shearing, similar to the bedded sulphide sequences in the eastern part of the deposit (McClay, 1983).

The movement of dismembered gabbro dyke segments within the sulphide interval for greater distances than dyke offsets between footwall and hanging wall sequences (Fig. 10) suggests a mass flow of sulphides independent of the amount of hanging wall movement. In the western part of the deposit this flow is dominantly towards the thickest part of the vent complex, and suggests that a component of tectonic thickening of the sulphide body has taken place. In the southern part of the orebody, the apparent direction of sulphide flow was toward the periphery of the deposit, which, coupled with the predominance of durchbewegt sulphide (Fig. 4a) suggests that most of the sulphide in this area has been tectonically transported. It is therefore very probable that the sulphides of the open pit area are a tectonic remobilization beyond their pre-deformation extent. On a larger scale, the predominance of durchbewegt sulphides at both the western and eastern margins of the Sullivan deposit may mean that in these areas also, sulphide has been remobilized outwards from their original boundaries. Recognition of this possibility is important to interpretation of the controlling factors on the extent of the original Sedex (sedimentary exhalative) deposit.

It is suggested here that the concentration of cusps along the eastern and western margins of the area that contains the major chaotic breccia zones, represents propagation from fault-controlled irregularities in the sulphide-footwall interface coincident with the bounding structures of the pre-ore dilatant zone. The axial planes of cusps dip toward the locus of maximum sulphide thickness in the central part of the vent complex (Fig. 6, 7). This geometry suggests that the cusps were formed by compressional reactivation of north-trending normal faults that formed an original graben structure, and within which there was a maximum primary accumulation of sulphide. The existence of such a pre-sulphide north-trending structure is indicated by the direction of elongation of the chaotic breccia zones and the linearity of their contacts (implying fault control) (Fig. 6). Chaotic breccia bodies represent a pre-ore hydraulic disruption of the footwall (Hamilton et al., 1982) which indicates that the faults were therefore in existence at the onset of ore deposition. The interpretation of a north-trending graben structure within the vent complex is consistent with the interpretation of Hamilton et al. (1982) of pre-ore synsedimentary extensional faults that produced a seafloor depression, on the scale of the entire Sullivan deposit, and that of Turner et al., (unpublished abstract, GSC 1994 Minerals Colloquium) that the larger scale Sullivan Corridor, a zone of disrupted and hydrothermally altered sediments that extends for over 7 km southwards from the Sullivan deposit, represents a synsedimentary graben.

The greatest thickness of durchbewegt sulphide is also bounded by this graben structure (see Fig. 4b). As mentioned above, the dominant clast type in durchbewegt sulphides at the base of the pyrrhotite zone in this area is coarsely crystalline carbonate (calcite), presumably of hydrothermal origin. There is a spatial association of calcite with pyrrhotite throughout the Sullivan deposit, of which the basal part of the vent complex pyrrhotite zone is an example, that seems distinct from the calcite associate with pyrite-chlorite alteration and gabbro dykes. It is suggested here that the concentration of carbonate clasts in durchbewegt sulphides at the base of the pyrrhotite zone represents the comminution of a network of calcite veins. If this suggestion is correct, the graben structure below the vent complex acted as a focus for hydrothermal upflow during this carbonate depositional stage of hydrothermal activity. Fault-controlled irregularities in a sulphide-wall rock interface form a prime site for detachment of wall rock to form clasts in durchbewegt sulphides (see Fig. 1a). The overall lack of sedimentary rock clasts in the thickest sections of durchbewegt sulphides is attributed to the high strength of footwall tourmalinite, which likely resisted detachment during deformation.

The margins of this graben structure are coincident with the areas of highest Pb:Zn ratios (Fig. 7). In Sedex deposits, the highest Pb:Zn ratios characteristically indicate the focus of hydrothermal upflow during the main stage of ore deposition (e.g., Lydon, 1983). Faults bounding the north-trending graben structure below the vent complex therefore appears to have also been the main upflow conduits for the ore-forming hydrothermal fluids, as well as for the carbonate-enrichment stage (if they were separate events). In this connection, the distribution of hanging wall ores along the eastern margin of graben structure might indicate that the bounding faults also acted as a conduit for hydrothermal upflow to form the hanging wall sulphides. However, this conjecture is obscured by the fact that this same lineament is also the locus of hanging wall piercement structures (Fig. 6), which has resulted in a continuity of galena-rich durchbewegt sulphides between the main sulphide zone and the hanging wall sulphides by the mechanical transfer of sulphides from the former to the latter. The hanging wall sulphides therefore could be of hybrid origin.

CONCLUSIONS

Durchbewegung and piercement structures within the vent complex sulphide body resulted from multi-directional translations of hanging wall over footwall sequences. Semi- concordant high strain zones occur along the base of the sulphide body, where they produced durchbewegung structure, and within overlying zones of galena-rich sulphides, where they produced highly foliated planar fabrics. The upper parts of the vent complex, where interlayered sedimentary and sulphide sequences are preserved, are least strained.

The small lateral aggregate displacement of hanging wall with respect to the footwall sequences (< 150 m) in comparison to the overall size of the Sullivan deposit (2000 m by 1500 m), means that the overall morphology of the deposit has probably been preserved. Mass flow of sulphides within the deposit has blurred details of the deposit architecture and, together with metamorphic recrystallization, obliterated smaller scale textural features. This mechanical remobilization of sulphides has resulted in sulphides now occurring beyond their original lateral extent at the peripheries of the deposit, and contributed to thickening in the central part of the vent complex.

Cusps developed at the interface between massive sulphide and tourmalinized footwall, propagating from north-trending, fault-controlled irregularities along the footwall-sulphide interface during compressional reactivation of north-trending faults. These faults formed the margins to a pre-ore graben structure, which has exerted a fundamental control on the evolution of the vent complex. This graben structure localized the footwall chaotic breccias and maximum sulphide accumulation, and also focussed hydrothermal upflow during the main ore-forming event and carbonate enrichment of the pyrrhotite zone. The bounding faults on the eastern margin of this graben also controlled the siting of hanging wall ores by localizing sulphide injection into the hanging wall or by focussing late stage hydrothermal upflow.

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Geological Survey of Canada Project 930046

A note on U-Pb dating of Middle Jurassic plutonic suites: Cumshewa Head pluton, southeastern Moresby Island, Queen Charlotte Islands, British Columbia

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Anderson, R.G. and McNicoll, V.J., 1995: A note on U-Pb dating of Middle Jurassic plutonic suites: Cumshewa Head pluton, southeastern Moresby Island, Queen Charlotte Islands, British Columbia; in Current Research 1995-A; Geological Survey of Canada, p. 91-96.

Abstract: A new U-Pb date of 167 ± 2 Ma for the Cumshewa Head pluton corroborates the pluton's previous correlation with the Middle Jurassic (168 to ≥ 158 Ma) Burnaby Island plutonic suite based on plutonic style.

Résumé : ne nouvelle datation U-Pb de 167 ± 2 Ma pour le pluton de Cumshewa Head corrobore la corrélation antérieure du pluton avec la suite plutonique de Burnaby Island du Jurassique moyen (de 168 à ≥ 158 Ma), laquelle est basée sur le style plutonique.

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INTRODUCTION

Jurassic plutonic and volcanic rocks form an important igneous province in the Queen Charlotte Islands (QCI) and they were important in the development of the Queen Charlotte Basin. Knowledge of the distribution, nature, age, thermal regime and tectonic setting of the magmatism contributes to the understanding of their genesis and ultimately in evaluation of the hydrocarbon potential of the basin. For example, thermal overmaturation of potential hydrocarbon source rocks of the Kunga Group, though cospatial with Tertiary and Middle Jurassic plutons on southeastern Moresby Island (Vellutini and Bustin, 1991; Orchard and Forster, 1991), is genetically most closely linked with the Middle Jurassic (168 to ≥ 158 Ma) Burnaby Island plutonic suite (Anderson and Greig, 1989; Anderson and Reichenbach, 1991).

GEOLOGICAL SETTING

Jurassic igneous rocks extend the length of the Queen Charlotte Islands. Middle Jurassic (172-≥158 Ma) plutons consist of the San Christoval plutonic suite (172-171 Ma; San Christoval plutonic suite) on the west coasts of Queen Charlotte Islands, and to the east, the Burnaby Island plutonic suite (Burnaby Island plutonic suite), the younger (168-≤158 Ma) of two post-tectonic Middle Jurassic plutonic suites (Fig. 1; Anderson, 1988a,b; Anderson and Greig, 1989; Anderson and Reichenbach, 1989, 1991; Anderson et al., 1992). The sample we dated (AT-87-61-2) was collected from the Cumshewa Head pluton of the Burnaby Island plutonic suite (Anderson and Greig, 1989; Anderson and Reichenbach, 1991). The suites are part of the Middle and Upper Jurassic assemblage of Lewis et al. (1991) which evolved after an important Aalenian deformation.

Middle Jurassic Burnaby Island plutonic suite

Geology and intrusive relationships

Burnaby Island plutonic suite is more heterogeneous than San Christoval plutonic suite, comprising sequentially intruded gabbro or diorite, quartz monzodiorite, quartz monzonite and trondhjemite (leucodiorite) phases. Rocks are typically unfoliated and pervasively brittly fractured and veined. Hornblende is subhedral and as abundant as biotite except in felsic phases (such as in the Cumshewa Head pluton), where biotite is the main mafic mineral. The trondhjemite phase lacks hornblende but contains muscovite and rare biotite.

Burnaby Island plutons crosscut Middle Jurassic (Bajocian) and older strata. Locally, Lower Cretaceous (Hauterivian) Longarm Formation rocks (Haggart and Gamba, 1990) nonconformably overlie the highly altered and veined granitic rocks (Anderson and Greig, 1989). Geological relations between the Cumshewa Head pluton and Bajocian Yakoun Group feldspar-porphyry volcanic rocks are poorly exposed but may indicate that the pluton is younger. Hydrothermal alteration and veining of the plutons, which also distinguish Burnaby Island plutonic suite from San Christoval plutonic suite and are cospatial with copper-iron skarns (Anderson, 1988b), are much more regionally extensive, intense and regular in the granitic rocks than in the overlying Longarm Formation.

Previous K-Ar and U-Pb isotopic dating

U-Pb (Anderson and Reichenbach, 1991) and K-Ar (Anderson and Reichenbach, 1989, 1991; Hunt and Roddick, 1990, 1991) dates for members of the Burnaby Island plutonic suite suggest that the suite intruded between 168-≥158 Ma. Generally the K-Ar dates agree within uncertainty with the U-Pb dates. Exceptions are: 1) the relatively late trondhjemite phase which has the youngest K-Ar and U-Pb dates (153 ± 3 Ma and ≥ 158 ± 4 Ma respectively); and, 2) the latest Jurassic K-Ar dates (147-145 Ma) attributable to a pre-Early Cretaceous hydrothermal event (Anderson and Reichenbach, 1991). The Cumshewa Head pluton was not precisely dated but preliminary U-Pb analyses from sample AT-87-61-2, (analysis for fraction A* in Table 1 and Fig. 2), reported in Anderson and Reichenbach (1991) helped corroborate its field relations that suggested it was part of the Burnaby Island plutonic suite and not the Tertiary Kano plutonic suite.

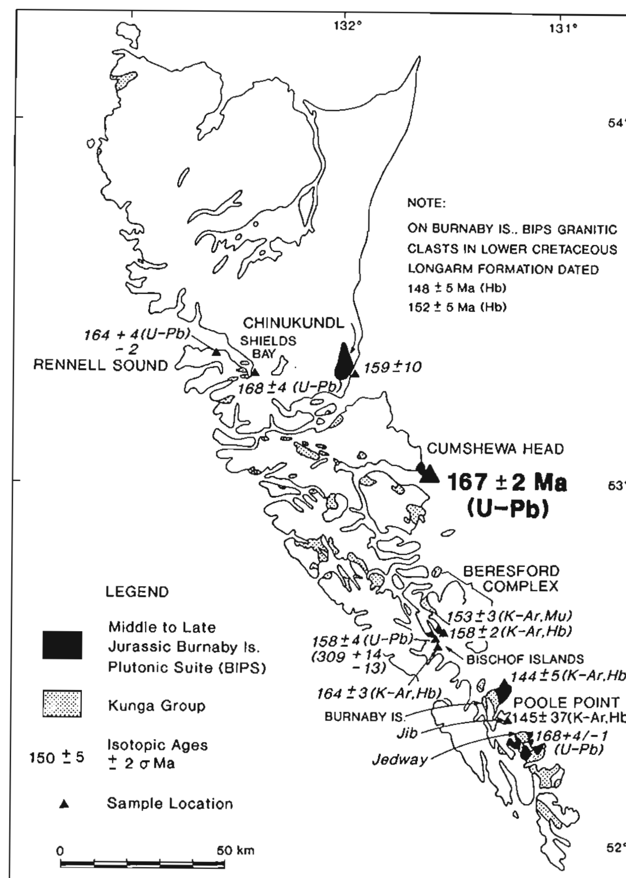

Figure 1. Distribution of Jurassic plutons of the Burnaby Island plutonic suite in Queen Charlotte Islands and their geochronology (modified from Anderson and Reichenbach, 1991).

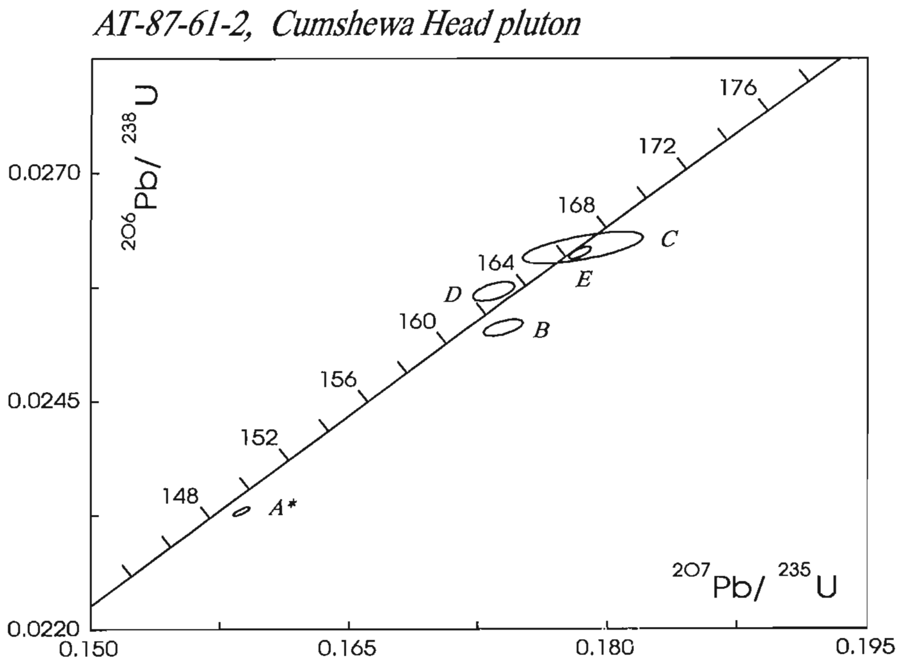

Figure 2.

U-Pb concordia plot for zircons from sample AT-87-61-2, the Cumshewa Head pluton. Error ellipses reflect the two sigma uncertainty.

Table 1. U-Pb analytical data for the Cumshewa Head pluton.

Fraction[a]	Wt.[b]	U	Pb[c]	$^{206}Pb^{d}$	Pb[e]	$^{208}Pb^{f}$	Radiogenic ratios (±1σ, %)[g]			Ages (Ma, ±2σ)[h]		
	mg	ppm	ppm	^{204}Pb	pg	%	$^{206}Pb/^{238}U$	$^{207}Pb/^{235}U$	$^{207}Pb/^{206}Pb$	$^{206}Pb/^{238}U$	$^{207}Pb/^{235}U$	$^{207}Pb/^{206}Pb$
AT-87-61-2, Cumshewa Head pluton [1]												
A*,105,el	0.165	191.46	4.478	1259	37	10.16	0.02328 ± 0.10	0.1586 ± 0.15	0.04942 ± 0.09	148.3 ± 0.3	149.5 ± 0.4	167.9 ± 4.3
B, 80,pr	0.056	168.3	4.283	774	19	10.21	0.02531 ± 0.18	0.1740 ± 0.33	0.04986 ± 0.26	161.1 ± 0.6	162.9 ± 1.0	188.6 ± 12.4
C, 90,el	0.018	99.05	2.643	200	15	11.41	0.02618 ± 0.33	0.1786 ± 0.98	0.04947 ± 0.80	166.6 ± 1.1	166.8 ± 3.0	169.9 ± 37.8
D, 90,el	0.070	168.6	4.406	1245	15	11.23	0.02570 ± 0.21	0.1734 ± 0.35	0.04894 ± 0.29	163.6 ± 0.7	162.4 ± 1.1	144.8 ± 13.6
E, 90,pr	0.099	112.9	2.998	1368	13	11.22	0.02611 ± 0.13	0.1784 ± 0.18	0.04955 ± 0.12	166.2 ± 0.4	166.7 ± 0.5	174.1 ± 5.7

[a]Zircon fractions are abraded except those marked with *; numbers refer to average size of zircons in microns, el = elongate, pr = prismatic (length to width ratio of about 2-3:1).
[b]Error on weight = ± 0.001 mg
[c]Radiogenic Pb
[d]Measured ratio corrected for spike and Pb fractionation of 0.09 ± 0.03%/AMU
[e]Total common Pb on analysis corrected for fractionation and spike
[f]Radiogenic Pb
[g]Corrected for blank Pb and U and common Pb (Stacey-Kramers model Pb composition equivalent to the $^{207}Pb/^{206}Pb$ age)
[h]Corrected for blank and common Pb

[1] Sample locality: Cumshewa Head pluton (BIPS), hornblende-biotite quartz monzonite; UTM zone 9, 5883730N - 323430E; 53°04'33" N - 131°38'09" W (NTS 103G/4); north of Cumshewa Head, 5.25 km south of Grey Point, 4.5 km north-northwest of Cumshewa Island, at sea level; collected by R.G. Anderson.

U-Pb ISOTOPIC DATA

Zircon was extracted from the crushed rock sample by conventional Wilfley table, heavy liquid, and Frantz magnetic separation techniques. The most euhedral, fracture-free, and inclusion-free zircon grains were chosen for analysis. All zircon fractions except A* were strongly air abraded following the method of Krogh (1982). U-Pb analytical methods used in this study are those outlined in Parrish et al. (1987). Techniques included mineral dissolution in microcapsules (Parrish, 1987), a mixed ^{205}Pb-^{233}U-^{235}U isotopic tracer (Parrish and Krogh, 1987), multicollector mass spectrometry (Roddick et al., 1987), and estimation of errors by numerical error propagation (Roddick, 1987). Analytical results are presented in Table 1, where errors on the ages are quoted at the two sigma level, and displayed in the concordia plot (Fig. 2).

Zircons analyzed from this sample include well-faceted, prismatic to more elongate crystals with elongate and round fluid inclusions and some opaque inclusions. The age of the rock is interpreted to be 167 ± 2 Ma based on the ages of the most concordant fractions C and E, which overlap each other and intersect concordia. Zircon fractions B and A* have slightly higher U concentrations and are interpreted to have

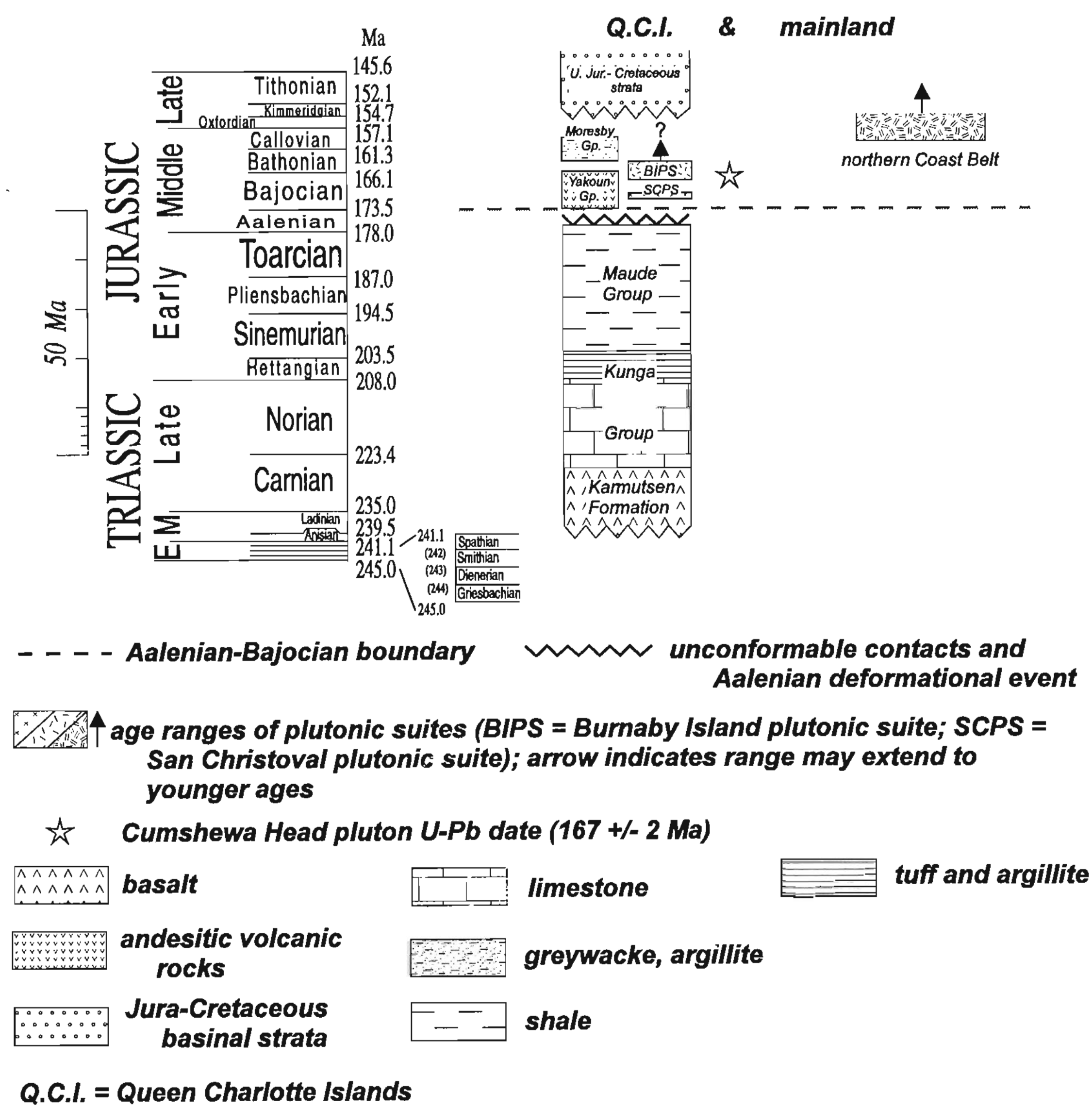

Figure 3. Time-stratigraphic summary based on the Harland et al. (1990) time scale emphasizes coeval linkage among some co-spatial plutonic-volcanic rock assemblages for Queen Charlotte Islands compared with that on the mainland between 52°-54° latitude (modified from Woodsworth, 1991; Lewis et al., 1991; Lyatsky and Haggart, 1992; and, van der Heyden, 1989).

undergone some Pb loss in accord with the highly altered and veined nature of this granitic rock. In particular, the analysis previously reported by Anderson and Reichenbach (1991) for fraction A*, was determined on unabraded zircons and consequently shows a stronger Pb loss effect.

The reverse discordance of fraction D is not well understood. However, the $^{207}Pb/^{206}Pb$ ratios on the mass spectrometer during the analysis of fraction D showed a sudden 3 per cent shift halfway through the run indicative of analytical problems. Fraction D could possibly be as young as ca. 164 Ma but the agreement and near concordance of fractions C and E suggest that the best interpretation for the age of the rock is 167 ± 2 Ma.

DISCUSSION

The new U-Pb date helps to confirm the correlation of the sampled unit based on its lithological characteristics and intrusive relations established in previous work. The age range of Burnaby Island plutonic suite defined by Anderson and Reichenbach (1991) is weighted strongly on porphyritic and fine grained intrusions within the older San Christoval plutonic suite. Of the Burnaby Island suite plutons on the east coasts of Queen Charlotte Islands, only the Poole Point pluton has been dated by U-Pb method (168 +4/-1 Ma; Anderson and Reichenbach, 1991). The new U-Pb date for the Cumshewa Head pluton is within the previously defined 168-≥158 Ma age range for Burnaby Island plutonic suite and is corroborated by a close similarity in plutonic style with other dated members of Burnaby Island plutonic suite. The age date also agrees with the pluton's poorly exposed intrusive relations with Bajocian Yakoun Group volcanic country rocks; according to the time scale of Harland et al. (1990), the Bajocian Yakoun Group is coeval with the 172-171 Ma San Christoval plutonic suite and older (168-164 Ma) members of Burnaby Island plutonic suite (Fig. 3). Finally, these youngest and easternmost of the Jurassic intrusions on Queen Charlotte Islands are significantly older than other plutonic belts to the east (Late Jurassic Banks Island belt (160-155 Ma), Early Cretaceous McCauley Island belt (131-123 Ma), and mid-Cretaceous Ecstall belt (110-94 Ma) on the mainland (van der Heyden, 1989). They are also older than Jurassic plutons along the western margin of the Coast Belt on the mainland to the southeast (164 Ma and younger; e.g., Monger and McNicoll, 1993; Friedman and Armstrong, in press).

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Geological Survey of Canada Project 840046

New K-Ar and U-Pb dates for the Cambria Icefield area, northwestern British Columbia

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Greig, C.J., McNicoll, V.J., Anderson, R.G., Daubeny, P.H., Harakal, J.E., and Runkle, D., 1995: New K-Ar and U-Pb dates for the Cambria Icefield area, northwestern British Columbia; in Current Research 1995-A; Geological Survey of Canada, p. 97-103.

Abstract: An Early Jurassic U-Pb zircon date (201.8 ± 0.5 Ma) for a sill of the Goldslide intrusions at the Red Mountain Au deposit helps to constrain the timing of mineralization and alteration, and places an older limit on regional deformation. An Early Jurassic date (194 ± 8 Ma, K-Ar, hornblende) for pyroclastic rocks of the Hazelton Group on strike with volcanic and volcaniclastic rocks that host, in part, the Red Mountain deposit supports the interpretation that the Goldslide intrusions and their country rocks are nearly contemporaneous. The Bulldog Creek pluton yields an Early to Middle Jurassic date (181 ± 8 Ma, K-Ar, hornblende), which corroborates its pre-kinematic structural setting. Eocene K-Ar biotite dates for three post-kinematic Tertiary plutons extend the known distribution of Tertiary plutons in the region.

Résumé : Une datation U-Pb sur zircon du Jurassique précoce (201,8 ± 0,5 Ma) pour un filon-couche des intrusions de Goldslide dans le gisement aurifère de Red Mountain aide à délimiter la chronologie de la minéralisation et de l'altération et à attribuer une limite plus ancienne à la déformation régionale. Une datation (194 ± 8 Ma, par K-Ar sur hornblende) du Jurassique précoce pour des roches pyroclastiques du Groupe de Hazelton qui sont parallèles à la direction des roches volcaniques et volcano-clastiques abritant, en partie, le gisement de Red Mountain, soutient l'interprétation selon laquelle les intrusions de Goldslide et leurs roches encaissantes sont presque contemporaines. Le pluton de Bulldog Creek donne un âge du Jurassique précoce à moyen (181 ± 8 Ma, par K-Ar sur hornblende), ce qui corrobore son cadre structural précinématique. Des datations K-Ar sur biotite de l'Éocène établies pour trois plutons postcinématiques du Tertiaire élargissent la distribution connue des plutons du Tertiaire dans la région.

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INTRODUCTION

Granitic and volcanic rocks in the Cambria Icefield area (NTS map areas 103P/13 and 14 and 104A/3 and 4; Fig. 1) are an important part of the Mesozoic and Tertiary geological development of the region. Their petrology and spatial and temporal distribution help constrain reconstruction of Mesozoic and Tertiary tectonics and refine exploration models for intrusion-related gold deposits such as at Red Mountain.

Volcanic stratigraphy and plutonic styles established in earlier field mapping (e.g., Greig et al., 1994a,b) and the existing regional geochronometric database (e.g., Smith, 1977; Alldrick et al., 1986, 1987; Brown, 1987; Anderson in Hunt and Roddick, 1991; Schroeter et al., 1992; Anderson et al., 1993) can be interpreted as numerous Mesozoic and Tertiary magmatic episodes including Late Triassic, Early Jurassic, Middle Jurassic, and Tertiary (Eocene). Compositional range, mineralogy, texture, structural fabric, spatial proximity with coeval volcanic rocks, and distinctive intraplutonic porphyritic intrusions provide criteria to establish the stratigraphic units and to group plutons. For example Late Triassic Stikine, Early Jurassic Texas Creek, Middle Jurassic Three Sisters and Eocene Hyder plutonic suites are recognized elsewhere in western Stikinia (e.g., Woodsworth et al., 1991).

Reconnaissance and detailed K-Ar and U-Pb dating are underway to complement the regional mapping of the Cambria Icefield area (e.g., Greig et al., 1994a) and biostratigraphic studies (e.g., Cordey et al., 1992). Reported here are some results of the U-Pb and K-Ar dating (Tables 1, 2; Fig. 1, 2). In all cases, the K-Ar dates provide minimum age estimates.

GEOLOGICAL SETTING AND PREVIOUS WORK

Early Jurassic plutonic and volcanic rocks

Goldslide intrusions

The Goldslide intrusions comprise stocks and related sills and dykes of seriate to porphyritic hornblende quartz monzonite, quartz monzodiorite, or monzodiorite which commonly contain distinctive books of biotite (≤ 0.5 cm), and locally, phenocrysts of quartz. The main intrusion on Red Mountain (Goldslide pluton, 160 Ma and 200 Ma, $^{40}Ar/^{39}Ar$ hornblende; Schroeter et al., 1992; T.G. Schroeter, written comm., 1994), and sills and dykes of similar texture and composition, are spatially and genetically related to the ore. A sample of one of the sills, a seriate biotite-hornblende quartz monzonite to monzodiorite about 50 m thick, was collected for U-Pb dating. The sill intrudes massive pyroxene-phyric mafic fragmental volcanic rocks and volcanic debris flows, and relatively well-layered, fine grained tuffaceous epiclastic rocks of probable Early Jurassic age. It lies above the ore zones, about 300 m west of the summit of Red Mountain (Fig. 1). The sill is cut by axinite and quartz-chlorite-epidote veins and contains pyrite as disseminations and fracture fillings.

Similar intrusive bodies to those on Red Mountain occur 6-7 km to the south and southwest, north of Homestake Ridge, northwest of North Flat Glacier, and west of Willoughby Glacier. All are typically dyked, veined, altered, and mineralized, and where in contact with uppermost Triassic and lowermost Jurassic country rocks, they resemble peperites (i.e., intrusions emplaced into unlithified or poorly lithified hosts). Peperite features present in the sampled sill and common to other intrusions at Red Mountain include: 1) highly disrupted wall rocks, with common plastically-deformed fragments of country rock contained within a tuffaceous matrix; 2) breccia-dykes within the country rocks and locally within the intrusions and which contain fragments of both; and 3) highly lobate contacts with distinct chill margins, local detached apophyses, and wide alteration haloes.

The Goldslide intrusions locally contain a regional cleavage prominent in their host rocks. Therefore the age of the intrusions places an older limit on the timing of regional deformation affecting rocks within the Bitter Creek antiform, an important regional structure. Where intensely developed within the intrusions, the cleavage is defined by dynamically recrystallized and aligned, very fine grained white mica and chlorite that are constituents of a widespread alteration assemblage which replaced feldspars and mafic minerals. In contrast to this 'ductile' strain evident in the secondary phyllosilicates, primary quartz, as well as secondary pyrite, were deformed brittley, and occur as porphyroclasts surrounded by phyllosilicates and common carbonate which formed tails and fracture fillings. Thus, the alteration event is interpreted to be pre-kinematic.

Bulldog Creek pluton

The Bulldog Creek pluton underlies much of the southwestern part of the map area (Fig. 1). It intrudes clastic and fragmental rocks of probable Late Triassic to earliest Jurassic age, and is itself intruded by the Kshwan Glacier and Sutton River plutons (see below) and by a variety of dykes. The Bulldog Creek pluton is distinguished from the Tertiary plutons by its darker grey weathering colour and common green cast. It consists predominantly of medium grained, equigranular to locally seriate, unfoliated to weakly foliated, biotite-hornblende quartz monzodiorite, quartz monzonite, monzonite, and monzodiorite. Near its northeast margin, a faint gneissosity comprises zones of weakly aligned, centimetre-scale lenses of medium grained biotite-hornblende monzodiorite contained in and grading into a finer grained groundmass of similar composition. The Bulldog Creek pluton is commonly epidotized and chloritized, and crosscut by extensive anastomosing epidote-chlorite-quartz veinlets. Locally, a garnet-epidote-pyrite-chalcopyrite(?) endoskarn(?) assemblage is developed across widths of up to 0.5 m. Discontinuous (5 or more metres along strike), roughly planar centimetre- to decimetre-wide chlorite-matrix breccia zones are common throughout the pluton and locally associated with skarn zones. Intensity of alteration and contact relations are consistent with a post-Triassic, pre-middle Cretaceous age for the Bulldog Creek pluton (Greig et al., 1994b). Formerly, plutonic rocks in this area were not subdivided and were assigned a Tertiary and older(?) age (Carter, 1981; Grove, 1986).

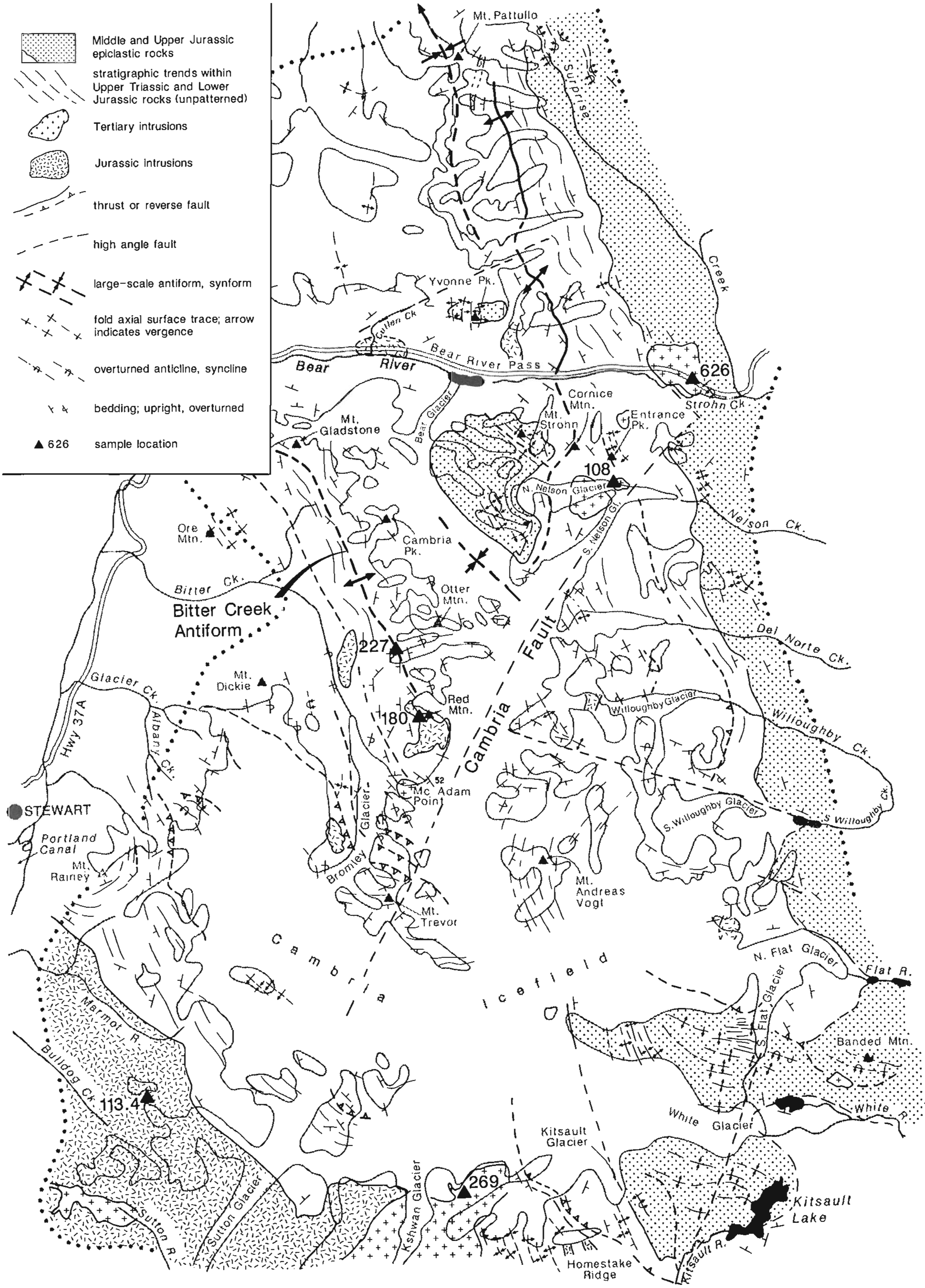

Figure 1. Distribution of igneous units in the Cambria Icefield area and location of U-Pb and K-Ar samples.

Hazelton Group coarse lapilli tuff

Steeply dipping, massive, dark grey-green, hornblende-feldspar-phyric basaltic andesite lapilli tuff or tuff-breccia of the Hazelton Group occurs within the core of a complexly faulted, north-northwest-trending regional antiform, the Bitter Creek antiform. They occur between an east-facing volcanic tuffaceous epiclastic sequence to the east, and a folded but generally west-facing predominantly epiclastic package immediately to the west. The tuffs are 2 km to the northwest of, and on strike with, host rocks for the Red Mountain deposit.

Tertiary(?) plutonic rocks

Tertiary plutons described in this paper belong to a suite of seven plutons in the Cambria Icefield area which were assigned Tertiary ages by Greig et al. (1994a,b), on the basis of intrusive relations, structural style, relative lack of alteration and dyking, and similarity to Middle Eocene intrusions of the Hyder plutonic suite (Grove, 1986; Woodsworth et al., 1991). Grove (1986) described the Strohn Creek and McAdam Point plutons, and mapped the Entrance Peak and Banded Mountain plutons, which were not examined in this study. A further description of the Strohn Creek pluton and its setting is given by Evenchick et al. (1992).

The Tertiary plutons vary in composition from quartz monzonite and monzogranite to granodiorite or quartz diorite, commonly contain biotite and scattered pink potassium feldspar megacrysts, and are typically medium grained and unfoliated. The north-northeast alignment of five of the plutons immediately northwest of the Cambria fault suggests structural control on their emplacement. Three of the plutons, the Sutton River, Kshwan Glacier and Nelson glaciers plutons, were not previously documented. Previously dated as Eocene in age are the McAdam Point (45 ± 4 Ma, $^{40}Ar/^{39}Ar$, biotite (Schroeter et al., 1992; T.G. Schroeter, written comm., 1994)) and Bitter Creek (48.4 ± 3.4 Ma, K-Ar, biotite (Alldrick, 1993)) plutons. Dating of the Sutton River pluton (U-Pb) is underway.

Two prominent dyke swarms are related to the Tertiary plutons: 1) the north-northwesterly striking Portland Canal swarm, which extends across Bitter Creek, underlies Mt. Dickie and part of Ore Mountain, and may emanate from the Bitter Creek pluton; and 2) the northwesterly striking Nelson glaciers swarm, which emanates from the Nelson glaciers pluton. The dyke swarms are characterized by composition as well as orientation. The Nelson swarm consists of homogeneous (biotite-, quartz-) hornblende-feldspar porphyritic dykes of probable intermediate composition, whereas those of the Portland Canal swarm range from lamprophyre and basalt to dacite, rhyolite and granite. Dykes of the Nelson swarm are almost invariably subvertical, compared with Portland Canal dykes, which generally dip west. Intrusion of the Portland Canal dykes was mainly controlled by orientations of the bedding and cleavage in country rocks (see Grove, 1986).

Nelson glaciers pluton

The Nelson glaciers pluton may comprise two intrusions, a larger one to the southwest underlying and exposed on both sides of the north Nelson glacier, and a northwestern body north of the junction between the north and south Nelson glaciers (Fig. 1). Both have irregular west-northwest- to northwest-trending contacts with subparallel northwest-striking dykes. The southwestern body consists of massive, equigranular biotite granodiorite containing approximately 1% dark grey, very fine grained inclusions, and cut by common centimetre-scale aplite and pegmatite dykes. The northeastern body, which was sampled for K-Ar dating, consists of massive, potassium feldspar megacrystic (hornblende?) biotite monzogranite with common crosscutting decimetre-scale aplite dykes. Dykes of the Nelson glaciers swarm are clearly post-kinematic with respect to folding of rocks as young as Oxfordian-Kimmeridgian.

Sutton River pluton

The Sutton River pluton occurs in the southwest corner of the map area and is a massive, fine- to medium-grained, seriate biotite granite or monzogranite (Fig. 1). It has well-developed stockwork and agmatite zones along its contacts with the Bulldog Creek pluton and Triassic and/or Jurassic clastic rocks. Fine grained biotite is locally intergrown with muscovite. The Sutton River pluton may be a marginal phase or a satellitic body of a much larger monzogranite pluton that underlies the region between the Georgie River and Anyox pendants to the south.

Kshwan Glacier pluton

The Kshwan Glacier pluton is a massive, medium-grained, hornblende-biotite granodiorite. It intrudes the Bulldog Creek pluton on its west and south margins, and Triassic and/or Lower Jurassic clastic and volcaniclastic rocks to the north and east (Fig. 1). Silty mudstone adjacent to its eastern contact contains abundant cordierite and very fine-grained brown biotite. Limy pods within adjacent muddy debris flows locally contain centimetre- and decimetre-scale skarn assemblages that include brown garnet, epidote, calcite, chlorite, amphibole and quartz(?). The Kshwan Glacier pluton is intruded by aplite, (biotite, quartz) hornblende needle feldspar porphyry, and basaltic-andesite dykes.

U-Pb ISOTOPIC DATING

Analytical techniques

Zircon was extracted from the crushed Goldslide sill sample (EPC-93-180) by conventional Wilfley table, heavy liquid, and Frantz magnetic separation techniques. The most euhedral, fracture-free, and inclusion-free zircon grains were chosen for analysis. Analytical methods used in this study are those outlined in Parrish et al. (1987). Techniques included strong air abrasion (Krogh, 1982) for all of the zircon fractions analyzed, mineral dissolution in microcapsules (Parrish, 1987), a mixed ^{205}Pb-^{233}U-^{235}U

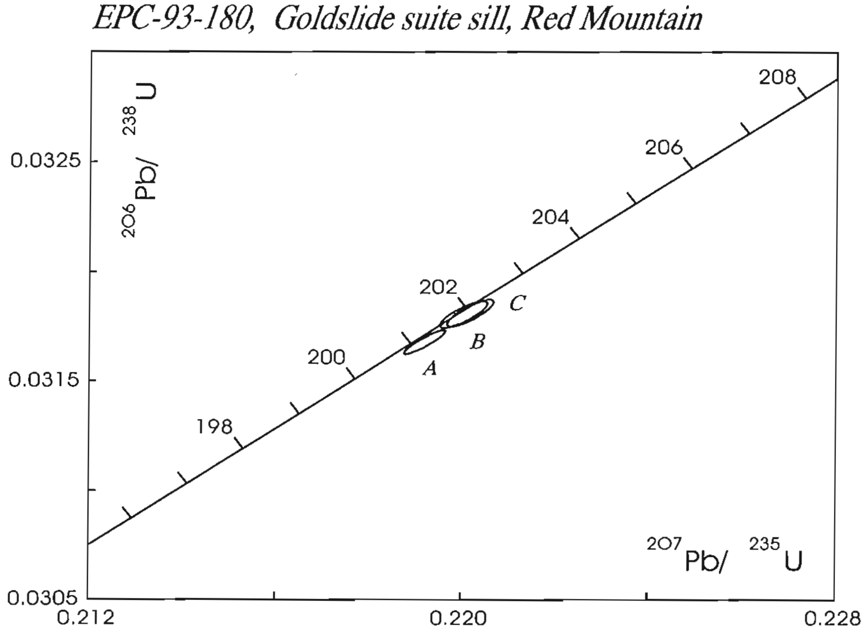

Figure 2.

U-Pb concordia plot for zircon from sample EPC-93-180, a sill of the Goldslide intrusions on Red Mountain.

Table 1. U-Pb analytical data for zircon from a sill of the Goldslide intrusions.

							Radiogenic ratios (±1σ, %)[g]			Ages (Ma, ±2σ)[h]		
Fraction[a]	Wt.[b] mg	U ppm	Pb[c] ppm	^{206}Pb[d] / ^{204}Pb	Pb[e] pg	^{208}Pb[f] %	$^{206}Pb/^{238}U$	$^{207}Pb/^{235}U$	$^{207}Pb/^{206}Pb$	$^{206}Pb/^{238}U$	$^{207}Pb/^{235}U$	$^{207}Pb/^{206}Pb$
<u>EPC-93-180, Goldslide suite biotite-hornblende quartz monzonite to monzodiorite sill</u> [1]												
A, 65,pr	0.048	876.0	26.94	9841	8	6.996	0.03168 ± 0.09	0.2192 ± 0.10	0.05019 ± 0.04	201.0 ± 0.3	201.3 ± 0.4	203.7 ± 1.9
B, 70,pr	0.045	480.4	14.72	6440	7	6.297	0.03180 ± 0.10	0.2201 ± 0.11	0.05019 ± 0.06	201.8 ± 0.4	202.0 ± 0.4	203.9 ± 2.6
C, 60,pr	0.029	897.8	27.56	5422	10	6.465	0.03180 ± 0.10	0.2199 ± 0.11	0.05016 ± 0.06	201.8 ± 0.4	201.8 ± 0.4	202.5 ± 2.6

[a]Numbers refer to average size of zircons in microns, pr = prismatic (crystal length to width ratio of about 2-3:1)
[b]Error on weight = ± 0.001 mg
[c]Radiogenic Pb
[d]Measured ratio corrected for spike and Pb fractionation of 0.09 ± 0.03%/AMU
[e]Total common Pb on analysis corrected for fractionation and spike
[f]Radiogenic Pb
[g]Corrected for blank Pb and U and common Pb (Stacey-Kramers model Pb composition equivalent to the $^{207}Pb/^{206}Pb$ age)
[h]Corrected for blank and common Pb

[1] Sample locality: collected from a Goldslide suite sill on Red Mountain, approximately 19 km ENE of the town of Stewart, B.C. (NTS 103P/13); UTM zone 9, 456725 E - 6202250 N; 55°57'56" N - 129°41'36" W.

Table 2. K-Ar data for plutonic and volcanic rocks in the Cambria Icefield area.

Sample No.	Mineral	K (wt.%)	Rad. ^{40}Ar cc/gm 10^{-6}	Rad. $^{40}Ar/^{40}Ar$	Age ± 2σ (Ma)	UTM (zone 09) Easting	UTM (zone 09) Northing	NTS
EPC-93-626 Strohn Creek pluton	biotite	5.73	11.720	87.5	51.9 ± 2.6	468700	6217000	104A
EPC-93-108 Nelson glaciers pluton	biotite	7.30	14.542	84.5	50.5 ± 2.4	465300	6212620	104A
EPC-93-227 Hazelton Group tuff	hornblende	0.880	7.017	88.0	194 ± 8	454925	6205455	103P
EPC-93-113.4 Bulldog Creek pluton	hornblende	0.595	4.411	86.4	181 ± 8	442900	6186510	103P
EPC-93-269 Kshwan Glacier pluton	biotite	6.10	11.607	89.7	48.3 ± 2.6	456780	6181020	103P

isotopic tracer (Parrish and Krogh, 1987), multicollector mass spectrometry (Roddick et al., 1987) and estimation of errors by numerical error propagation (Roddick, 1987). Analytical results are presented in Table 1, where errors on the ages are quoted at the 2 sigma level, and displayed in the concordia plot (Fig. 2).

U-Pb results and interpretation

Three fractions of clear, colourless, well faceted, prismatic zircon with minor, mostly fluidal, inclusions and rare fractures were analyzed from the Goldslide suite sill on Red Mountain. The interpreted age is 201.8 ± 0.5 Ma, based on the ages of fractions B and C, which overlap concordia and each other (Fig. 2). Fraction A is interpreted to have undergone a minor amount of Pb loss, consistent with a fairly high U content. The Early Jurassic age of the sill is in agreement with the older of the two $^{40}Ar/^{39}Ar$ dates for the Goldslide pluton (200 Ma, Schroeter et al., 1992; T.G. Schroeter, written comm., 1994), and probably dates the main mineralizing and alteration events at Red Mountain. Such an age is also consistent with the contact relationships and with regional evidence for an Early Jurassic mineralizing event (Greig et al., 1994b). In addition, it places an older limit on development of the regional cleavage.

K-Ar ISOTOPIC DATING

Analytical techniques

Minerals were separated and K and Ar were analyzed at the Geochronology Laboratory of the Department of Geological Sciences, University of British Columbia, using techniques described by Greig et al. (1992).

K-Ar results and interpretation

Massive, equigranular, biotite hornblende quartz monzodiorite from the Bulldog Creek pluton yielded a 181 ± 8 Ma hornblende age (Early to Middle Jurassic) (Table 2). This date confirms the pluton's suspected Jurassic age and represents the youngest age for a pluton that is considerably altered. However, the hornblende in the mineral separate was partially chloritized and some degree of resetting may have occurred. U-Pb zircon dating of a nearby sample of the Bulldog Creek pluton is underway.

The 194 ± 8 Ma age for Hazelton Group tuff-breccia northwest of Red Mountain (Table 2) provides a minimum age for volcanic rocks within the Bitter Creek antiform. The correlation of the tuffs with the host rocks for the Red Mountain deposit, and overlap of the date, within error, with the age of the Goldslide intrusions supports the interpretation that the Goldslide intrusions, their country rocks, and mineralization are nearly contemporaneous. The possibility of some degree of resetting must be considered because the sample from which the hornblende was separated was considerably altered (although the hornblende itself appears unaltered) and because of the unit's complex structural setting.

Middle Eocene K-Ar dates for biotite from the Strohn Creek (51.9 ± 2.6 Ma), Nelson glaciers (50.5 ± 2.4 Ma) and Kshwan Glacier (48.3 ± 2.6 Ma) plutons provide minimum ages for pluton emplacement (Table 2). They are essentially coeval with the McAdam Point and Bitter Creek plutons, and with the Alice Arm and Hyder intrusions to the southeast and northwest of the Cambria Icefield area, along the eastern margin of the Coast Belt (Smith, 1977; Carter, 1981; Woodsworth et al., 1991). They are part of an extensive

Eocene magmatic arc, which is locally genetically-related to significant porphyry molybdenum resources, particulary to the southeast (Carter, 1981).

CONCLUDING REMARKS

The dates reported herein confirm that in this part of Stikinia, as elsewhere, Early Jurassic magmatism and metallogenesis are important. The dates also help to bracket the timing of regional deformation in the Cambria Icefield area to be between Late Jurassic and Eocene age.

ACKNOWLEDGMENTS

The fieldwork and laboratory analyses were supported by the Industrial Partners Project involving the Geological Survey of Canada and LAC Minerals Ltd. The co-operation during this project of many LAC Minerals Ltd. geologists, in particular Kate Bull, Dave Cawood, Garfield MacVeigh, Marc Prefontaine, Dave Rhys, Hans Smit, and Gernot Wober is gratefully acknowledged. The staff of the GSC Geochronology Laboratory are thanked for their assistance in generating the U-Pb data. Thanks go to Bev Vanlier for proofing the manuscript and to Bert Struik for a review.

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Revised stratigraphy for the Hoodoo Mountain volcanic centre, northwestern British Columbia

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Edwards, B.R., Edwards, G., and Russell, J.K., 1995: Revised stratigraphy for the Hoodoo Mountain volcanic centre, northwestern British Columbia; in *Current Research 1995-A; Geological Survey of Canada, p. 105-115.*

Abstract: Detailed mapping during 1994 at Hoodoo Mountain volcanic centre indicates: 1) some hyaloclastite deposits on Hoodoo Mountain are genetically related to subglacial lava flows, 2) pillow lavas and pillow lava breccias occur on Little Bear mountain, and 3) tillites and glaciolacustrine sediments are an important part of the volcanic stratigraphy. The revised stratigraphy for the Hoodoo Mountain volcanic centre suggests: 1) at least five periods of subaerial and subglacial eruption at Hoodoo Mountain volcano, 2) the Little Bear mountain volcano is older than the youngest subglacial lava flows of the Hoodoo Mountain volcano, and 3) grey phonolite dykes associated with subglacial eruptions from Hoodoo Mountain volcano postdate the two oldest lava series at that volcano.

Résumé : Une cartographie détaillée effectuée en 1994 au centre volcanique du mont Hoodoo laisse voir que : 1) certains dépôts de hyaloclastites sur le mont Hoodoo sont génétiquement liés à des coulées de lave sous-glaciaires, 2) des laves en coussins et des brèches de lave en coussins sont présentes sur le mont Little Bear, et 3) les tillites et les sédiments glaciolacustres constituent une partie importante de la stratigraphie volcanique. La stratigraphie révisée du centre volcanique du mont Hoodoo laisse supposer que : 1) le volcan du mont Hoodoo a connu au moins cinq périodes d'éruption subaérienne et sous-glaciaire, 2) le volcan du mont Little Bear est plus ancien que les plus récentes coulées de lave sous-glaciaires du volcan du mont Hoodoo, et 3) les dykes de phonolite grise associés aux éruptions sous-glaciaires du volcan du mont Hoodoo sont postérieurs aux deux plus anciennes séries de laves de ce volcan.

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INTRODUCTION

The Hoodoo Mountain volcanic centre comprises two volcanoes immediately north of the Iskut River, in northwestern British Columbia (Fig. 1). Hoodoo Mountain volcano (UTM centre 360000E/6295000N) consists of lava flows, dykes, pyroclastic ejecta, and debris flows (Edwards and Russell, 1994a) which erupted both subaerially and subglacially. A small basaltic volcano immediately north of Hoodoo Mountain (UTM centre 359500E/6297500N), informally referred to as Little Bear mountain volcano, is interpreted as a tuya or subglacial volcano (Edwards and Russell, 1994a). The geography and general geology of Hoodoo Mountain volcanic centre have been described briefly by Kerr (1948) and Souther (1990) and in more detail by Edwards and Russell (1994a,b).

This report summarizes results from geological mapping completed between July and August 1994, during which several new units were recognized: peralkaline subglacial lava flows and hyaloclastite, basaltic pillow lavas and pillow lava breccias, tillites, glaciolacustrine sediments, debris flows, and intermediate to felsic dykes which intrude the basement rocks. The three most important of these deposits are described below sequentially: 1) peralkaline subglacial lava flows, 2) basaltic pillow lavas and pillow lava breccias, and 3) tillites and glaciolacustrine sediments. Stratigraphic observations on these three deposits lead to a revision of the stratigraphy of the Hoodoo Mountain volcanic centre presented by Edwards and Russell (1994b).

STRATIGRAPHY OF THE HOODOO MOUNTAIN VOLCANIC CENTRE

Previous work

Kerr (1948), B.C. Hydro (1985), and Edwards and Russell (1994a) described stratigraphic relationships from the Hoodoo Mountain volcanic centre in varying degrees of detail. Kerr (1948) recognized at least six stratigraphic intervals at Hoodoo Mountain. The oldest lava (unit 1), which is distinctly columnar jointed, green-black and fine grained, is only exposed in a small area on the west flank of the volcano and presumably rests on top of older Permian and Triassic basement rocks. Grey and brown glacial "drift" (units 2 and 4) and, on the west side in Hoodoo Valley, deltaic, glacial lake sediments is widespread. The oldest abundant lava flows (unit 3), which were described as pahoehoe, aphanitic to porphyritic, ice-dammed lavas, were later glaciated. Younger pahoehoe lava flows (unit 5) were described from the south and east sides and contain interbedded ash layers. Aa lava flows (unit 6) were recognized as forming most of the top of

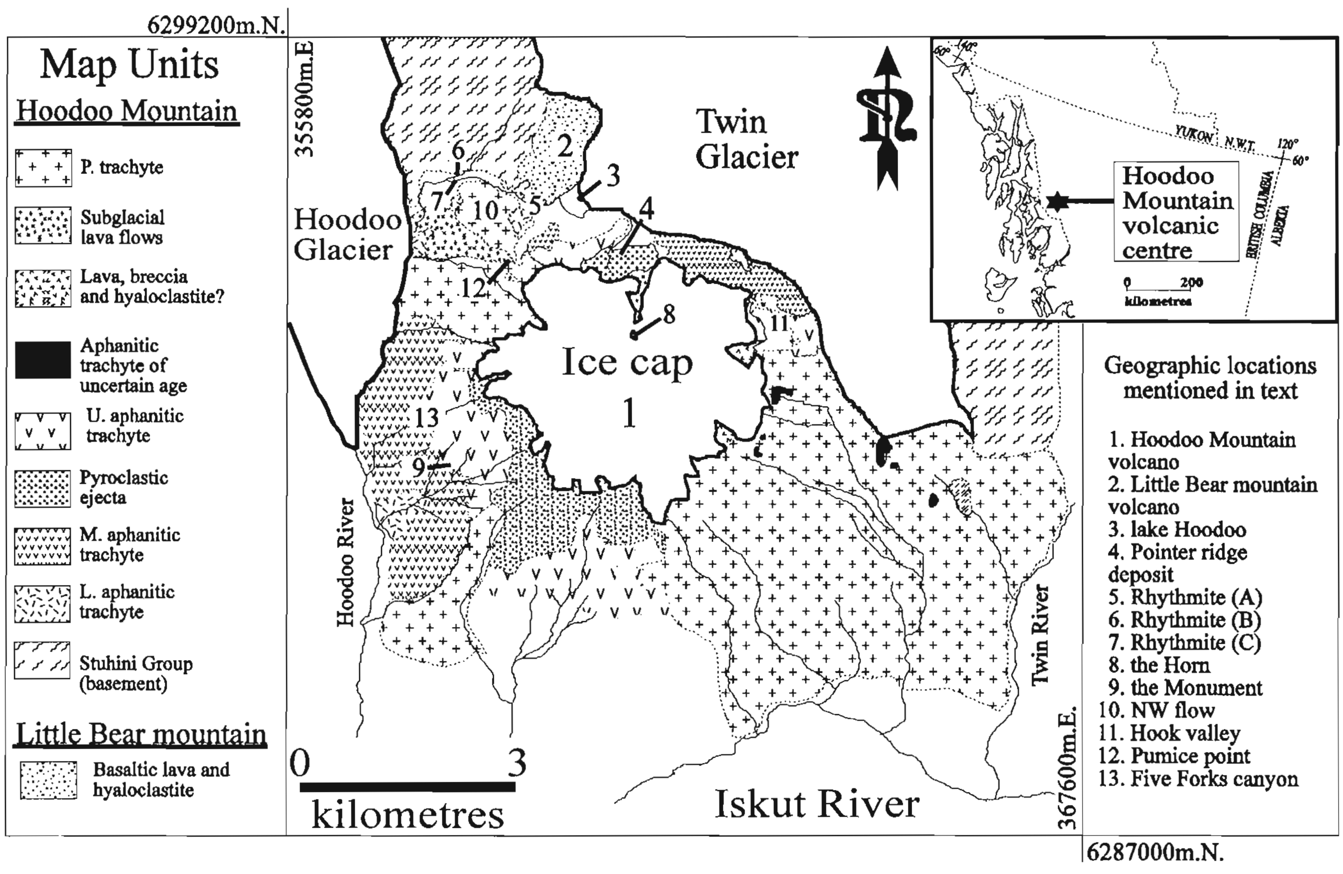


Figure 1. Revised geologic map of Hoodoo Mountain volcanic complex. Inset map of the Canadian Cordillera shows the location of the complex on a regional scale.

the mountain. B.C. Hydro (1985) followed Kerr's stratigraphy and descriptions, adding that by tree ring counts on live trees growing on top of one of the most recent lava flows a minimum age for the flow is 180 years. They also recognized a 10 m thick ash layer on the north side of the mountain.

Few revisions were made to Kerr's overall stratigraphic relationships by Edwards and Russell (1994a). However, Kerr apparently did not work at all on the north side of Hoodoo, where glaciation exposed more of the volcanic stratigraphy of Hoodoo Mountain as well as Little Bear mountain volcano. Significant revisions to the stratigraphy derived from the formerly unexplored the north and west sides include the discovery of an interval between Kerr's (1948) units 1 and 5 of extensive explosive and/or subglacial/subaqueous volcanic activity, including deposits of welded pyroclastic material and hyaloclastite.

PERALKALINE, SUBGLACIAL LAVA FLOWS, AND HYALOCLASTITE

The close spatial association between "grey aphanite" and pumiceous deposits was noted by Edwards and Russell (1994a); however, the genetic relationship between them was unclear. Electron microprobe analyses of glass bombs within the grey aphanite showed them to be peralkaline (mole % Al_2O_3 mole % $Na_2O + K_2O$) and phonolitic in composition (Edwards and Russell, 1994b). Close inspection of four areas in 1994 (informally referred to as Hook valley, Horn ridge, Pumice point, and Five Forks canyon; see Fig. 1) suggest that these deposits are cogenetic and represent subglacial lava flows.

"The Horn", a prominent nunatak on the northern edge of the Hoodoo ice cap (Fig. 1, location 8), is one of many examples of grey phonolite which is overlain by yellow, pumiceous hyaloclastite (Fig. 3) and underlies a ridge which extends northward above the surrounding ice. The grey phonolite is aphanitic, highly vesicular, and has radial columnar joints (Fig. 2). Some of the hyaloclastite on this ridge also contains locally abundant clasts of aphanitic trachyte.

The most complete exposures of the grey phonolite-hyaloclastite association are along a ridge on the south side of a distinctive, recent aa lava flow informally referred to as NW flow (Fig. 1, location 10) in Edwards and Russell (1994a). The ridge is composed of: (1) yellow, lithified hyaloclastite consisting of heterogeneously distributed, highly vesiculated and elongate lapilli and bombs in a yellow, palagonitized matrix (Fig. 4, 5); (2) green, nonlithified hyaloclastite comprising mainly lapilli and ash with local bombs of vesicular grey phonolite (Fig. 6, 7); and (3) grey phonolite. On the south side of the ridge, the subglacial lava flows cascaded down on top of domes of columnar jointed aphanitic trachyte and were subsequently covered by porphyritic trachyte lava flows from vents higher up the volcano (Fig. 6). At one location (Fig. 6), three coarsening upwards sequences in the green hyaloclastite culminate in deposits of large, glass rimmed, grey phonolite bombs. From Pumice point down to Hoodoo Glacier, the grey phonolite is underlain by green hyaloclastite.

Above 1300 m elevation the grey phonolite occurs as dykes which intrude older aphanitic trachyte flows and breccias. Several large, spine-shaped hoodoos, including one referred to by Kerr (1948) as "the Monument", are made of grey phonolite. This spine has horizontal columnar jointing and is associated with a west-trending dyke (Fig. 8, cf. Fig. 1, location 9). Five hundred metres north of Hoodoo canyon, in upper Five Forks canyon (Fig. 1, location 13), another grey phonolite dyke has radial columnar joints. Radial jointing, underlying and overlying hyaloclastite and aphanitic texture all support the hypothesis that the grey phonolite erupted subglacially.

BASALTIC PILLOW LAVAS AND HYALOCLASTITE

Four units were recognized by Edwards and Russell (1994a) as underlying Little Bear mountain: basaltic breccia, basalt dykes, hyaloclastite, and crystal lithic tuff. Detailed mapping in 1994 resulted in a revision of units at this volcano. Massive basalt and pillow lava were found near the base of the northern summit on the east, south and west sides (Fig. 9). Individual pillows range in size from 45 by 48 cm to 1 by 3 m (Fig. 9, 10). The pillows locally have radial jointing and white xenoliths oriented parallel to the pillow exterior surfaces. Concentric, vesicle-rich bands similar to those reported by Jones (1970) in subglacial pillow lavas from Iceland also occur at Little Bear mountain.

Orange-brown basaltic breccia comprising cobble- and gravel-sized basalt clasts in a matrix of sand-sized, vitric, lithic and crystal grains is the volumetrically dominant unit at Little Bear mountain volcano (Fig. 11). Matrix- and clast-supported varieties exist and locally have abundant conjugate joint sets which have been filled with orange sand-sized matrix. On the northern half of the volcano, basalt clasts which are generally round and contain heavily oxidized rims and vesicular interiors are interpreted as pillow fragments (Fig. 11). The breccia that contains pillow fragments is equivalent to para-pillow lava of Jones (1970). In several areas the basaltic breccia lava filled channels in underlying volcanic sandstone (Fig. 11). Along the south fork of upper Hoodoo River and extending down in elevation to the confluence with the north fork, reworked basaltic breccia unconformably overlies basement rocks (Fig. 12). Just above the confluence, the basaltic breccia is laminated and interbedded with reworked hyaloclastite.

The breccia is considered to be coeval with the basaltic lavas at Little Bear mountain and is interpreted as having formed by the disruption of pillows during subaqueous/subglacial emplacement of the basaltic lavas. In a topographically lower basin to the southwest of Little Bear mountain, the breccia was deposited by debris flows shed off the edifice. Basaltic breccia forms the middle part of the Standard Depositional Unit of Bergh and Sigvaldason (1991) for subglacial/subaqueous lava flows.

Grey hyaloclastite and dark brown volcanic mudstone are subordinate to basaltic breccia at Little Bear mountain. Both are most abundant on the southern flank of the volcano.

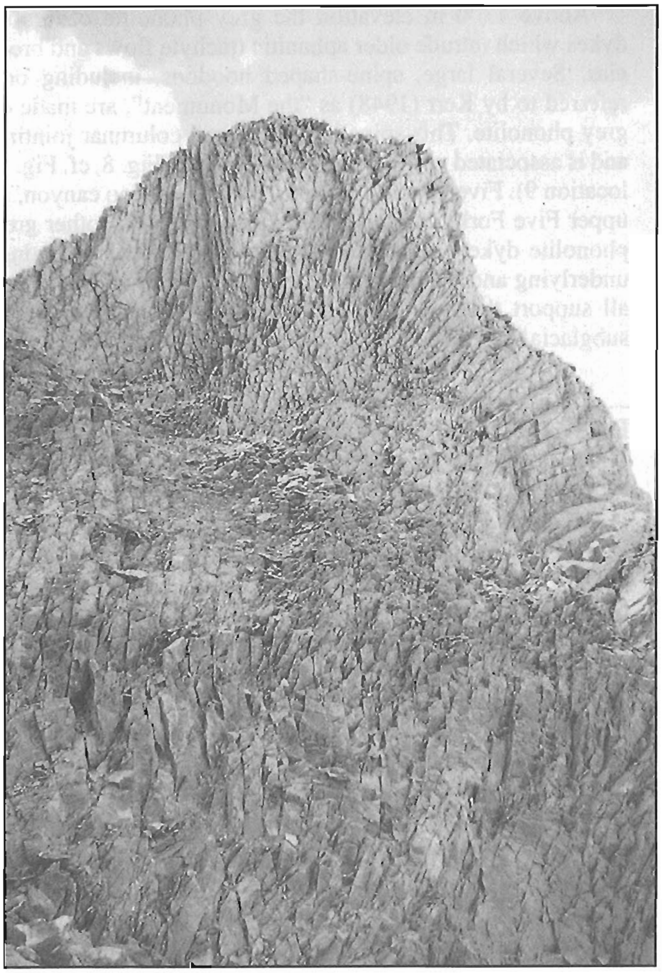


Figure 2. *Radial columnar jointing in grey, aphanitic, highly vesicular phonolite of "the Horn". The exposure is approximately 15 m high.*

Figure 4. *Glassy, highly vesicular bomb in hyaloclastite. The hyaloclastite comprises lapilli and elongate bombs (greater than 30 cm) of highly vesicular, phonolite magma in a palagonitized matrix. Ice axe is 3 cm wide.*

Figure 3.

Subglacial lava flow (A) that forms a sharp ridge on the north side of Hoodoo Mountain volcano and is surrounded by hyaloclastite (B). The highest exposure of this flow is a nunatak ("the Horn") which protrudes through the Hoodoo Mountain ice cap. Person is 2 m tall and is standing on the hyaloclastite.

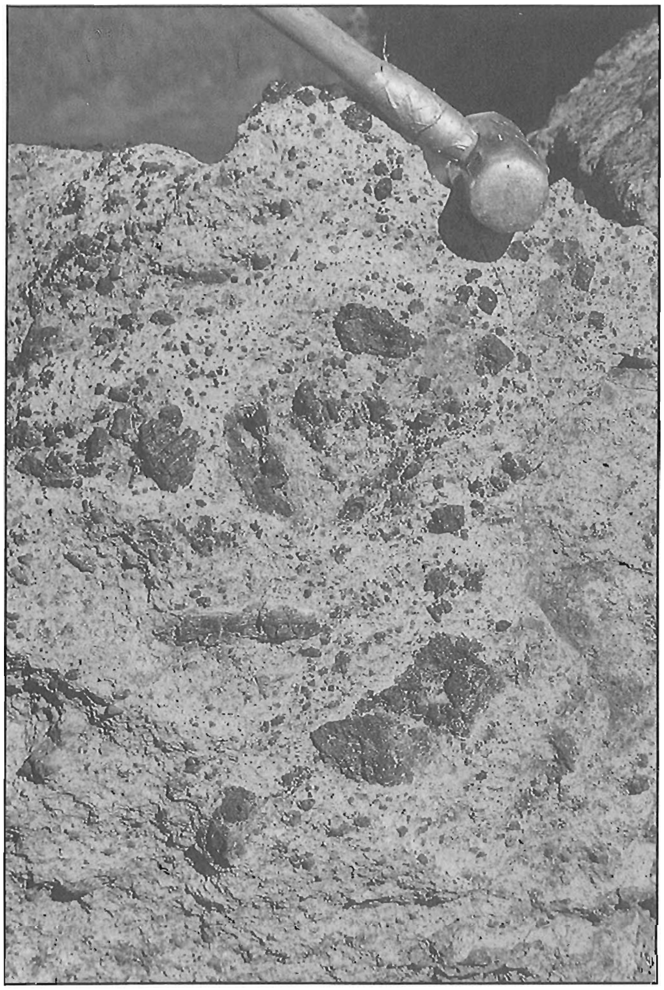


Figure 5. Higher than normal abundance of glass lapilli in hyaloclastite. Sledge hammer is 38 cm long.

Hyaloclastite locally contains thin (<2 m) lenses of basaltic breccia (Fig. 12) as well as isolated cobbles of basalt and rounded metavolcanic rocks. The mudstone locally contains up to 10 per cent angular, black vitric fragments. The hyaloclastite is interpreted as having formed during subglacial/subaqueous eruption of basaltic lava by spallation of glass from the rinds of pillow lavas.

TILLITES AND GLACIOLACUSTRINE RHYTHMITES

Both lithified and nonlithified glacial sediments occur within the Hoodoo Mountain volcanic centre stratigraphy. Polymictic diamictites interpreted as tillites and glaciolacustrine rhythmites are exposed in several locations along streams to the north and northwest of Hoodoo Mountain. Both are described in detail below.

Tillites

The most widespread sedimentary deposits are polymictic diamictites that are interpreted as tillites. They are interbedded with units from both Little Bear and Hoodoo Mountain volcanoes. Along the eastern edge of Hoodoo Glacier and adjacent west slope of Hoodoo Mountain three different diamictites both over- and underlie subglacial lava flows from Hoodoo Mountain. The lowest unit is polymictic, poorly lithified, approximately 5 m thick and is directly on top of a pre-Hoodoo(?), jointed, rhyolitic to dacitic volcanic unit. The lowest unit is overlain by a brown, well-lithified diamictite with rounded clasts of both metamorphic basement rocks (Fig. 13) and fresh, vesicular, plagioclase-phyric basalt (Fig. 14). The basalt clasts probably are derived from Little Bear mountain volcano, as no other basaltic volcanoes occur immediately north of this area. The middle diamictite is about

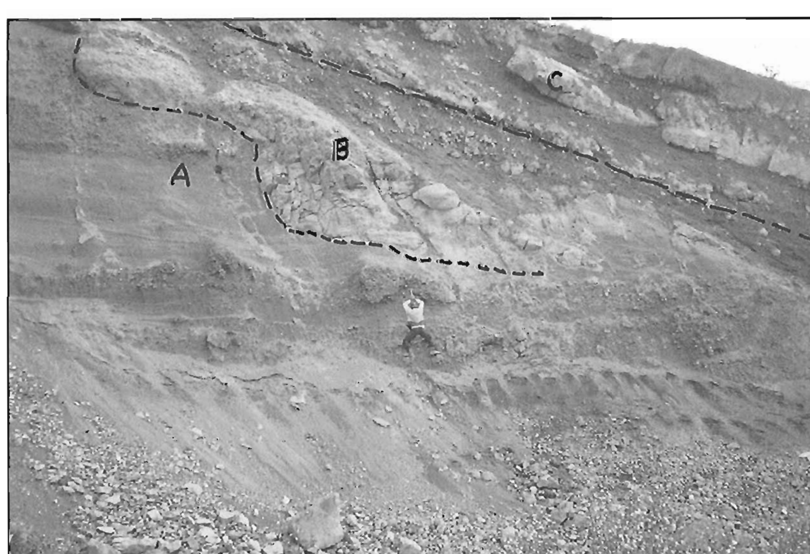


Figure 6.

Unlithified hyaloclastite deposit (A) underlying a subglacial lava flow on the west side of Hoodoo Mountain. The deposit varies from 5 to 10 m in thickness and is crudely bedded. A small block of lithified hyaloclastite (B) and younger porphyritic trachyte lava flows (C) overlie the deposit. Person is 2 m tall.

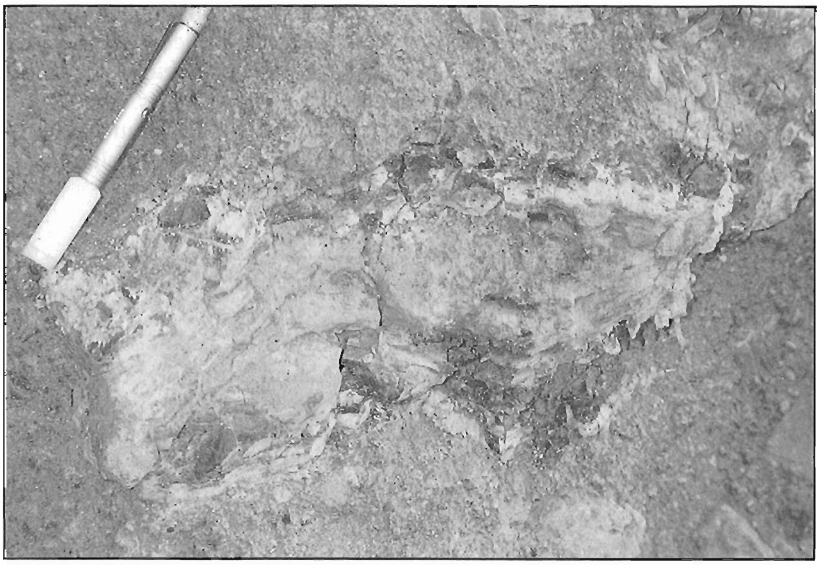
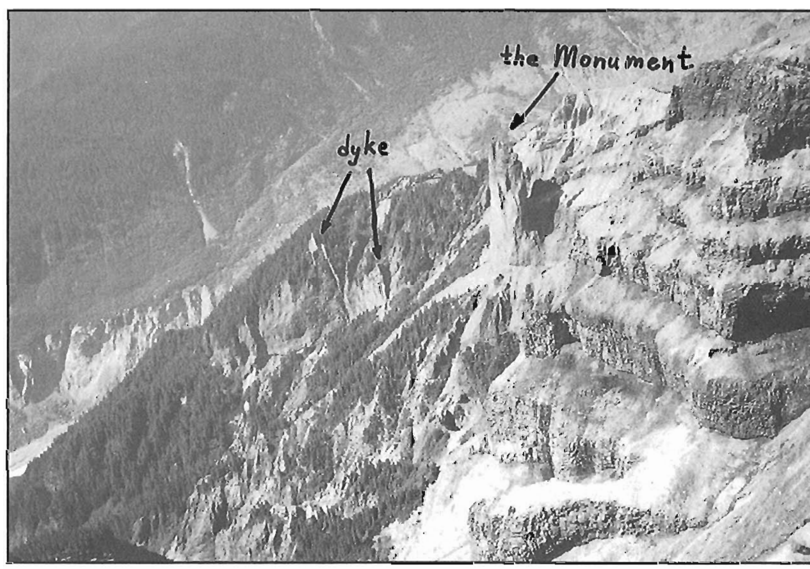


Figure 7.

Glass rimmed bomb of grey phonolite in hyaloclastite. Pencil flare is 18 cm long.

Figure 8.

Dyke (arrows) emanating from a lava spine approximately 150 m high, referred to as "the Monument" by Kerr (1948), represents an eroded volcanic neck which was the source for a dyke of grey phonolite, visible to the west (left) and downslope. The spine is in an extremely rugged canyon on the west central side of Hoodoo Mountain. Similar spines occur in several places above the 1300 m contour interval on Hoodoo Mountain.



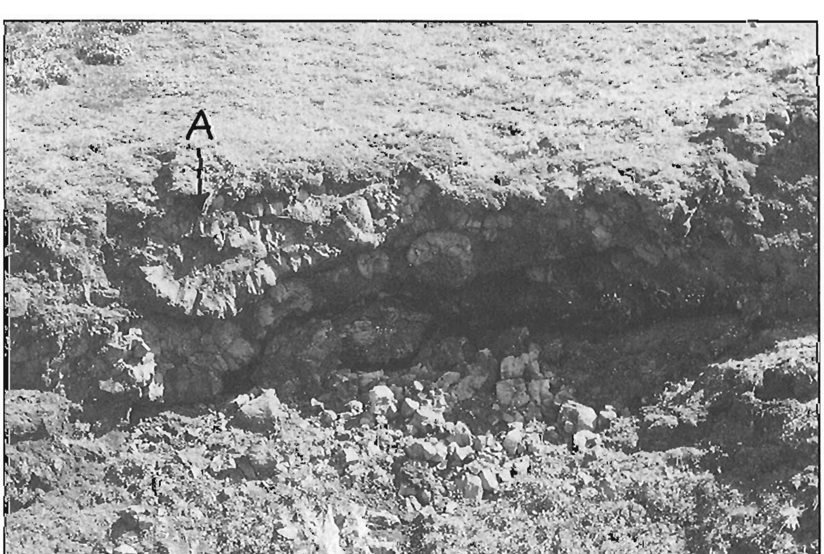


Figure 9.

Basaltic pillow lavas on the northwest flank of Little Bear mountain volcano. The large, elliptical pillow (A) is approximately 1 m wide.

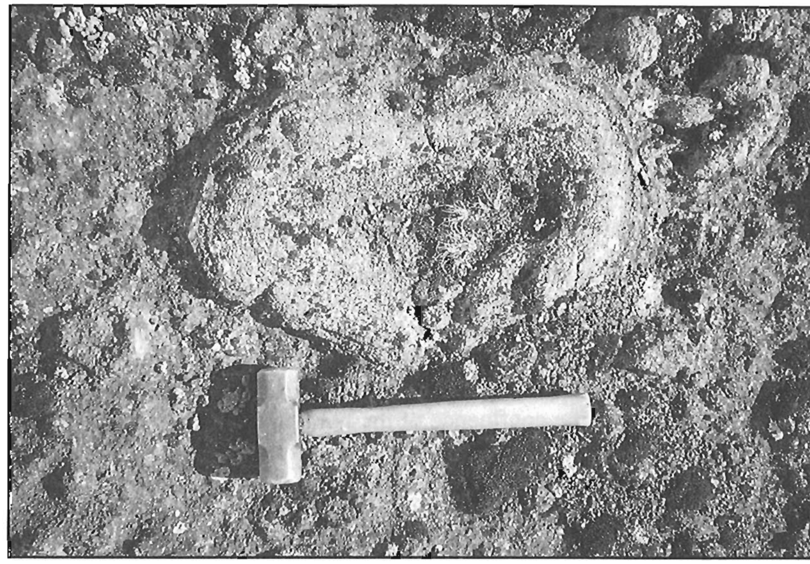


Figure 10.

Isolated basalt lava pillow surrounded by hyaloclastite matrix. Most of the exposed rocks on Little Bear mountain are basaltic breccias containing and devoid of lava pillows. Hammer is 38 cm long.

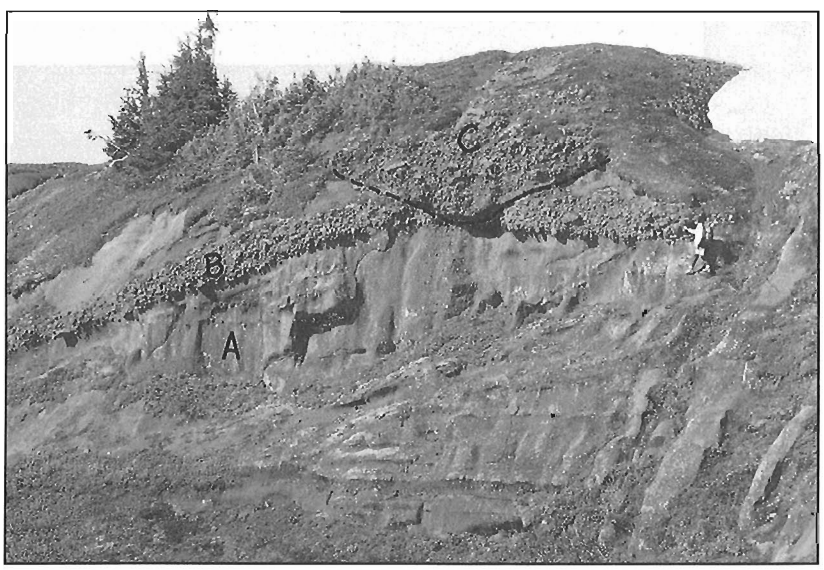


Figure 11.

Clast-filled channel of basaltic breccia (C) crosscutting breccia layer (B) within basaltic hyaloclastite (A), southeastern Little Bear mountain. Exposures of jointed, matrix-rich, poorly-bedded hyaloclastite are common on the southern and eastern flanks of the volcano. Person is 2 m tall.

10 m thick and is conformably overlain by a subglacial lava flow and associated hyaloclastite which is in turn overlain by the upper, greyish tan diamictite (Fig. 15).

Along the south base of Little Bear mountain is another greyish tan, matrix-supported diamictite, which overlies basaltic breccia. Round clasts of green quartzite, metavolcanic rocks and vesicular basalt make up 15-20 per cent of the unit and occur in a fine sand to coarse silt matrix. The diamictite also locally has irregular, flaggy horizontal joints and locally, flattened clasts are oriented parallel to the jointing. On the south fork of upper Hoodoo River, the diamictite rests on top of a basaltic breccia on both the north and south stream banks and is overlain by one of the most recent lava flows from Hoodoo Mountain on the south bank (see Fig. 12). This diamictite ranges from only a few metres thick on most of Little Bear mountain to 10 m thick on the south fork of upper Hoodoo River. Based on stratigraphic position, colour and clast type, it may be correlative with the upper tillite of Hoodoo Mountain. Both of these units are interpreted as lodgment tills based on: (1) the abundance of matrix over clasts, (2) subhorizontal jointing, (3) local clast imbrication, and (4) bullet-shaped clasts.

Rhythmites

Variegated tan, brown, green and yellow, rhythmically interbedded mud and sands occur at two different locations along the upper Hoodoo River. The first is 200 m west of lake Hoodoo, on the northeastern edge of NW flow (Fig. 1, location 5). Here the rhythmite sequence, comprising mudstones at least 1 m thick composed mainly of silt (<0.062 mm) with some clay size material, is underlain by a coarse, gravel-rich diamictite and is overlain by gravelly black sand and brown to orange silt. The rhythmite has laminations that are less than 1 mm thick and vary in colour from brown to grey-green to yellow (Fig. 15). The upper contact with the black sand is sharp and locally the upper part of the rhythmite has centimetre-scale, gravel-filled load structures. Micro-thrust faults disrupt some of the beds. The rhythmite appears to have been deposited in a small basin formed by blockage of the stream drainage by a lava from Hoodoo Mountain. Three different terrace levels are obvious at this location: two stratigraphically above the rhythmites and another at the base. On the west end of the paleobasin a levée of the Little Bear mountain diamictite dips steeply to the east, from the upper terrace down to a level below the rhythmites.

The second rhythmite occurrence is laterally more extensive than the first. At two localities multiple sets of rhythmically alternating sand and silt beds occur within a thicker package of poorly sorted sand. These units are found in a large basin topographically below NW flow but above Hoodoo Glacier, near the confluence of the north and south

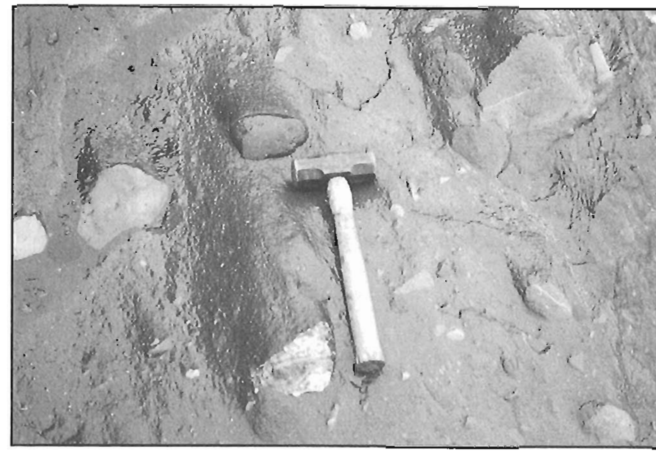


Figure 13. Tillite with a variety of metamorphic clasts. Polymictic diamictites, exposed in several places in a basin which is northwest of Hoodoo Mountain, are interpreted as lodgment tillites. Sledge hammer is 38 cm long.

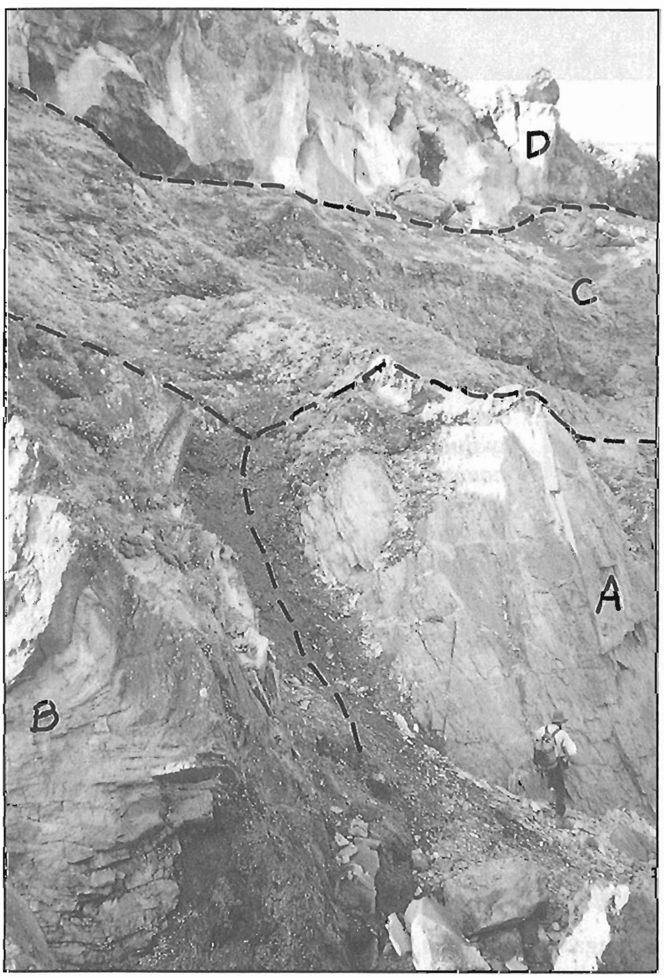


Figure 12. Stuhini Group(?) rocks (A) and Little Bear mountain basaltic breccia (B) are shown unconformably overlain by tillite (C) in a narrow gorge cut by the south fork of upper Hoodoo River along the northern edge of a recent porphyritic trachyte aa lava flow (D) from Hoodoo Mountain volcano. From top to bottom the sequence is Stuhini Group (A), basaltic breccia (B), tillite (C), and porphyritic trachyte (D). The country rock is a crenulated phyllite and is presumably part of the Stuhini Group (Fillipone and Ross, 1989; Anderson, 1993). Person at the base of the Stuhini Group exposure is 2 m tall.

Figure 14. Vesicular, highly plagioclase-phyric clasts in diamictite. Basaltic clasts in tillites 2 km southwest of Little Bear mountain volcano resemble flow rocks from that volcano. Hand flare is 18 cm.

forks of upper Hoodoo River (Fig. 1, locations 6 and 7). Two rhythmite sets occur near the base of a stream cut along the north fork of upper Hoodoo River (location 6), and three are found along the south fork (location 7).

The two sets of rhythmically layered sand and silt/clay on the north fork are exposed in a 15 m high stream bank (Fig. 16), composed mainly of poorly laminated, black sandy gravel (average size is 2 mm) which contains wood fragments.

On the south fork of upper Hoodoo River rhythmites are exposed in a 37 m high stream bank. The top of the upper rhythmite set is 31 m above the base and is 0.7 m thick. It is overlain by a dark, moist, gravelly sand, which also occurs between the other two rhythmite sets. The individual beds are generally less than 1 mm thick and have been variably folded and faulted on a centimetre scale. The second rhythmite set is at a height of 25 m from the bottom of the stream bank, consists mainly of alternating light and dark sand size grains and is 0.64 m thick. The top of the lowest rhythmite set is 22 m from the bottom and 0.5 to 1 m thick. It consists of alternating dark beds of sand size grains and light beds of silt to clay size grains. The base of the bank comprises tan, matrix-supported, polymictic diamictite. On the top of the stream bank and 12 m to the south, a flow of porphyritic trachyte appears to overlie the rhythmite deposit.

Figure 15. *Glaciolacustrine rhythmites, 200 m west of lake Hoodoo. The green and tan rhythmically layered, fine grained sediments occur on the northeast side of a Hoodoo Mountain lava flow. They are interpreted as having formed when the stream draining lake Hoodoo was temporarily blocked by lava flows. The informal name, "lake Hoodoo" was given to the small, glacier-fed lake on a spur of Twin Glacier, between Hoodoo and Little Bear mountains (Fig. 1). Bear spray is 22 cm tall.*

DISCUSSION

Revised stratigraphy

The new unit descriptions and stratigraphy provide further insight into previous stratigraphic work for Hoodoo Mountain and may also provide stratigraphic linkage between Hoodoo Mountain and Little Bear mountain volcanoes (Fig. 17).

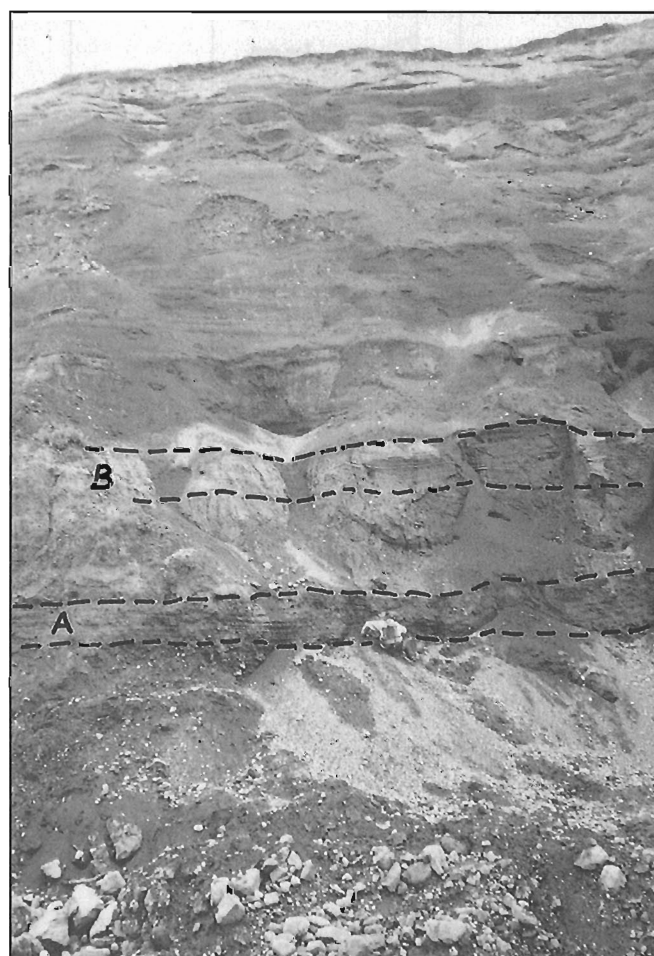


Figure 16. *Glaciolacustrine rhythmites on the north fork of "upper" Hoodoo River. Two sets of rhythmically bedded silt and sand (A and B) occur in the lower one third of this stream bank. The rhythmites are overlain by poorly bedded gravelly sand and underlain by stream gravel and till. Person is 2 m tall.*

The basement to the Hoodoo Mountain volcanic complex comprises Paleozoic and Mesozoic metavolcanic and metasedimentary units, including the Upper Triassic Stuhini Group (Kerr, 1948; Fillipone and Ross, 1989; Anderson, 1993). The basement rocks were intruded by three distinct dyke sets including grey phonolite (Edwards and Russell, 1994a) whose affinities to Hoodoo and Little Bear mountain volcanoes are unknown.

The oldest unit of the Hoodoo Mountain volcanic complex comprises highly jointed, orange weathering, locally flow banded aphanitic trachyte and local interbedded hyaloclastite found on the north, west, and possibly on the southwest flanks of Hoodoo Mountain (Kerr, 1948). The trachyte is overlain by a second series of microporphyritic trachyte lava flows, which form cliffs over 200 m high on the northeast and west sides of Hoodoo Mountain. No direct stratigraphic link between this oldest unit at Hoodoo Mountain volcano and the volcanic rocks at Little Bear mountain volcano was observed.

The cliff-forming microporphyritic lava flows are products of the next phase of volcanic activity at Hoodoo Mountain. On the north flank of Hoodoo Mountain they are unconformably overlain by the Pointer ridge pyroclastic deposit and another series of aphanitic to slightly porphyritic, black trachyte lava flows. On the western side of Hoodoo Mountain, the cliff-forming microporphyritic lavas overlie hyaloclastite associated with the oldest exposed lavas, and are

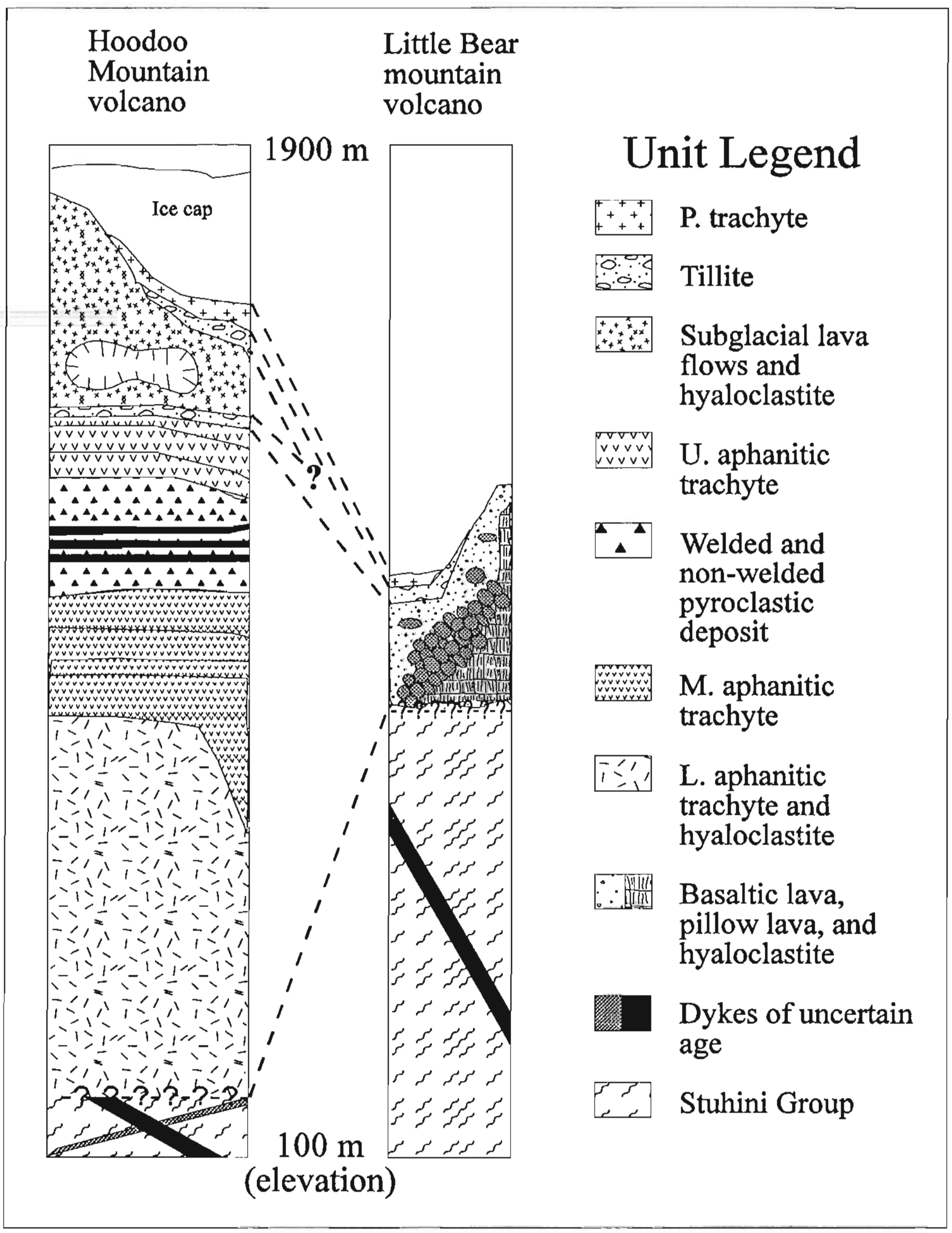


Figure 17. *Revised stratigraphy for Hoodoo Mountain volcanic complex.*

overlain in turn by a thick, nonresistant yellow unit, multiple aphanitic lava flows, and a thick sequence of intermingled lava and hyaloclastite. The thick yellow unit probably formed during an interval of either subglacial eruptions or extensive mass wasting and is possibly correlative with the Pointer ridge pyroclastic deposit. The cliff-forming lavas are also crosscut by a dyke associated with a spine of grey phonolite.

On the north and east sides of Hoodoo Mountain volcano, pyroclastic flows are overlain sequentially by a third series of weakly porphyritic lava flows, a tillite, subglacial lava flows and hyaloclastite, tillite and finally the youngest lava flows which are highly porphyritic anorthoclase trachytes.

No maximum stratigraphic position can be assigned to the Little Bear mountain volcano, but several field relationships indicate that it is older than many of the Hoodoo Mountain volcano lavas. The basaltic breccia is overlain by both glacial till and porphyritic trachyte at the southern edge of Little Bear mountain. Along the eastern margin of Hoodoo Glacier, clasts of basalt, very similar in appearance to basalt of Little Bear mountain volcano, occur in tillite that is overlain by grey phonolite.

SUMMARY

Detailed mapping during 1994 at Hoodoo Mountain volcanic centre revealed new information about the stratigraphy of the centre. The discoveries made included: (1) new links between hyaloclastite deposits and peralkaline, subglacial lava flows; (2) basaltic pillow lavas on Little Bear mountain volcano; and (3) tillites and glaciolacustrine sediments within the volcanic stratigraphy. These and other new observations lead to a revision of the stratigraphy of the Hoodoo Mountain volcanic centre as follows: (1) at least five periods of subaerial and subglacial lava extrusion occurred at Hoodoo Mountain volcano, (2) tillites with clasts of basalt from Little Bear mountain volcano underlie hyaloclastite from the subglacial lava flows, indicating that Little Bear mountain is older than the subglacial Hoodoo lava flows, and (3) dykes of grey phonolite cut across at least the lower three sets of Hoodoo lava flows. The stratigraphy for this complex indicates that several periods of volcanism have been interspersed with regional glaciation.

ACKNOWLEDGMENTS

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Geological Survey Project Number 840046

Investigations in the Dorsey terrane, Part 1: stratigraphy, structure, and metamorphism in the Dorsey Range, southern Yukon Territory and northern British Columbia[1]

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Stevens, R.A. and Harms, T.A., 1995: Investigations in the Dorsey terrane, Part 1: stratigraphy, structure, and metamorphism in the Dorsey Range, southern Yukon Territory and northern British Columbia; in Current Research 1995-A; Geological Survey of Canada, p. 117-127.

Abstract: In the Dorsey Range, the informally named Dorsey assemblage and the mid-Permian Ram Stock characterize the Dorsey terrane. The Dorsey assemblage can be subdivided into three units. The structurally lowest unit, which is intruded by the top of the Ram Stock, consists of greenschist to amphibolite facies siliciclastic rocks and interleaved metaplutonic rocks. The middle unit, in part Pennsylvanian, consists of greenschist to sub-greenschist facies phyllite, chert, tuff, siliciclastic, volcanic and carbonate rocks.The upper unit consists of sub-greenschist facies epiclastic, tuffaceous, siliciclastic, carbonate, and volcanic rocks. Penetrative ductile deformation in the lower unit grades to localized and more brittle deformation in the upper unit.

The base of the Ram Stock is in thrust-shear contact with an underlying package of rocks referred to as the imbricate assemblage. Tectonic fabrics in the Ram Stock and imbricate assemblage are cut by an Early Jurassic(?) pluton.

Résumé : Dans le chaînon Dorsey, l'assemblage informel de Dorsey et le Stock de Ram du Permien moyen caractérisent le terrane de Dorsey. L'assemblage de Dorsey peut être subdivisé en trois unités. La plus basse unité, sur le plan structural, que recoupe le sommet du Stock de Ram, consiste en des roches silicoclastiques allant du faciès des schistes verts au faciès des amphibolites, ainsi que des roches métaplutoniques interfoliées. L'unité du milieu, en partie pennsylvanienne, consiste en phyllades, chert, tuf et roches silicoclastiques, volcaniques et carbonatées allant du faciès des schistes verts au faciès des schistes verts inférieur. L'unité supérieure consiste en roches épiclastiques, tufacées, silicoclastiques, carbonatées et volcaniques du faciès des schistes verts inférieur. Une déformation ductile pénétrative dans l'unité inférieure passe graduellement à une déformation localisée et plus cassante dans l'unité supérieure.

La base du Stock de Ram est en contact par chevauchement-cisaillement avec un groupe de roches sous-jacentes appelé l'assemblage imbriqué. Un pluton du Jurassique précoce(?) recoupe la fabrique tectonique dans le Stock de Ram et l'assemblage imbriqué.

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INTRODUCTION

The allochthonous Dorsey terrane (Fig. 1) as defined by Monger et al. (1991) is a distinctive succession of upper Paleozoic sedimentary rocks that overlies rocks assigned to the Slide Mountain terrane. It is characterized by abundant chert, and carbonate and siliciclastic rocks, and by a lack of volcanic rocks. The Dorsey terrane is considered similar to the Slide Mountain terrane and the Harper Ranch subterrane of Quesnellia, but differs from them by its lack of volcanics (Monger et al., 1991). Abbott (1981) presented a different interpretation for the tectonic affinity of rocks in the central part of the Dorsey terrane (Dorsey Range; Fig. 1). According to Abbott (1981), the Dorsey Range is underlain by rocks of the Yukon Cataclastic Complex (Nisutlin subterrane of Gabrielse et al., 1991), and volcanic and sedimentary rocks at the base of the entire succession are part of the allochthonous Anvil Assemblage of the Yukon Cataclastic Complex, rather than the Slide Mountain terrane. Recent field work in the Teslin map area (105C), by Harms (1992) and Gordey and Stevens (1994a,b) has led them to interpret rocks assigned to the Dorsey and Slide Mountain terranes, in the Teslin area, as North American strata of the Cassiar terrane.

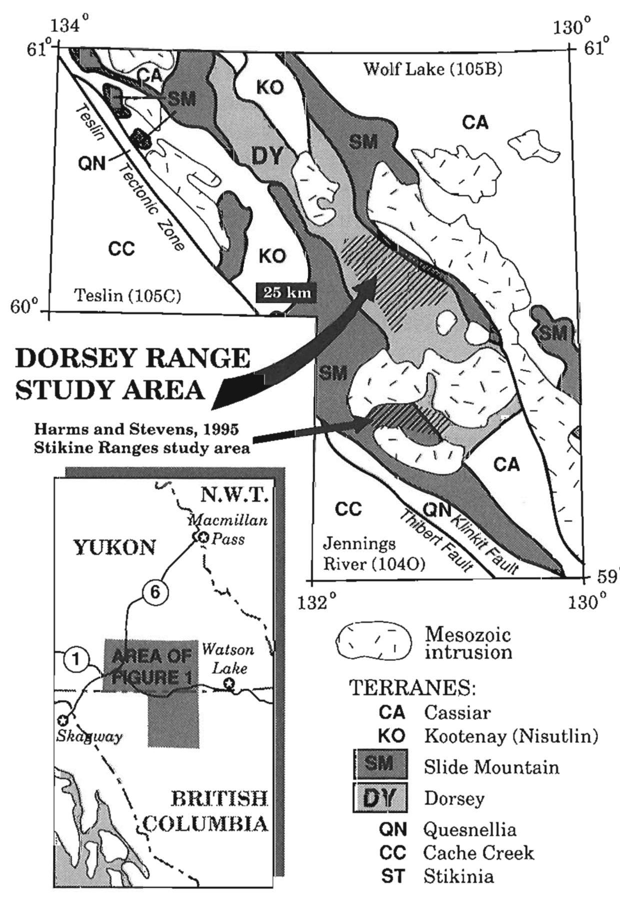

Figure 1. *Location map and tectonic framework of the project area.*

During the 1994 field season, Lithoprobe-supported mapping, as part of the Slave – Northern Cordillera Lithoprobe transect, was aimed at evaluating the tectonic significance of strata assigned to the Dorsey terrane, and adjacent Slide Mountain terrane, and at locating the suture zone between Paleozoic rocks of the North American margin and the allochthonous terranes of the Cordilleran Orogen. Field work was focused in two principal areas (Fig. 1): the Dorsey Range in the Wolf Lake (105B) and Jennings River (104O) map areas (this report), and in the Stikine Ranges in Jennings River map area (Harms and Stevens, 1995).

Our mapping in the Dorsey Range has established a new stratigraphic, structural and metamorphic framework for the central part of the Dorsey terrane, and has resolved the nature of several critical contacts. This paper summarizes the results of this field study by presenting lithological sections, and by discussing the important map relationships, the metamorphism and structure, and preliminary tectonic implications for the significance of rocks in the Dorsey Range.

LITHOSTRATIGRAPHIC UNITS

In the Dorsey Range, Poole (1956) identified three members, which according to Monger et al. (1991) characterize the Dorsey terrane. These members were interpreted by Poole (1956) to conformably overlie volcanic and sedimentary rocks of the Sylvester Group which, in turn, conformably and structurally rest above sedimentary rocks of the North American margin. The lowest member (Dorsey Group; unit 8 of Poole et al., 1960) consists of fine grained quartzite, grey to black argillite and slate, ribbon chert and discontinuous bands of limestone, cherty limestone and dolostone. The middle member (Screw Creek Formation; unit 9 of Poole et al., 1960) consists of Pennsylvanian, bedded, cherty limestone, locally more than 300 m thick. The upper member (Partridge Group; unit 10 and 11 of Poole et al., 1960) consists of a lower unit (>300 m) of conglomerate, grit, quartzite, sandstone and argillite and a upper unit (>500 m) of ribbon chert, argillite, slate and minor limestone.

We have identified different map units in the Dorsey Range (Fig. 2). Our units are essentially the same as those described by Abbott (1981), but our correlations differ from his. Rocks assigned to the Dorsey terrane by Wheeler and McFeely (1991) are included here in the informal Dorsey assemblage, which consists of three units that are of Paleozoic age or older (Fig. 2). The lower unit contains metamorphosed and deformed siliciclastic sedimentary rocks and tonalitic to dioritic plutonic rocks. The middle unit comprises two laterally distinct facies: to the southeast are black argillite to phyllite, quartzite, chert and minor siliciclastic rocks (middle phyllite/chert unit), whereas to the northwest are cherty carbonate rocks, volcanic rocks, tuff, chert and minor clastic rocks (middle volcanic/carbonate unit). The upper unit which is confined to the area southwest of the Seagull Batholith (Fig. 2) consists of epiclastic and siliciclastic rocks, carbonate rocks, tuff and minor volcanic rocks. The Dorsey assemblage includes the Dorsey Group of Poole (1956), and also the large

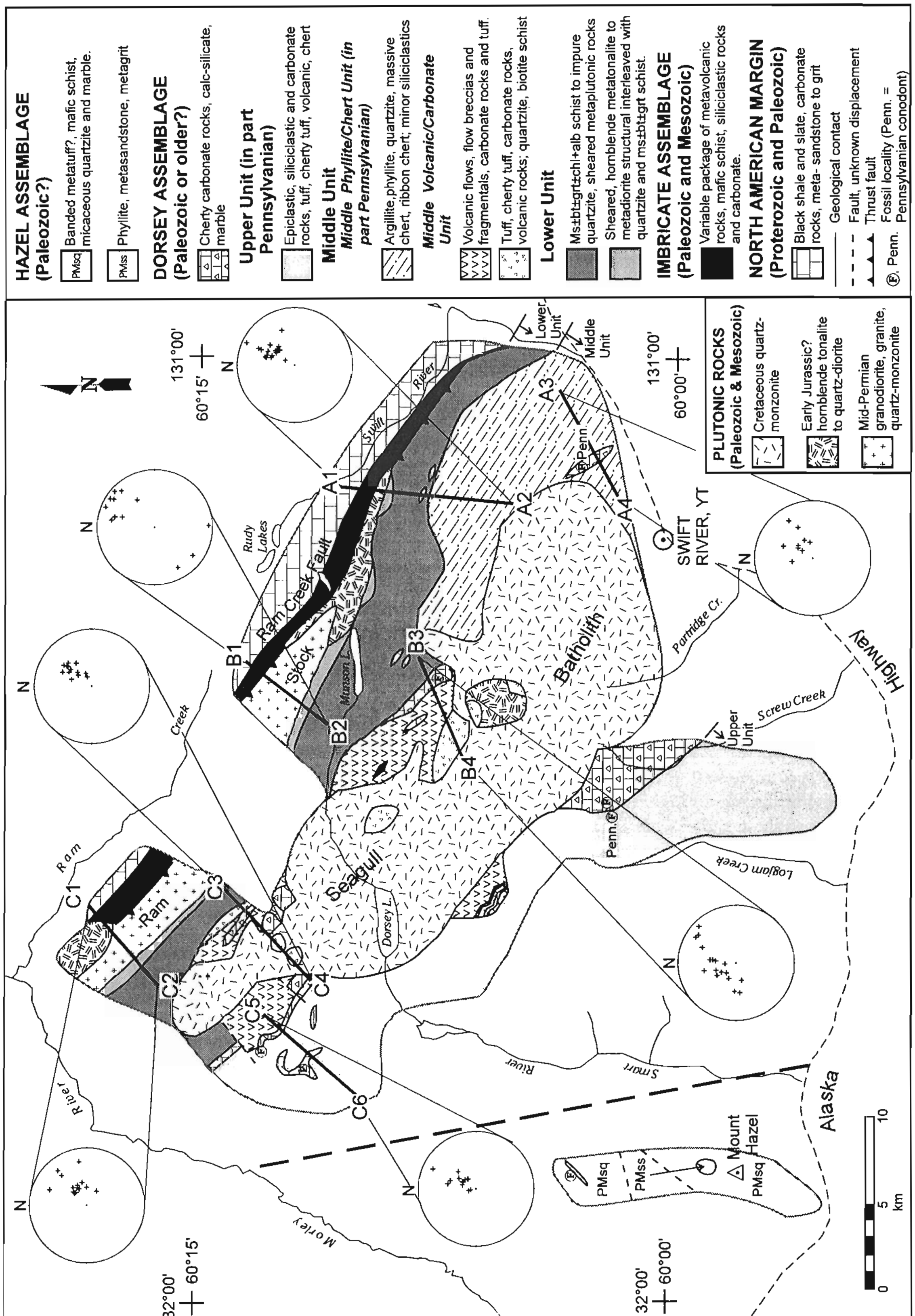

***Figure 2.** Simplified geology map of the Dorsey Range, southern Yukon Territory and northern British Columbia. Scatter plots are poles to planar fabrics (primary and tectonic). Geology by Abbott (1981) and this paper.*

band of carbonate rocks at the headwaters of Screw Creek (Fig. 2) which is part of the Screw Creek Formation of Poole (1956).

Rocks that structurally underlie the Dorsey assemblage and that were interpreted by Poole (1956) as part of the Sylvester Group (Slide Mountain terrane of Wheeler and McFeely, 1991) are referred to here as the imbricate assemblage (Fig. 2). The imbricate assemblage includes Paleozoic and/or Mesozoic metamorphosed, volcanic, plutonic and sedimentary rocks.

Southwest of the Dorsey Range, around Mount Hazel (Fig. 2), a separate and distinct package of metamorphosed green and yellow banded biotite porphyroblastic tuffs?, carbonate rocks, chlorite schist, pink micaceous quartzite, and dirty brown sandstone and grit are referred to as the Hazel assemblage. The Hazel assemblage is part of the Sylvester Group of Poole (1956).

Simplified lithological sections through the Dorsey Range (Fig. 3, 4, 5) are presented to summarize the principal rock types in each assemblage and to show the nature of along-strike variations. The important structural relationships within and between the sections are discussed below.

STRUCTURAL RELATIONSHIPS

In the Dorsey Range planar fabrics (primary and tectonic) dip dominantly to the southwest. The structurally lowest assemblage exposed in the Dorsey Range occurs to the northeast and consists of Upper Proterozoic and lower Paleozoic miogeoclinal rocks of the continental margin of North America (Cassiar terrane of Wheeler and McFeely, 1991). Description and correlation of these rocks is contained in Abbott (1981) and in Poole (1956). The top of the North American assemblage comprises black graphitic shale and siliceous graphitic siltstone. These rocks are in sharp contact along a moderate to steeply southwest-dipping fault surface (Ram Creek fault of Poole, 1956) with rocks of the imbricate assemblage. The Ram Creek fault is marked by changes in rock type, metamorphic grade, degree of deformation and locally in fabric orientations. Sense of movement indicators along the Ram Creek fault were not observed, however it is interpreted as a thrust fault because it juxtaposes greenschist facies metamorphic rocks over sub-greenschist facies sedimentary rocks.

The imbricate assemblage is dominated by mafic schist and metavolcanic rocks but also includes variably deformed metaplutonic and metasedimentary rocks. Individual rock units or groups of units are discontinuous laterally on the scale of 1-5 km. Along much of its length the top contact is a distinct mylonitic shear zone or strongly foliated zone, involving rocks of the Ram Stock and the imbricate assemblage. Steeply-plunging mineral lineations are preserved in this high strain zone. Near the eastern edge of the map area, south of Swift River, the contact was not observed; however, rocks of the lower unit of the Dorsey assemblage, which lie structurally above the imbricate assemblage in this area (see Fig. 2) have been metamorphosed to a higher grade. The shearing and mylonitization along the contact, the plunge of the lineations, and the metamorphic juxtaposition suggest that the contact is a thrust zone, and that the Dorsey assemblage has been thrust over the imbricate assemblage. The most recent deformation of the imbricate assemblage and most recent movement along this thrust zone is constrained to post-mid-Permian time since the Ram Stock which is affected by the deformation crystallized at 259 ± 2 Ma (zircon date; J.K. Mortensen, pers. comm., 1994).

The Ram Stock is a resistant, homogeneous body of granodiorite, granite and quartz-monzonite situated between the imbricate and Dorsey assemblages. It is massive or weakly deformed with a few small (<2 m wide), discontinuous shear zones throughout. The base of the Ram Stock (northeast contact), where clearly exposed, is a distinct 10-20 m wide mylonitic shear zone in which mylonite grades abruptly upwards into massive granodiorite. The top of the Ram Stock is an intrusive contact with the lower unit of the Dorsey assemblage. The intrusive nature of this contact is clearly demonstrated by the occurrence of inclusions of foliated rock of the Dorsey assemblage within massive rock of the Ram Stock, and by dykes and offshoots of massive rock of the Ram Stock cutting foliated and metamorphosed rock of the Dorsey assemblage (Fig. 6). This indicates that youngest penetrative deformation in the lower unit of the Dorsey assemblage, above the Ram Stock, is pre-mid-Permian. This contrasts with the imbricate assemblage, below the Ram Stock, in which the youngest deformation is post-mid-Permian. The foliated zone at the base of the Ram Stock and the tectonic deformation fabrics in the imbricate assemblage are cut by a massive hornblende-bearing tonalite to quartz-diorite pluton in the northwestern part of the map area near the confluence of the Ram Creek and Morely River (Fig. 2). Regional correlation with similar bodies to the west (Stevens et al., 1993; Gordey and Stevens, 1994a,b) and elsewhere in the Yukon (Mortensen, 1992) suggests that this body is Early Jurassic in age. If so, deformation in the imbricate assemblage and Ram Stock had ceased by the Early Jurassic.

The lower unit of the Dorsey assemblage lies southwest of the Ram Stock and includes two members: a lower member of structural interleaved intermediate to mafic metaplutonic rocks and quartzite to metapelite, and an upper siliciclastic member of quartzite to metapelite (Fig. 7). The lower unit is continuous along the length of the map area (Fig. 2) and shows little variation. The most conspicuous along-strike changes are: an increase to the northwest in the degree of interleaving of metaplutonic and metasedimentary rocks in the lower member, and an increase to the southeast in pelitic rocks (garnet schist) in the upper member. In the southeast part of the map area, along section line A1-A2 (Fig. 2), the lower unit is in contact with black graphitic phyllite and white quartzite of the middle phyllite/chert unit over a structurally conformable, steeply southwest-dipping zone. This contact is parallel to compositional layering and foliation in both units and does not involve a change in structural style or degree of deformation and metamorphism between the two units. Locally, black graphitic phyllite occurs as minor interbands in siliceous schist in the top few hundred metres of the lower unit. These relationships suggest that the contact is not a fault, but probably a stratigraphic or transposed stratigraphic

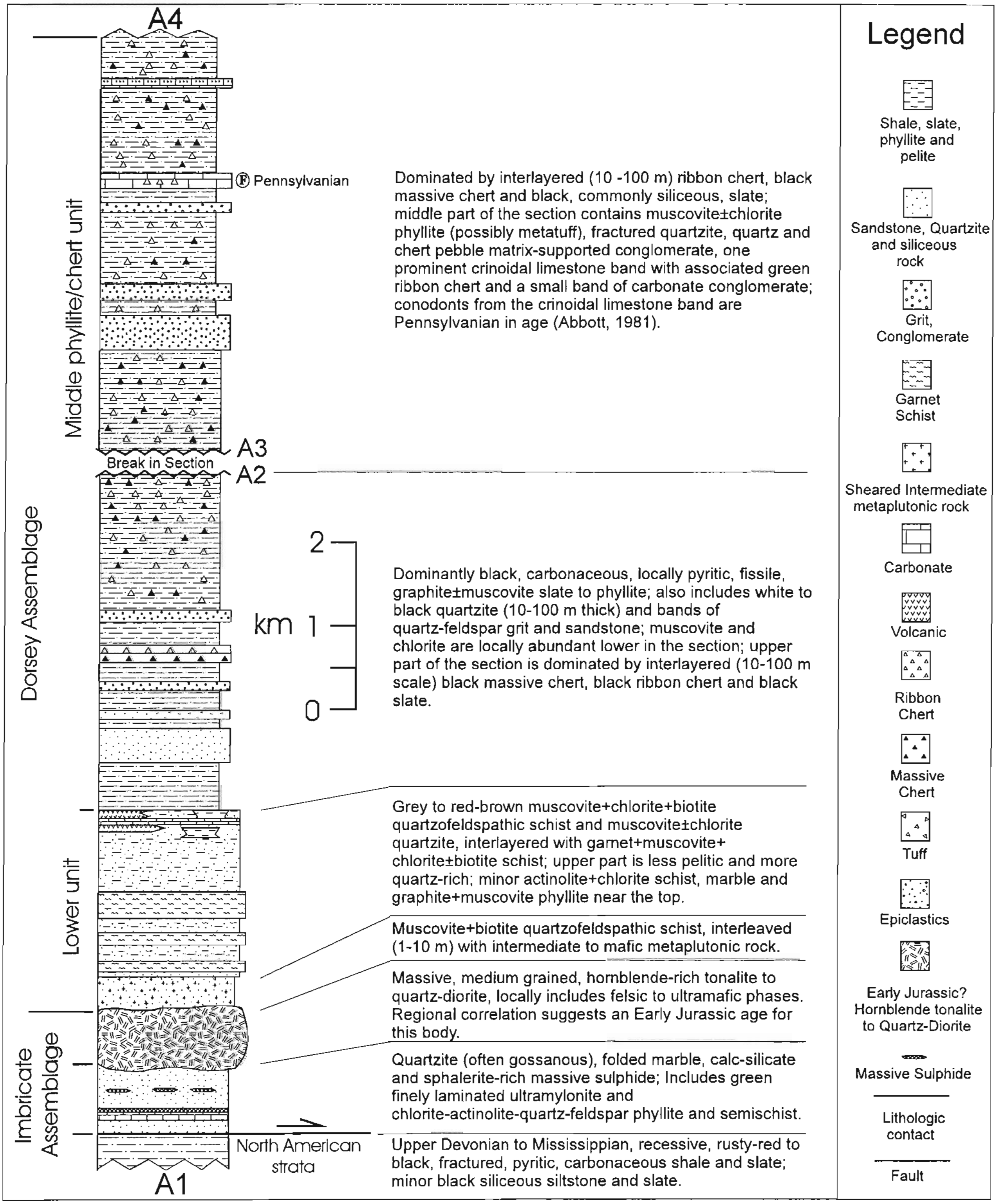

Figure 3. *Simplified lithological section and rock descriptions along section lines A1-A2 and A3-A4 (see Fig. 2 for section location). Unit thicknesses are approximate and were calculated by measuring the horizontal outcrop distance along the transect line and then correcting for dip of the strata. True stratigraphic thicknesses are unknown due to structural disruption. Massive sulphide near the base of the section is part of the Bar claim which is a zinc-skarn drilled prospect (DIAND, 1993).*

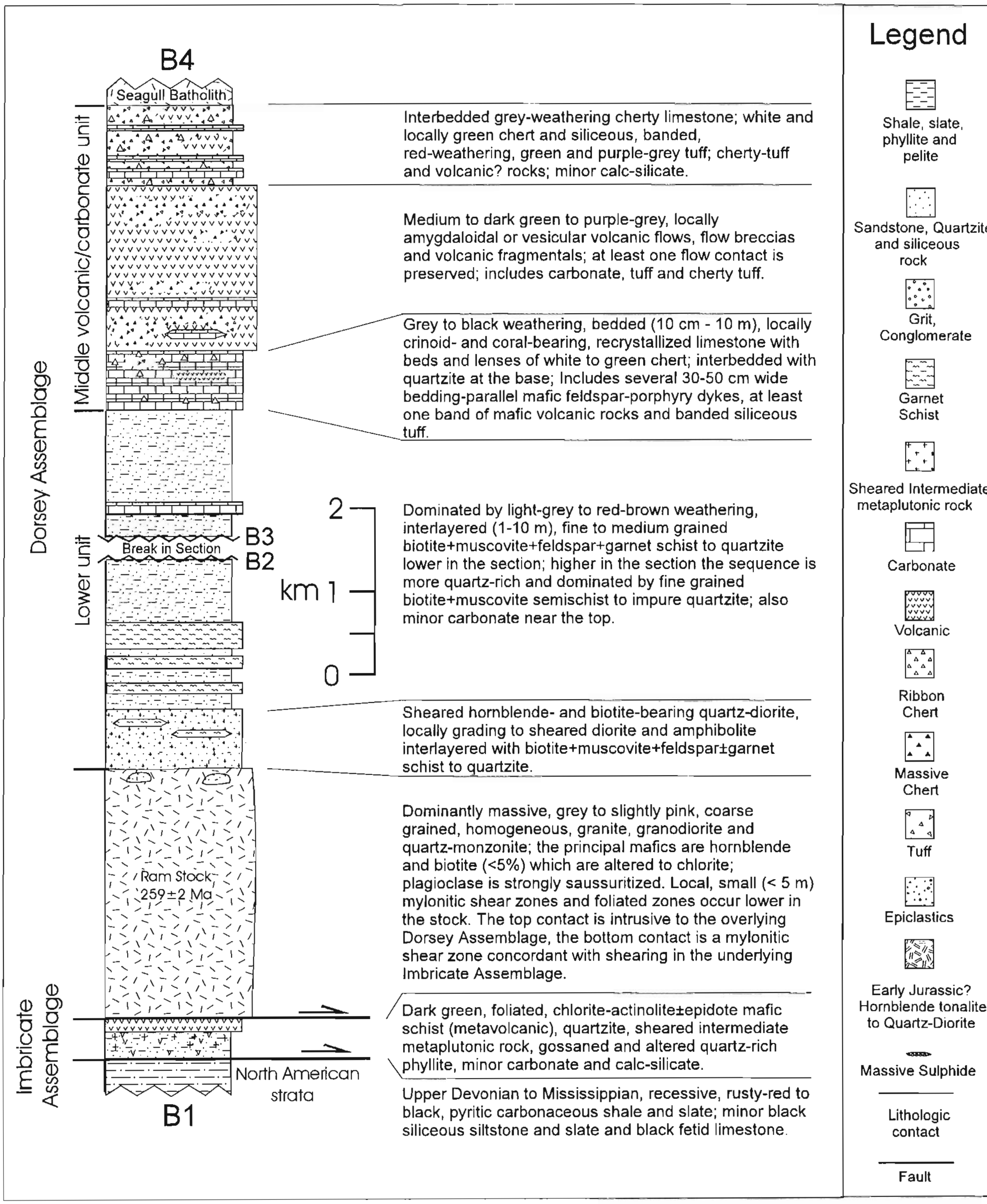

Figure 4. Simplified lithological section and rock descriptions along section lines B1-B2 and B3-B4 (see Fig. 2 for section location). Unit thicknesses are approximate and were calculated by measuring the horizontal outcrop distance along the transect line and then correcting for dip of the strata. True stratigraphic thicknesses are unknown due to structural disruption.

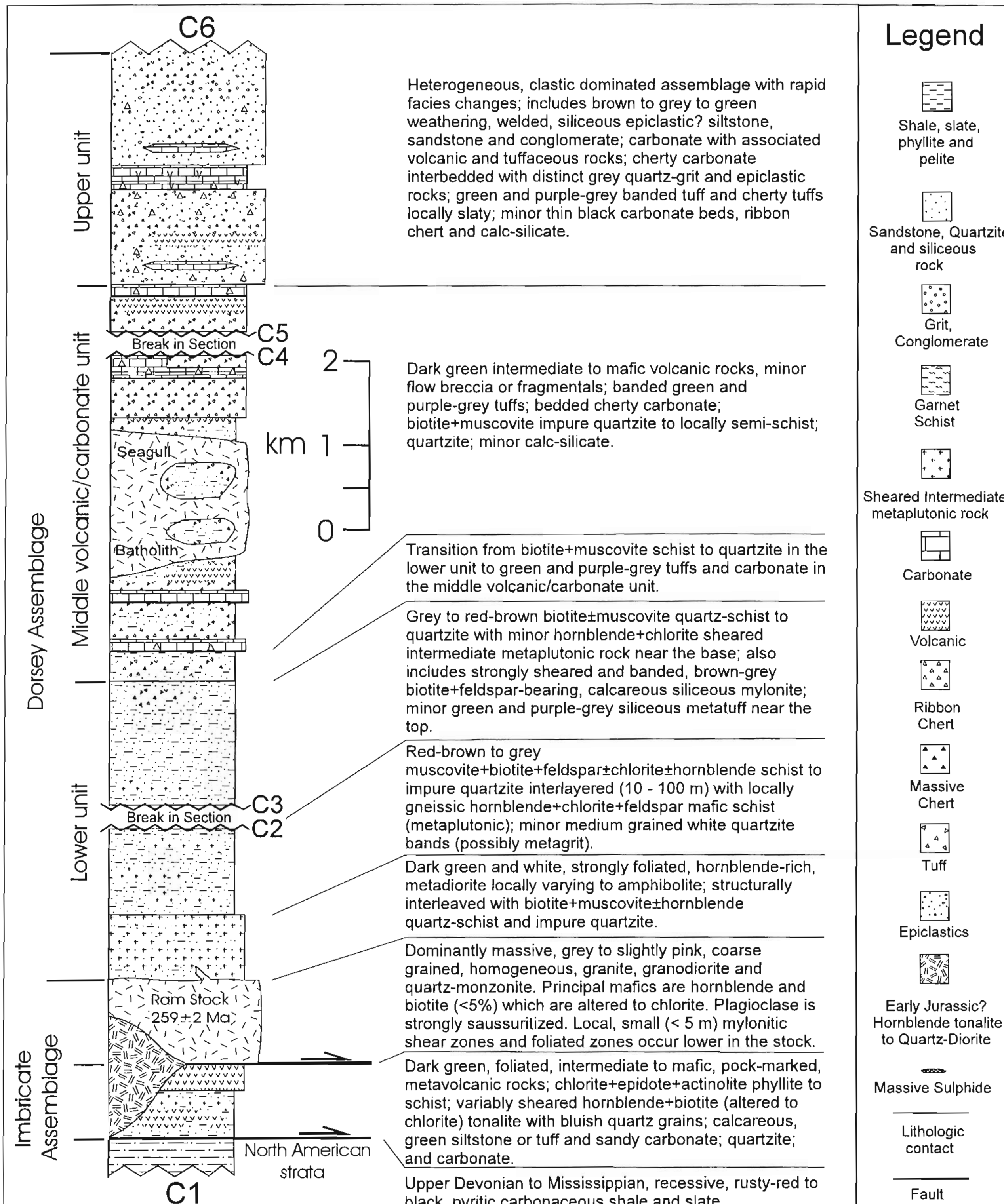

Figure 5. *Simplified lithological section and rock descriptions along section lines C1-C2, C3-C4, and C5-C6 (see Fig. 2 for section location). Unit thicknesses are approximate and were calculated by measuring the horizontal outcrop distance along the transect line and then correcting for dip of the strata. True stratigraphic thicknesses are unknown due to structural disruption.*

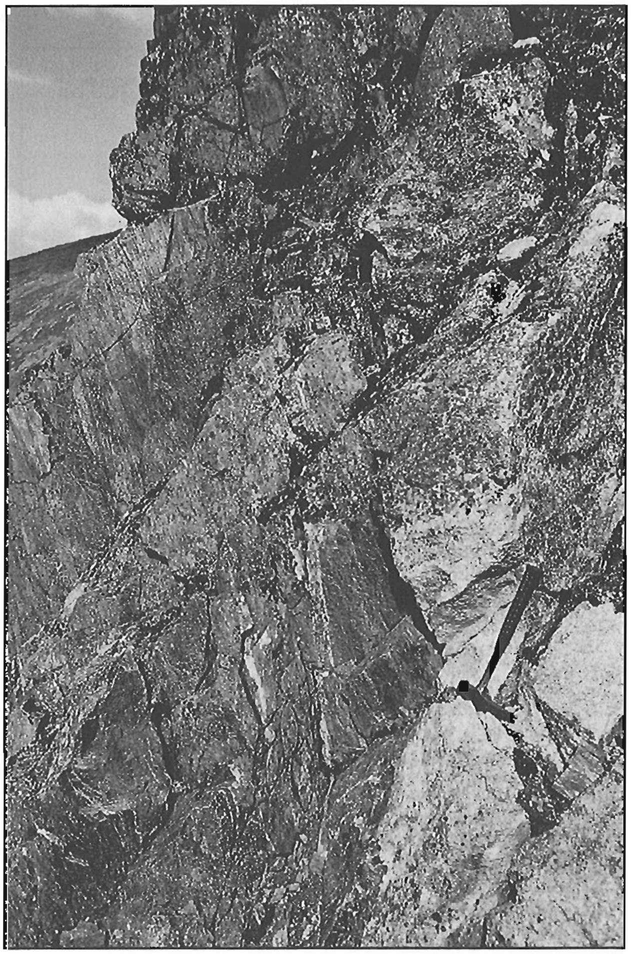

Figure 6. Intrusive contact between massive rock of the mid-Permian Ram Stock and metamorphosed and deformed sedimentary and plutonic rocks of the lower unit of the Dorsey assemblage. View to the northwest around section line C1-C2.

contact. To the northwest, along section line B3-B4, biotite quartz semi-schist and quartzite of the lower unit are interlayered with the lowest carbonate bed of the middle volcanic/carbonate unit. Along section C3-C4, the contact is gradational with green and purple-grey banded tuffs typical of the middle volcanic/carbonate unit occurring within biotite quartz semi-schist of the lower unit. These relationships suggest that the contact between the middle volcanic/carbonate unit and lower unit is also a stratigraphic or transposed stratigraphic contact.

The contact relationship between the middle phyllite/chert and volcanic/carbonate subunits is obscured by the Seagull Batholith (Fig. 2). However, the apparent lack of intercalation between the dominant rock types of the two units (e.g., little evidence of volcanism is preserved in the middle phyllite/chert unit) raises the possibility that a fault may lie between them, or conversely, one or both subunits may actually be in fault contact with the underlying lower unit.

The upper contact of the middle volcanic/carbonate unit is exposed northwest of Dorsey Lake (Fig. 2), where it is marked by the top of a bedded and laterally continuous carbonate horizon. Volcanic and tuffaceous rocks stratigraphically above the carbonate horizon are similar to those below it. However, to the southwest the volcanic and tuffaceous rocks occur with epiclastics, chert, quartz grit and sandstones typical of the upper unit, which suggest that rocks above the carbonate horizon are part of the upper unit.

The upper unit is characterized by internal heterogeneity of rock types, such that individual distinct beds or composite horizons cannot be followed along strike for more than a few kilometres. This heterogeneity is due to both rapid facies changes within the stratigraphic succession and to localized structural disruption. Although the unit is internally heterogeneous, the same suite of rocks can be traced along the length of the Dorsey Range, from the Alaska Highway to the Morely River. The top of the upper unit was not observed.

Figure 7.

Interlayered sequence of quartzite to metapelite in the lower unit of the Dorsey assemblage along section A1-A2. Planar deformation fabrics parallel to layering, and fabric asymmetry suggest that layer-parallel shearing dominated in the lower unit of the Dorsey assemblage.

Early Pennsylvanian conodonts have been obtained from a limestone band in the Dorsey assemblage within the upper part of the middle phyllite/chert unit, and from a thin carbonate band in the upper unit just west of the large carbonate exposure near the head waters of Screw Creek (Abbott, 1981) (Fig. 2). The age of the entire Dorsey assemblage is tentatively considered Pennsylvanian and older. However deformation or regional unconformities may have juxtaposed rocks of different ages, especially in the upper unit where internal heterogeneity of rock types is characteristic, and in the lower unit where deformation is the strongest.

A distinctive succession mapped southwest of the Dorsey Range near Mount Hazel, is referred to as the Hazel assemblage (Fig. 2). Based on the distinct change in rock types between these rocks and the upper unit of the Dorsey assemblage, a fault between the two is inferred. The Hazel assemblage includes two fault bounded(?) units of Paleozoic or Mesozoic age. One consists of locally strongly crenulated, banded, green and yellow biotite porphyroblastic schist (metatuff?), pink muscovite quartzite, muscovite-chlorite-feldspar±biotite±epidote±actinolite schist which is locally feldspar porphyroblastic and one layer of coralline micaceous marble. The second consists of dirty brown clastic rocks ranging from phyllite to gritty sandstone. Rocks around Mount Hazel were correlated with the Sylvester Group by Poole (1956).

METAMORPHISM

Rocks of the Cassiar terrane vary from low grade sub-greenschist facies shale and slate at the top of the succession near the Ram Creek fault, to greenschist facies schist farther northeast and close to the Cassiar Batholith. Abbott (1981) reported andalusite in some of these schists suggesting that an increase in metamorphic grade to the northeast is likely related to contact metamorphism around the batholith. Murphy (1988) also documented a strong contact metamorphic aureole around other mid-Cretaceous intrusions northeast of the Cassiar Batholith.

Above the Ram Creek fault, rocks of the imbricate assemblage have been metamorphosed to low-greenschist facies conditions with chlorite, muscovite, epidote and actinolite as the dominant metamorphic minerals.

Rocks of the Dorsey assemblage vary from amphibolite grade schist near the base of the lower unit to sub-greenschist facies sedimentary and volcanic rock near the top of the middle and upper units. Within 2-2.5 km of the contact with the Ram Stock coarse garnet-schist, lineated and foliated hornblende schist and metamorphically banded orthogneiss indicate that the rocks have been metamorphosed to amphibolite facies conditions. To the southwest, near the top of the lower unit and base of the middle unit, metamorphic grade apparently decreases to greenschist facies, with schist dominated by biotite, muscovite, chlorite and locally actinolite. Garnet and hornblende are absent in these schists. Further "upsection", near the top of the middle phyllite/chert unit, black ribbon chert is weakly recrystallized and locally appears cryptocrystalline, and epiclastics in the upper unit lack the development of any macroscopically visible metamorphic minerals.

Carbonate horizons throughout the Dorsey assemblage are recrystallized to some degree, but are more coarsely crystalline in the lower unit. A distinct metamorphic aureole around the Seagull Batholith was not observed except near section B3-B4 where tuffaceous and volcanic rocks are hornfelsed. However, within the Dorsey Range a number of carbonate horizons host skarn mineralization, particulary tin, which are thought to be associated with the Seagull Batholith (Abbott, 1981; Layne and Spooner, 1991). Calcite-garnet-diopside-epidote-tremolite calc-silicates are common within carbonate horizons and may be the product of regional metamorphism and/or skarn development associated with mid-Cretaceous and other plutons in the map area (Fig. 2).

Rocks of the Hazel assemblage have been metamorphosed to greenschist facies or lower. In the schist and quartzite unit (PMsq) rocks have well developed fine- to medium-grained metamorphic minerals, dominated by muscovite, biotite, chlorite, actinolite, epidote and albite. In the dirty sandstone unit (PMss) rocks appear less metamorphosed with weak development of biotite, muscovite and chlorite.

STRUCTURAL CHARACTER

Rocks of the Cassiar terrane preserve a variety of planar fabrics, including bedding, slaty cleavage and schistosity. Folding of shale and calcareous sandstone beds into a series of anticline/syncline pairs with an associated axial planar cleavage is evident west of Rudy Lakes. Discontinuity of stratigraphic units along strike suggests several faults cut the North American succession. For a more complete discussion of structure in rocks of the Cassiar terrane, northeast of the project area, see Murphy (1988).

In the imbricate assemblage rocks have been ductilely deformed and range from ultramylonites to weakly foliated metaplutonic rock. No distinct or significant faults are recognized, but the lack of along-strike continuity of units suggests structural disruption. The imbricate assemblage may be a collection of juxtaposed fault slices or slivers of varying sizes.

Rocks of the Dorsey assemblage preserve strong, penetrative, ductile fabrics at the base of the lower unit, but weaker, localized and more brittle fabrics at the top of the middle unit and in the upper unit. The lower unit is characterized by lineated protomylonite and mylonite, layer-parallel shear, isoclinal folding and preservation of shear-sense indicators (especially shear-bands). In the middle unit deformation varies from ductile and locally mylonitic near the base to discontinuous near the top with preservation of bedding (ribbon chert; Fig. 8) and volcanic flow contacts. Near the top, ribbon chert is commonly deformed by brittle fractures and folds. In the upper unit, deformation is localized and dominated by layer-parallel folding (commonly in carbonate layers) and by a flattening fabric (ellipsoidal quartz-grit grains with maximum 2:1 ratios) and by minor shear. Preservation of primary clastic textures is typical in the upper unit.

Figure 8.

Black ribbon chert from the middle phyllite/ chert unit of the Dorsey assemblage along section A1-A2. Phyllitic and slaty cleavage are parallel to bedding throughout most of the middle phyllite/ chert unit.

In the Dorsey assemblage bedding is parallel to cleavage and phyllitic foliation in the middle unit (Fig. 8) and layering of pelitic schist and quartzite is parallel to foliation in the lower unit (Fig. 7). These features combined with layer-parallel folds and asymmetric tectonic fabrics suggest that layer-parallel shear dominated in the Dorsey assemblage.

Throughout most of the map area planar fabrics dip moderately to steeply southwest (Fig. 2). Along section B3-B4 near the Seagull Batholith fabrics apparently mark the closure of a broad southeast-plunging synform. Previous mapping in the southern part of the Dorsey Range (Abbott, 1981; Gabrielse, 1968a,b) suggests that rocks lie in a regional southeast-plunging syncline. Our mapping indicates that foliation and bedding orientations may trace out a synform across the southern end of the Dorsey Range, but stratigraphy does not support the presence of a syncline. Fabrics dip consistently southwest across the northern end of the Dorsey Range (Fig. 2).

Rocks of the Hazel assemblage display a moderate to strong schistosity which is overprinted, in the northern part of the exposure, by strong crenulation folding with southerly plunging fold axes (20-30°). Bedding features (i.e., gradation from metagrit to metasandstone to metashale) are preserved in unit PMss.

DISCUSSION

Rocks of the imbricate assemblage and Hazel assemblage are interpreted to be part of the Sylvester Group by Poole (1956), and the Slide Mountain terrane by Wheeler and McFeely (1991). However the rock types in the Hazel assemblage and their metamorphic character suggest possible correlation to the Nisutlin subterrane of Kootenay terrane which is exposed immediately to the west in the Teslin area (Gordey and Stevens, 1994a,b). The imbricate assemblage lacks any diagnostic feature that could confirm its tectonostratigraphic correlation.

In the Dorsey assemblage, metamorphic grade and penetrative ductile deformation increase "down-section", which suggests that the Dorsey assemblage preserves an oblique-section through part of the upper crust. Currently, more than 15 km of section is preserved in the Dorsey Range, however the true stratigraphic thickness of this section is unknown as structural thickening or thinning likely occurred during deformation. The section is dominated by siliciclastic, continental margin-type rocks at the base that grade upward to a more oceanic or basinal sequence characterized by intermediate to basic volcanic rocks, carbonate rocks, chert and black shale, tuffaceous chert, and epiclastic rocks, at the middle and top.

The Dorsey assemblage was penetratively deformed and metamorphosed prior to the mid-Permian, intruded by the mid-Permian Ram Stock and then thrust over rocks of the imbricate assemblage between the mid-Permian and Early Jurassic(?). The imbricate assemblage is interpreted as a series of thrust slices that formed between the Dorsey assemblage and the North American margin as the Dorsey assemblage was thrust upwards. Pennsylvanian conodonts identified from the top of the middle phyllite/chert unit and from the upper unit (Abbott, 1981) (Fig. 2), together with evidence of pre-mid-Permian penetrative deformation in the lower unit suggest that the Dorsey assemblage may have been deformed between the early Pennsylvanian and mid-Permian.

Preliminary assessment of the Dorsey assemblage and Ram Stock suggest that although they are similar to rocks in other terranes (particularly the Kootenay and Slide Mountain terranes), they appear to have a tectonostratigraphy and history of metamorphism and deformation that is distinct from other terranes in the Canadian Cordillera. This feature combined with their apparent fault bounded position supports the concept of a separate Dorsey terrane.

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Investigations in the Dorsey terrane, Part 2: lithologies and structure of (?)Paleozoic stratified rocks in the Stikine Ranges, northern British Columbia[1]

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Harms, T.A. and Stevens, R.A., 1995: Investigations in the Dorsey terrane, Part 2: lithologies and structure of (?)Paleozoic stratified rocks in the Stikine Ranges, northern British Columbia; in Current Research 1995-A; Geological Survey of Canada, p. 129-133.

Abstract: Although presently divided between Slide Mountain and Dorsey terranes, (?)Paleozoic strata across the Stikine Ranges of northern British Columbia have demonstrable lithological similarities. Areas of the Stikine Ranges studied during the 1994 field season are dominated by clean quartz sandstone, fossiliferous and unfossiliferous limestone, and shale, which occur in laterally continuous, lithologically distinctive and homogeneous, mappable units. The character of this suite of rocks suggests a continental margin origin for the sequence, despite the presence of volcanic rocks in the domain assigned to the Slide Mountain terrane. These strata are deformed by both map scale and outcrop scale, tight to isoclinal, reclined to recumbent, similar folds that appear to be south-vergent. Full description of the stratigraphy and geological history of the Stikine Ranges and Dorsey terrane will require further detailed mapping of the units and structures of the area.

Résumé : Bien que présentement divisées entre les terranes de Slide Mountain et de Dorsey, les couches du Paléozoïque(?) dans les chaînons Stikine, dans le nord de la Colombie-Britannique, présentent des similarités lithologiques qui peuvent être démontrées. Les roches prédominantes dans les secteurs des chaînons Stikine étudiés durant la campagne 1994 sur le terrain sont un grès quartzeux pur, des calcaires fossilifères et non fossilifères et un shale, qui se présentent dans des unités continues latéralement, distinctives et homogènes sur le plan lithologique et que l'on peut cartographier. La nature de cette suite de roches suggère, pour la séquence, une origine de marge continentale, en dépit de la présence de roches volcaniques dans le domaine assigné au terrane de Slide Mountain. Ces couches sont déformées, à l'échelle cartographique et à l'échelle des affleurements, par des plis semblables, fermés à isoclinaux, réclinés à couchés, qui semblent être à vergence sud. Une description complète de la stratigraphie et du passé géologique des chaînons Stikine et du terrane de Dorsey nécessitera une cartographie plus détaillée des unités et des structures de la région.

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INTRODUCTION

Stratified rocks in the Stikine Ranges of northern British Columbia have been mapped as one unit and, although largely unfossiliferous, interpreted to be Paleozoic in age (Gabrielse, 1970; Monger et al., 1991). These rocks are presently assigned to the Dorsey terrane in the east half of the Stikine Ranges and to Slide Mountain terrane in the west (Fig. 1). The boundary between these two domains is based on the distribution of volcanic rocks, as known (Monger et al., 1991; Wheeler and McFeely, 1991), and has not, as yet, been recognized as a fault. The Stikine Ranges lie immediately west and north of the Cassiar terrane, where the westernmost sedimentary strata associated with North America's Paleozoic continental margin occur. In upcoming years, the Slave-Northern Cordilleran LITHOPROBE transect will cross the inboard part of the Cordillera along strike immediately to the north and the south of the Stikine Ranges. At present, however, there is no valid paleogeographic model for the rocks that lie in this critical transitional position between indigenous North America and accreted Paleozoic and Mesozoic oceanic and arc terranes to the west. Consequently, we have undertaken a project that seeks to establish the stratigraphy, structure, and geological history of those stratified rocks that predate Mesozoic intrusions within the Stikine Ranges of northern British Columbia (Fig. 2). We hope to provide an interpretation of the setting in which these sequences originated and thereby help to constrain models of crustal structure in this part of the transect.

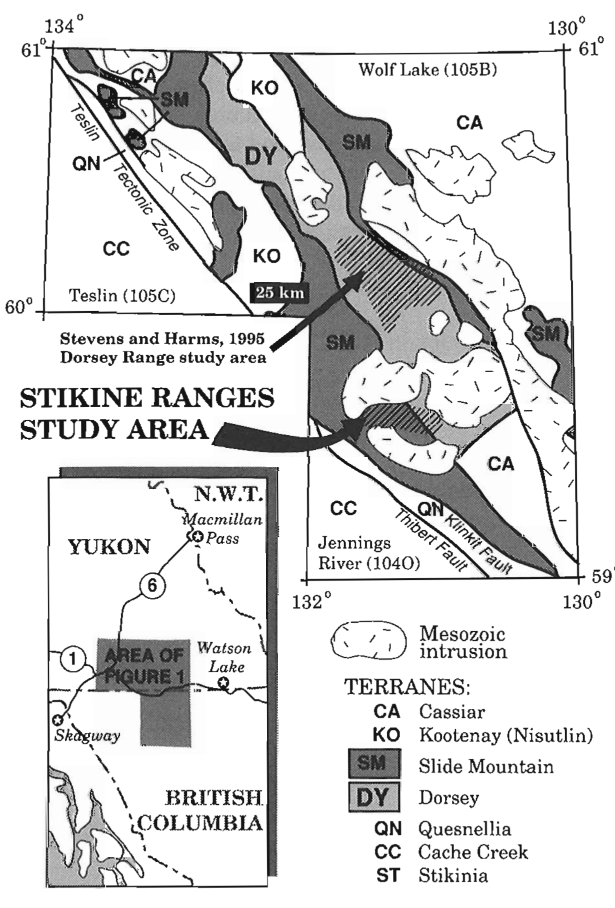


Figure 1. Location of the Stikine Ranges study area. Adapted from Wheeler and McFeely (1991).

LITHOLOGIES

Mapping at 1:25 000 scale within the Stikine Ranges in 1994 was conducted in three areas; one within the domain assigned to the Dorsey terrane and two flanking Butsih Creek within the area of Slide Mountain terrane (Fig. 2). All three were previously mapped at 1:250 000 scale by Gabrielse (1970) and were assigned by him to a single map unit (unit 11). Our work demonstrates that laterally continuous, lithologically distinctive rock units, generally several tens of metres thick, can be subdivided and mapped across each study area. However, large-scale deformation is present (described below). Our experience suggests that detailed mapping is necessary if a reliable and complete stratigraphy for the region is to be determined. Furthermore, very few units are fossiliferous; no units in the study area are, as yet, dated. Consequently, we must await laboratory analyses for microfossils and increased mapping coverage before we can properly correlate units from one study area to another. Short, provisional stratigraphic sequences that result from the work of 1994 cannot yet be built into a characteristic stratigraphic column or columns for the Stikine Ranges as a whole.

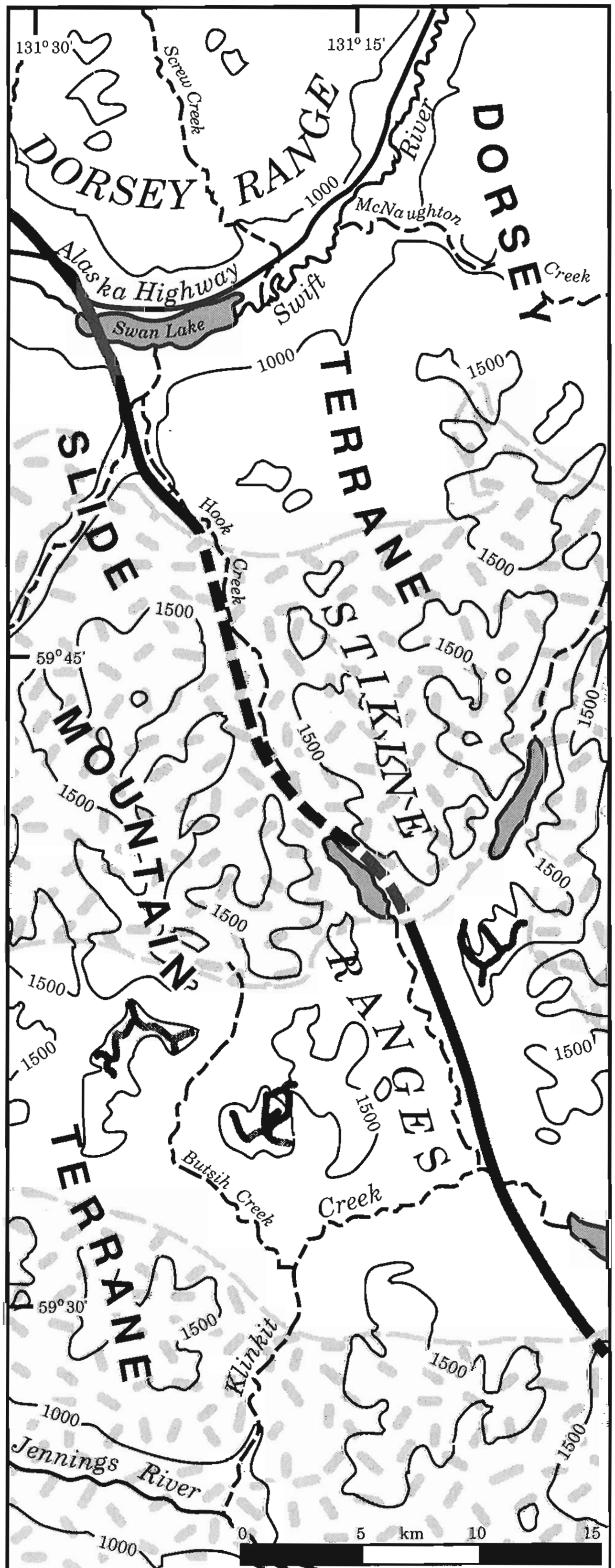
A review of the lithological units present in the study area, on the other hand, is instructive. All three map areas contain a similar suite of lithologies: clean quartz sandstone, well bedded limestone, and black shale to slate with or without associated coarser clastics. In addition, study areas within the Slide Mountain terrane contain volcaniclastics. Descriptions of these lithologies as they occur in the study areas follow.

Quartz sandstone

All three study areas contain compositionally very mature, clean quartz sandstone, which underlies as much as 30 per cent of the study areas. In all cases, the sandstone is made of well rounded and spherical quartz grains that are moderately size sorted. Sandstone may be either framework supported with quartz cement or matrix supported with a quartz silt/quartz cement matrix; carbonate cement occurs rarely. Sandstones range in colour from white to buff, light olive green or pale purple, to black, occurring most commonly as a thick (100-200 m), massive unit locally interbedded with horizons of siltstone and grey to black shale. Bedding is the only primary sedimentary structure that was observed in the sandstone.

Limestone

One or more horizons of limestone occur in each of the three study areas, but do not share the same physical characteristics. Two limestone units occur in the Dorsey terrane study area: one which is light grey, severely recrystallized, and contains relict large crinoid columnals; and another that is dark grey,



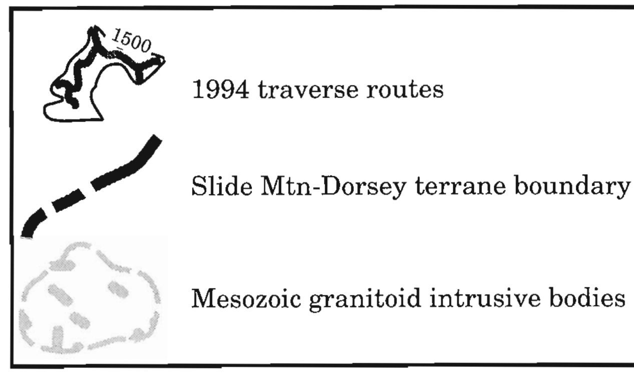


Figure 2.

Map of the Stikine Ranges. Study area traverses shown with dark stippled lines. Slide Mountain/Dorsey terrane boundary (thick line) is from Wheeler and McFeely (1991). Distribution of intrusive rocks from Gabrielse (1970). Contours and drainage taken from 1:250 000 scale topographic base NTS 104O (contour interval 500 m).

fetid in places, and unfossiliferous. Neither is more than a few tens of metres thick. In the Slide Mountain study areas, carbonate occurs in one case as four, 1-2 m thick, black limestone marker horizons within a thick black shale unit, in another case as a 10-20 m thick unit of well-bedded, interlayered unfossiliferous light grey limestone and dolomite, and in a third case as a 75 m thick unit of pervasively silicified, interlayered white, grey, dark grey and cream coloured limestones. The last includes two horizons that contain megafauna: one of abundant colonial corals and a second with (?)Pentamerid brachiopods and uncoiled Nautiloids. These units are laterally continuous but thin, together they constitute no more than approximately 10 per cent of the area mapped.

Black shale and slate

Black shale or slate units, 50 to 100 m thick, are common to all areas mapped. In the Dorsey terrane study area, black shale occurs with grey siltstone, bedded black chert, and abundant banded black porcellanite. This underlies up to 50 per cent of the area mapped. In the Slide Mountain study areas, black shale is interlayered with siltstone and granule conglomerate horizons, and with black carbonate lenses. It is slightly less abundant in the Slide Mountain study areas, underlying approximately 30 per cent of the ridges mapped.

Conglomerate

Both study areas in Slide Mountain terrane contain compositionally and texturally mature conglomerate in a horizon or horizons from one to several metres thick. Conglomerate clasts range in size from 20 cm down to a sand-sized matrix. They are predominately well rounded, although some are subangular. Clast composition is uniformly siliceous but includes chert, quartz sandstone, quartzite, and quartz phyllite parent rock types. No conglomerate was found within the confines of the Dorsey terrane study area.

Volcaniclastic rocks

Only the Slide Mountain terrane study areas contain volcaniclastic rocks. They are fine grained, massive and green, and contain visible feldspar, biotite and amphibole. A siliceous, amphibole-bearing hornfels within the Slide Mountain study area is also interpreted as volcaniclastic. Together, these rocks cover 25 to 35 per cent of the Slide Mountain study areas.

STRUCTURE

Deformation within the Stikine Ranges is dominated by reclined to recumbent, tight to isoclinal, similar folds. Folds with 0.5-1 km limb length fold map units and their contacts in both Slide Mountain study areas. Axes of these folds appear to trend east-west, with shallowly north-dipping axial planes. Furthermore, well-bedded lithologies (predominately limestones) in the Slide Mountain study areas show abundant small-scale, 1-2 m limb-length, intrafolial folds of the same style. This smaller-scale deformation does not fold unit contacts and implies some structural detachment along those horizons. Folds of both scale have considerable thinning in the overturned limb. Intrafolial folds were also recognized within the black shale, chert and porcellanite unit in the Dorsey terrane study area, however, that area is disrupted by one or more steep faults and map-scale fold noses, if they exist, could not be identified.

Without established ages for any map units, the presence or absence of layer-parallel faults in the study area is difficult to assess. Outcrop scale intrafolial folding suggests it may be a factor in the structural development of the Stikine Ranges. Along one folded contact, structurally lower units are clearly truncated at a low angle by structurally higher units, but whether this represents an unconformity in the stratigraphic section or a contractional or extensional fault cannot yet be determined.

The existence of tight to isoclinal folds throughout the area studied, accompanied as they are by thinning in overturned limbs and intrafolial fold thickening in other domains, indicates that observed unit thicknesses cannot be directly interpreted as depositional thicknesses.

INTERPRETATIONS

Several generalizations and working hypotheses arise from this study. First, there is a consistent similarity of the lithological suite, although not of detailed stratigraphic sequence, between all parts of the Stikine Ranges studied. The region is dominated by clean quartz sandstone and, where determinable through fossils, platform carbonate. Notably, no bedded chert has yet been found within the Slide Mountain domain of the ranges, and very little in the Dorsey domain. This suggests a continental margin affinity for the Stikine Ranges. It is in contrast to the oceanic character that typifies the Slide Mountain terrane throughout the Cordillera (Harms et al., 1984). The general distribution of volcanic rocks, which by definition are excluded from the Dorsey terrane (Monger et al., 1991) and provide the rationale for assigning the western Stikine Ranges to Slide Mountain terrane, is an exception to this regional lithological similarity. Extrusive and volcaniclastic rocks, however, have recently been identified in other portions of the Dorsey terrane – in the Thirtymile Range (Harms, 1992) and in the Dorsey Range (Stevens and Harms, 1995) of southern Yukon. The absence of volcanic rocks, therefore, is not an applicable discriminator of Dorsey terrane and does not justify separation of those rocks designated as Slide Mountain terrane in the Stikine Ranges from the Dorsey terrane as a whole. Finally, map-scale deformation within the area assigned to Slide Mountain terrane in the Stikine Ranges is unlike the structural stack of fault-bounded, thin and discontinuous, lithotectonic slices that characterizes the Slide Mountain terrane in the Sylvester allochthon directly to the east (Harms, 1984), to the north in Yukon (Erdmer, 1985), or at Sliding Mountain in central British Columbia (Struik and Orchard, 1985). This, too, suggests that separation of parts of the Stikine Ranges into Slide Mountain terrane is unwarranted.

ACKNOWLEDGMENTS

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Structure and terrane relationships of Cassiar and Kootenay (Yukon-Tanana) terranes, Teslin map area, southern Yukon Territory

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Gordey, S.P., 1995: Structure and terrane relationships of Cassiar and Kootenay (Yukon-Tanana) terranes, Teslin map area, southern Yukon Territory; in Current Research 1995-A; Geological Survey of Canada, p. 135-140.

Abstract: Cassiar Terrane in Teslin map area is imbricated by at least two flat thrust faults. One emplaces Proterozoic to Cambrian clastics (Ingenika Group) above Devono-Mississippian chert conglomerate (equivalent to Earn Group). The conglomerate in turn is thrust over Mississippian carbonate, quartzite, volcanics and phyllite. Meagre evidence suggests the faults are west-directed, and possibly as old as Early Jurassic.

The boundary between Kootenay and Stikine terranes is straighter than earlier mapped, making previously suggested dextral slip along this boundary kinematically more feasible.

A small window or re-entrant of Cassiar Terrane beneath Kootenay Terrane is speculated.

Résumé : Le terrane de Cassiar dans la région cartographique de Teslin est imbriqué par au moins deux failles de chevauchement planes. L'une fait en sorte que les roches détritiques du Protérozoïque-Cambrien (Groupe d'Ingenika) se trouvent au-dessus du conglomérat cherteux du Dévonien-Mississippien (équivalent du Groupe d'Earn). Le conglomérat chevauche à son tour des roches carbonatées, du quartzite, des roches volcaniques et des phyllades du Mississippien. De rares indices laissent supposer que les failles ont une orientation ouest et qu'elles pourraient remonter au Jurassique précoce.

La frontière entre les terranes de Kootenay et de Stikine est plus droite qu'indiquée dans la cartographie antérieure, ce qui rend le rejet dextre le long de cette frontière, postulé précédemment, plus plausible sur le plan cinématique.

On suppose la présence d'une petite fenêtre ou d'une indentation du terrane de Cassiar sous le terrane de Kootenay.

INTRODUCTION

Fieldwork in 1994 was the fourth and last full field season of a project to remap the geology of Teslin map area (NTS 105C; 60-61°N; 132-134°W) in southern Yukon Territory (Fig. 1). The aim of this project is to produce a new 1:250 000 scale map and reports as well as 1:50 000 scale maps of selected areas to understand the regional stratigraphic, structural and tectonic context of the area's mineral resources. The fieldwork of Mulligan in the early 1950s (Mulligan, 1963) established a broad geological framework, but little other work has been done until recently. Later reports (e.g., Jackson, 1990; Cordey et al., 1991; Gordey, 1991, 1992; Stevens, 1991, 1992, 1993; Gareau, 1992; Harms, 1992; Stevens and Erdmer, 1993; and Stevens et al., 1993) have touched on paleontology, structure, stratigraphy, geochronology and metamorphism of specific parts of the area. Gordey and Stevens (1994a,b) summarized the regional terrane and plutonic relationships, and presented a preliminary geological map at 1:250 000 scale.

The Teslin area is underlain by six terranes of contrasting sedimentary and structural histories which were juxtaposed in Jura-Cretaceous time and intruded by three main suites of post-tectonic plutonic rocks. This report focuses on newly recognized large west-vergent(?) thrust faults within Cassiar Terrane (area A, Fig. 1). Field observations of Kootenay Terrane (also referred to as Yukon-Tanana Terrane (Mortensen, 1992) or Teslin Suture Zone (Hansen et al., 1989) at two areas

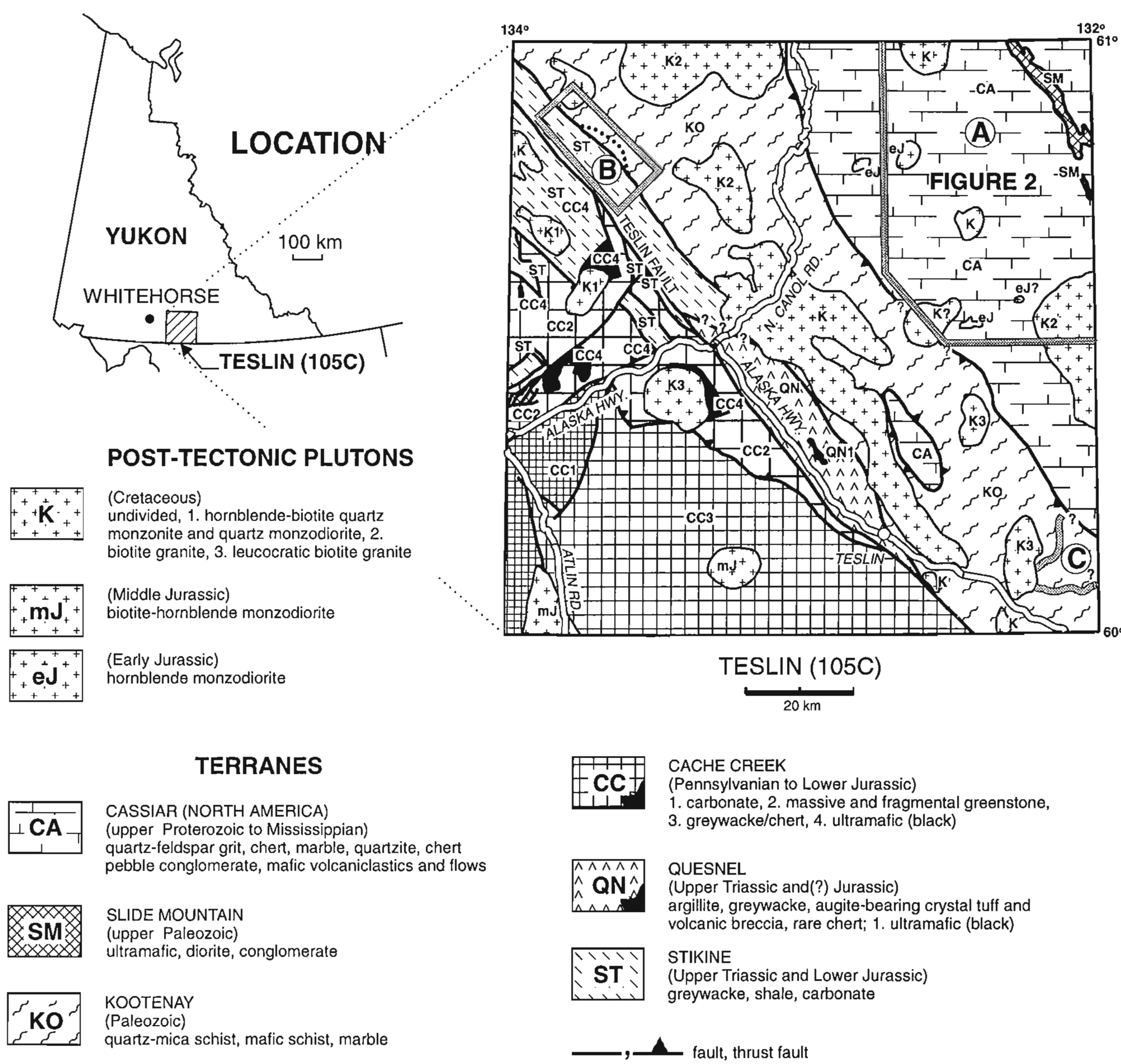

Figure 1. Location and tectonic framework of Teslin map area (105C) (from Gordey and Stevens, 1994a). Areas discussed in the text are labelled A, B and C. Area A corresponds to Figure 2.

that bear on regional tectonic relationships (areas B, C, Fig. 1) are also described briefly. No new mineral occurrences were noted. The tectonic setting of the occurrences has been outlined previously (Gordey and Stevens, 1994a,b).

CASSIAR TERRANE

Cassiar Terrane is a displaced sliver of Ancestral North America margin strata (Wheeler et al., 1991). In Teslin map area it is underlain by quartz sandstone and quartz-feldspar grit, chert, chert pebble conglomerate, andesite, quartzite, carbonate and phyllite affected by locally intense bedding-parallel mylonitization and steep normal faults. Foliation and bedding dip gently to moderately, but the nature of contacts is obscured by poor exposure. A tentative stratigraphic framework and lithological descriptions have been presented previously (Gordey, 1992).

Strata of Cassiar Terrane can be divided into three structural panels (A, B, C, Fig. 2). Lowermost strata (A, Fig. 2), of probable Devono-Mississippian age, consist of sheared chert pebble conglomerate (50 m+?) overlain (location 1, Fig. 2) by three, in part laterally equivalent, formations of early to mid-Mississippian age. Locally crinoidal, coralline and conodont-bearing carbonate and quartzite (300 m?) are facies equivalents of phyllite and black fine grained quartzite interspersed with carbonate lenses to the northwest. To the southeast intermediate to mafic volcanic flows and tuffs containing carbonate lenses form a thick succession (200 m?) within the carbonate. Panel A is overlain along flat contacts at two localities (2, 3, Fig. 2) by sheared to unsheared shale, quartz-chert wacke and chert-quartzite pebble conglomerate (300 m?) (panel B, Fig. 2). These rocks, like those at the base of panel A, are diagnostic of the Earn Group of Late Devonian to early Mississippian age which is widespread along the Ancestral North American margin. At one locality (2, Fig. 2) the base of the panel consists of massive, very fine grained siliceous rock (recrystallized chert?) that may be of Earn Group age or that could be equivalent to the older fine clastic to cherty Ordovician to Devonian Road River Group. Panel C comprises quartz-feldspar grit, shale and minor limestone (600 m+?), above which is locally preserved bedded to massive chert (200 m?) of possible early Paleozoic age. The quartz-feldspar grit is diagnostic of strata of the Proterozoic-Cambrian Ingenika Group which is widespread along the Ancestral North American margin. At a locality (4, Fig. 2) first identified by Harms (1992), coarse quartz-feldspar grit of panel C overlies panel B along a flat contact.

Although age control is poor, the strong lithological correlations with strata of known age elsewhere in Cassiar Terrane indicate older-over-younger relationships between the panels and therefore that their mutual gently dipping contacts are thrust faults. The trace of the western margin of Proterozoic strata of panel C north and south of Thirtymile Lake (Fig. 2) is tentatively indicated as a thrust fault along its trace. However, its dip is not well constrained and parts of the boundary may result from offset along post-thrusting normal faults which have down-dropped panel C relative to panel A.

Scant evidence suggests the direction of vergence of the above-mentioned faults is westerly. Along the eastern two-thirds of exposure of Proterozoic strata, from Thirtymile Lake to Fish Lake, a cleavage ranging from a fracture cleavage to a slaty cleavage or schistosity dips moderately to the east. South of Thirtymile Lake (5, Fig. 2) the intensity of cleavage begins to increase westerly within about 500 m of the probable west limit of panel C. Near this limit, the intense cleavage maintains its moderate east dip, but is also thrown into metre scale west-vergent, tight to isoclinal minor folds with moderate east-dipping axial planes. Cleaved quartzose grits are juxtaposed against poorly exposed cleaved siliceous argillite. From near this area and to the north of Thirtymile Lake, the fault is difficult to locate as it juxtaposes two successions which share nondiagnostic fine clastic components. Easterly direction of transport is dominant in Pelly Mountains to the north and northeast (Tempelman-Kluit, 1977, 1979a,b; Gordey, 1981). However, 50 km to the east in adjacent central Wolf Lake area, the dominant structures are west-directed, and include map-scale and smaller folds and a related cleavage (Murphy, 1988).

The age of thrust faulting is uncertain. The strong regional cleavage is likely of Early Jurassic or older age as strata affected by this fabric are intruded by unfoliated Early Jurassic (186 Ma; Gordey and Stevens, 1994a,b) plutons. If the thrust faults formed in response to the same deformation, then a pre-Early to Early Jurassic age is indicated. However, they could be as young as Cretaceous, the upper limit being fixed by crosscutting, mid-Cretaceous plutons. Murphy (1988) indicated an age of post-early Middle Jurassic to pre-late Early Cretaceous for west-vergent structures in central Wolf Lake area. However, the older age limit there is poorly constrained.

Ultramafics, siliceous schist, and conglomerate form poor scattered exposures in the northeast corner of the area (D, E, F, Fig. 2). Orange weathering, fine- to medium-grained, massive, variably serpentinized peridotite occurs in a southeast- trending belt, whose continuity and extent is approximated from a magnetic high on regional aeromagnetic maps (Geological Survey of Canada, 1962, 1963). Locally the rocks display a planar mylonitic fabric. The siliceous white mica bearing schist, possibly originally quartz arenite, has a strong mylonitic planar fabric locally accompanied by a strong lineation. The conglomerate consists of at least two compositional varieties. Both are massive, poorly sorted and poor to moderately indurated. Neither is penetratively deformed. In one variety (6, Fig. 2) clasts include sheared hornblende granodiorite (15%?), muscovite and biotite quartzofeldspathic schist and muscovitic quartzite (75%), dark grey limestone (5%), and dark grey chert (5%). The larger clasts are commonly better rounded than the smaller. Clast size ranges up to 18 cm across. In the second variety (7, Fig. 2; a large outcrop immediately east of the map boundary) the largest proportion of the clasts (90%) are unfoliated, medium- to coarse-grained siliceous granitoids rich in quartz (15%?) and relatively poor in fine grained mafic minerals (<5%?). Other clasts include minor grey chert, carbonate, and mica schist. The clasts are angular to subrounded and range from 1 cm to 30 cm across.

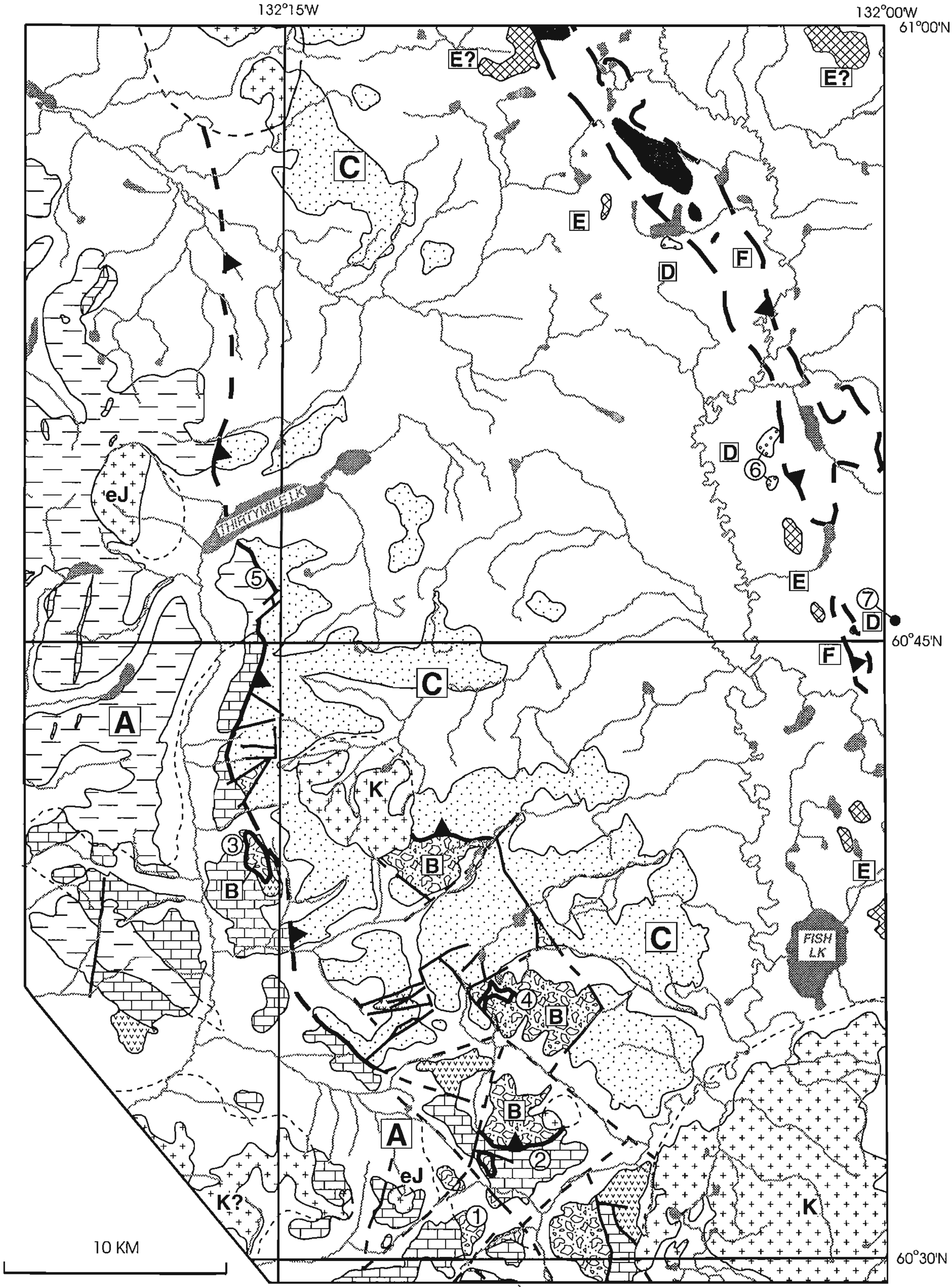

Figure 2. Map: Geology of northeast Teslin map area. Localities 1 to 7 are referenced in the text.

Relationships of the above three units to Cassiar Terrane cannot be demonstrated because of poor and scattered exposure in the Teslin area. However, they form an association that to the north and northeast in the Pelly Mountains rests structurally above Cassiar Terrane. The ultramafic rocks and siliceous schist correspond to the Slide Mountain and Kootenay (Nisutlin sub-terrane or allochthon) terranes respectively as mapped by Tempelman-Kluit (1977, 1979a,b). The conglomerate forms part of the "synorogenic clastic rocks" of Tempelman-Kluit (1979a) interpreted as recording cataclasis, collision and uplift along Teslin suture in the Late Triassic (?and Early Jurassic) and finally emplaced above autochthonous strata (Cassiar Terrane) in Early Cretaceous time.

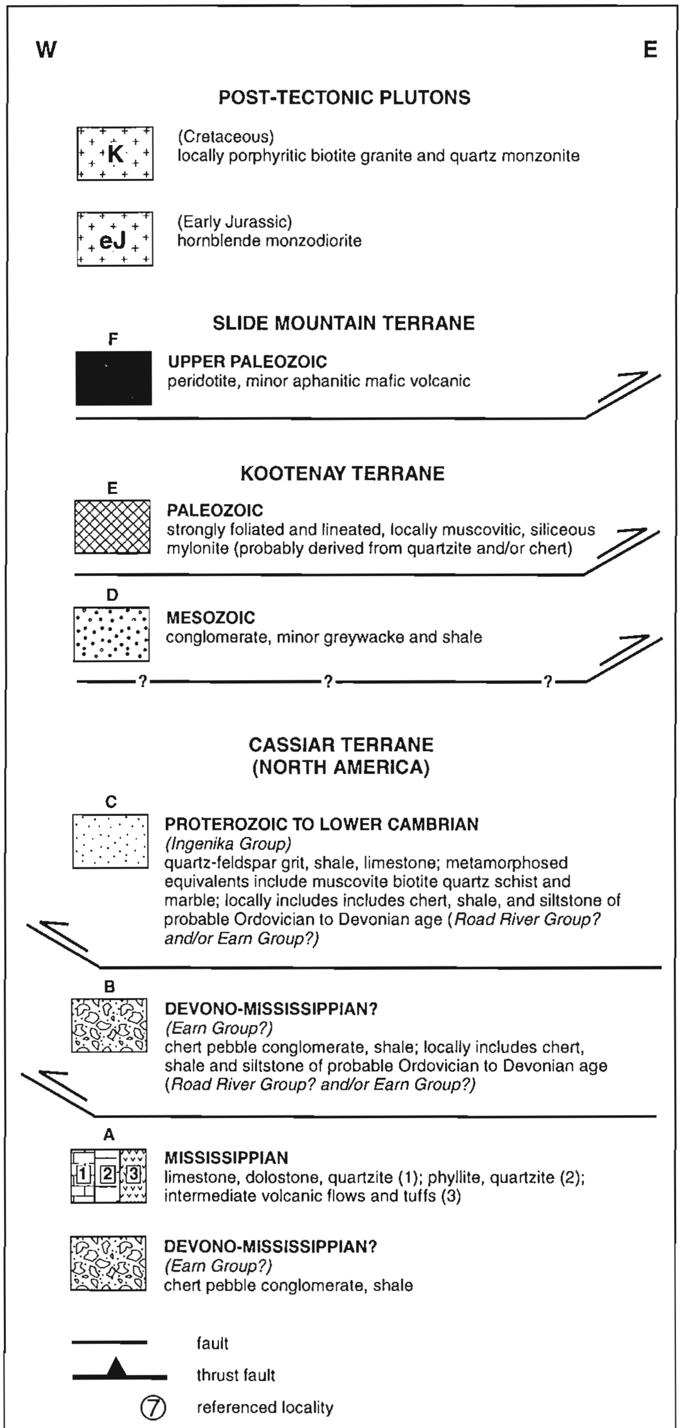

Figure 2. *Legend: Structural and stratigraphic succession in northeast Teslin map area.*

KOOTENAY (YUKON-TANANA) TERRANE

Observations on two separate areas of Kootenay (Yukon-Tanana) Terrane not examined previously bear on the tectonic framework of the region. In a relatively poorly exposed area (B, Fig. 1) Kootenay Terrane rocks including fine grained quartz mica schist and recrystallized and sheared carbonate occur further to the southwest than previously thought (Mulligan, 1963; and as interpreted by Gordey and Stevens, 1994a,b), eliminating a large bend in the Stikine - Kootenay terrane boundary (Fig. 1). With the elimination of a potential compressional fault-bend, dextral strike-slip as earlier suggested along this boundary (Gordey, 1992) is kinematically more feasible.

Observations within the second area (C, Fig. 1) pose a problem rather than resolve one. In this area a moderately southwest-dipping succession interpreted as part of Kootenay Terrane (Gordey and Stevens, 1994a,b), resembles components of Mississippian strata of Cassiar Terrane as described above. It consists of thick bands (60 m+) of medium bedded, orange and white weathering, very fine crystalline, grey to white dolostone. The orange variety typically has a lamination accentuated by differential weathering whereas the white dolostone has a colour lamination in shades of medium grey to white, some of which forms a fine wispy fabric that may be mylonitic. The dolostone contains interbands of mafic fine grained chlorite-amphibole schist from 0.5 m to 20 m thick, and at one locality white mica quartz-eye schist 2 m thick. A band of psammitic mica schist 100 m thick within the carbonate displayed 2 m scale isoclinal folds of its schistosity. Foliation in the pelitic and mafic schist is generally concordant to the large scale compositional layering. Fine grained amphibole-chlorite mafic schist is the most common lithology in scattered outcrop beneath the carbonate whereas fine grained, white, locally muscovitic quartzite about 50? m in thickness lies above it. The total structural or apparent thickness of the succession from base of the lowest carbonate to the top of the capping quartzite is at least 300 m.

South of and separated from the above succession by a gap in outcrop a kilometre wide is structurally concordant, strongly foliated biotite quartz feldspar schist. A relict, sheared granitic fabric is preserved locally as 15 per cent fine grained clots of biotite representing recrystallized mafic minerals.

The association of quartzite, carbonate and metavolcanic rocks resembles that of Cassiar Terrane described above (panel A, Fig. 2), rather than the sucrosic, locally tremolite-, garnet-, and diopside-bearing marble typical of Kootenay Terrane (Stevens, 1991). If the above-described succession represents a window or re-entrant through Kootenay Terrane into Cassiar Terrane, the former would be represented by the structurally overlying metaplutonic rocks. The extent of the window or re-entrant and the nature of the structures bounding it, whether a single thrust or a combination of thrust and normal faults, is unclear. Uncertainty in extent is compounded by poor scattered exposure and the similarity of mafic and pelitic schist components within the window to schists within Kootenay Terrane. On the basis of present evidence the window or re-entrant is unlikely to be larger than that shown in Figure 1. Its strata are of apparently higher metamorphic grade (uppermost greenschist to amphibolite facies) than those within the window 37 km to the northwest (Fig. 1) (Gordey and Stevens, 1994a) which consists of low grade, massive, intermediate to mafic volcanics with locally well preserved primary fabrics.

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Geological Survey of Canada Project 900036

New stratigraphy and structures in eastern Lansing map area, central Yukon Territory[1]

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Roots, C.F., Abbott, J.G., Cecile, M.P., Gordey, S.P., and Orchard, M.J., 1995: New stratigraphy and structures in eastern Lansing map area, central Yukon Territory; in Current Research 1995-A; Geological Survey of Canada, p. 141-147.

Abstract: The east half of Lansing map area is underlain by a succession of deep water strata ranging from Late Proterozoic to Triassic in age. New conodont data and lithological characteristics define map units which can be correlated with Hyland Group, Gull Lake Formation, Road River and Earn groups, Keno Hill quartzite and Tsichu formation, Mount Christie Formation and Jones Lake Formation.

A central belt containing middle Paleozoic to Triassic strata is bounded to the northeast and southwest by Hyland Group and lower Paleozoic rocks. Strained Hyland Group rocks in the southwest are dislocated from adjacent, relatively unstrained Hyland Group and overlapping strata along a northwest-trending shear that could be the eastern edge of the Robert Service thrust panel. A fault parallel to the axis of the central belt links the northwest-directed Tombstone Thrust in the west with the dextral Hess Fault in the east.

Résumé : La moitié est de la région cartographique de Lansing repose sur une succession de couches déposées en eau profonde, qui s'échelonnent du Protérozoïque tardif au Trias. De nouvelles données sur les conodontes et les lithologies permettent de définir des unités cartographiques qui peuvent être mises en corrélation avec le Groupe de Hyland, la Formation de Gull Lake, les groupes de Road River et d'Earn, le quartzite de Keno Hill et la formation de Tsichu, la Formation de Mount Christie et la Formation de Jones Lake.

Une zone centrale renfermant des couches du Paléozoïque moyen au Trias est limitée au nord-est et au sud-ouest par des roches du Groupe de Hyland et du Paléozoïque inférieur. Au sud-ouest, des roches déformées du Groupe de Hyland sont disloquées par rapport au Groupe de Hyland adjacent, relativement peu déformé, et aux couches chevauchantes le long d'un cisaillement à direction nord-ouest qui pourrait être le bord est du panneau de charriage de Robert Service. Une faille parallèle à l'axe de la zone centrale relie la nappe de charriage de Tombstone à orientation nord-ouest, dans l'ouest, avec la faille dextre de Hess, dans l'est.

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INTRODUCTION

Lansing map area (105N) is one of the last parts of the Selwyn Basin to be systematically mapped and explored for stratiform base metal deposits. The region is remote from roads and 120 km from Mayo and Ross River (Fig. 1). Widely scattered, isolated bedrock exposures cannot be mapped efficiently employing the normal method of small flycamps moved periodically by a helicopter based in the distant community. Therefore the east half of Lansing map area was mapped by the four first authors using a contract helicopter and small base camp within the field area. Cumulatively the mapping team has 30 field seasons of experience with Selwyn Basin rocks, including completion of adjacent map areas to the east and south (e.g., Gordey and Irwin, 1987; Cecile and Abbott, 1992; Gordey and Anderson, 1993).

Lansing map area not only contains positively identified Late Proterozoic to middle Paleozoic rocks (Blusson, 1974; Gabrielse et al., 1980; Wheeler and McFeely, 1991) but newly discovered upper Paleozoic to Triassic strata. Discovery of these units is based on conodonts recovered from limestone samples collected in 1993 and earlier (Table 1) as well as this summer's mapping. This report presents the new map (Fig. 2), and outlines the units and major structures discovered during mapping. It updates the preliminary work of Roots and Brent (1994a,b).

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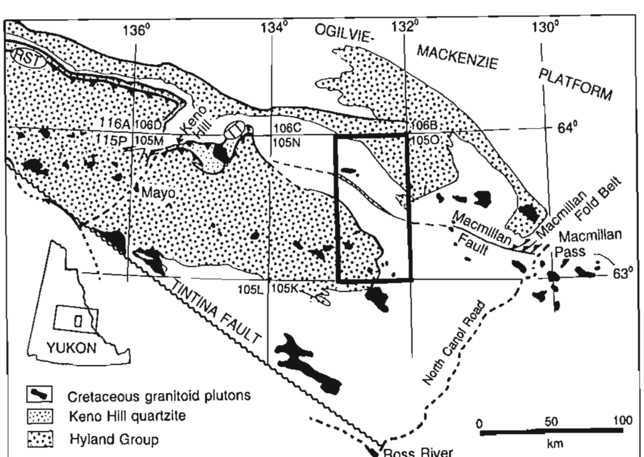

Figure 1. Portion of central Yukon with the east half of Lansing map area (105N) outlined. Two regional thrust faults, TT = Tombstone Thrust and RST = Robert Service Thrust are separated by Devonian and Carboniferous rocks, including the Keno Hill quartzite. Regional geological reports are available for the map areas as follows: 116A and 106D: Green (1972); 115P: Bostock (1964) and Murphy et al. (1993); 105M: Roots and Murphy (1992b); 105L: Campbell (1967); 105J and K: Gordey and Irwin (1987); 105O: Cecile and Abbott (1992); and 106B and C: Blusson (1974).

GEOLOGY OF LANSING MAP AREA, EAST HALF

Selwyn Basin comprises a 250 km wide belt of Late Proterozoic through Triassic sediments (Abbott et al., 1986) deposited outboard of the stable continental margin (Ogilvie-Mackenzie Platform) of ancestral North America. These strata have undergone Jura-Cretaceous contraction by folds and thrust faults and are of lower greenschist grade.

In eastern Lansing map area Paleozoic and Triassic strata form an elongate northwest-trending belt bounded by Hyland Group to the northeast and southwest. Penetrative deformation has left only a few areas where macrofossils are preserved. The stratigraphic units are locally dated by microfossils for which Table 1 is a preliminary list. No stratigraphic sequences were measured because units are incompletely exposed and thickness is tectonically modified.

The following comments are preliminary observations on the stratigraphic units of the east half of Lansing map area.

The oldest exposed rocks belong to the Hyland Group (Gordey and Anderson, 1993) which is composed of a lower formation dominated by sandstone-grit and topped by limestone (Yusezyu) and a shaly upper formation including a distinctive maroon argillite (Narchilla). In northeast Lansing map area these formations are extensive and continuous into northern Niddery Lake map area, where they are similar to strata seen by Cecile and Abbott (1992). Maroon shale is also locally present below the limestone. Southeast of Fairweather Lake, and extending westward across southern Lansing map area quartz-mica schist, psammite and dark-coloured phyllite are Yusezyu Formation of the Hyland Group. These strained rocks are described in Roots and Brent (1994a).

Disconformably overlying the Hyland Group is a thick recessive olive and brown silty argillite. This unit includes strata described under Paleozoic units 1 and 2 by Roots and Brent (1994a). Protoconodonts were recovered from this unit; they indicate Cambro-Ordovician age and lithologically this unit resembles the Gull Lake Formation (Gordey and Anderson, 1993). In the northeast part of Lansing map area this formation includes dark shale with thin calcareous sandstone interbeds, as well as orange weathering quartzite, with rare mafic volcanic flows or sills and conglomerate. The Rabbitkettle Formation, a prominent limestone which overlies an unconformity across southeastern Selwyn Basin, has not been recognized in Lansing map area.

The next youngest unit, which apparently disconformably overlies the Gull Lake Formation is the Road River Group. It includes two formations. The Ordovician unit is dominated by white weathering thick-bedded chert, whose fresh surface colour is typically grey but locally green, blue and black. At the graptolite locality north of the Hess River (no. 7 in Fig. 2 and Table 1) this chert appears to be 100 m thick, but 10 km to the north the chert is structurally interleaved with the Gull Lake Formation so that the two units cannot be mapped separately. Ordovician strata are not present west of the Plata property and were presumably removed before deposition of the Earn Group.

The upper unit of the Road River Group is characterized by orange or light brown weathering green argillite which locally includes bright orange weathering carbonate. It matches the Steel Formation (Silurian; Gordey and Anderson, 1993).

The Earn Group overlies the Steel Formation in the central part of eastern Lansing map area, but near the Plata property and in the south it disconformably overlies Gull Lake and Hyland Group rocks. Characteristic gunsteel-blue weathering carbonaceous siliceous siltstone, and dark chert-pebble conglomerate define this unit, but brown shale and sandstone are present. East of Mount Osgoode the conglomerate contains sandstone boulders and directly overlies Hyland Group, similar to adjacent northeastern Tay River map area (105K; Gordey and Anderson, 1993, p. 66). Elsewhere chert clasts of smaller size predominate. No complete undeformed sections are known and separation of the Earn Group into formations is not possible. Mafic flows and pods of white weathering limestone are present within the Earn Group east of Fairweather Lake.

The Earn Group is overlain by a succession of dark siltstone containing detrital mica, fine sandstone interbeds and rare sandy limestone. One collection of Viséan conodonts (no. 4 in Fig. 2 and Table 1) has been recovered, indicating that it is correlative with the lower part of the Tsichu formation (informal unit of Gordey and Anderson, 1993) which is extensively exposed in southern Niddery Lake map area (Cecile and Abbott, 1992). Furthermore, a persistent marker band of dark grey or black quartzite, varying to brown weathering quartz sandstone and finely recrystallized rock resembling black chert, has been traced over 40 km from Murray Creek to Fido Creek. This lithology is correlated with the Keno Hill quartzite, also of Lower Carboniferous age (Mortensen and Thompson, 1990). The presence of this unit, heretofore unrecognized in Lansing map area (Roots and Brent, 1994a) has structural implications discussed below.

Permian strata are identified in two places. Orange weathering green argillite and siliceous shale northeast of Fairweather Lake is correlated with the Mount Christie Formation (Gordey and Anderson, 1993). Another succession, east and south of Fairweather Lake, includes rust-brown weathering jet black siliceous shale and silty shale with thin beds of calcareous sandstone. Both successions contain locally abundant barite nodules up to 15 cm diameter. In the Wilson Range and west of the Plata property a third succession includes purple, brown and grey siliceous siltstone, locally laminated and limy, dark brown weathering black siltstone and interbedded green and grey chert. A thin black limestone bed contains Permian conodonts.

Table 1. Fossil determinations and age assignments from GSC fossil reports MJO-1994-25, except 1 (MJO 1994-27), 6 (D-3-BSN-1994) and 7 (O 8-BSN-1994).

No.	GSC Loc.	Age	Faunal List	Location	UTM coordinates
1.	C-108166	Late Carnian, Late Triassic	*Metapolygnathus* ex. gr. *nodosus* Hayashi 1968	63°36'10"N;132°00'00"W	9 351219E 7056045N
2.	C-300480	Artinskian?, Early Permian	*Neogondolella* sp. cf. *N. idahoensis* Youngquist, Hawley and Miller 1951	63°35'03"N;132°04'52"W	8 644850E 7053800N
3.	C-301303	Artinskian, Early Permian	*Neogondolella* sp. cf. *N. intermedia* Igo 1981, *Sweetognathus* sp.	63°33'31"N;132°16'23"W	8 635450E 7050550N
4.	C-301307	Visean-Serpukovian, Early Carboniferous	*Vogelgnathus* sp. cf. *V. campbelli* Rexroad 1957, *Rhachistognathus*? sp.	63°39'10"N;132°25'32"W	8 627450E 7060700N
5.	C-301309	Middle Fammenian, Late Devonian	*Palmatolepis glabra pectinata* Ziegler 1962, *Palmatolepis marginifera* Helms 1959	63°42'57"N;132°17'43"W	8 633600E 7068000N
6.	C-203018	Pragian, Early Devonian	*Monograptus yukonensis* Jackson and Lenz 1962	63°39'35"N;132°01'30"W	8 647247E 7062326N
7.	C-203020	Ashgill?, prob. Late Ordovician	*Arachniograptus* sp., ?*Climacograptus* sp., *Dicellograptus* sp., *Orthograptus* sp.	63°28'10"N;132°11'30"W	8 639932E 7040766N
8.	C-300496	Llanvirn-Caradoc, Ordovician	*Belodella*? sp., *Drepanoistodus* sp., *Periodon* sp., *Eoplacognathus*? sp., *Protopanderodus* sp., *Spinodus* sp., *Walliserodus*? sp	63°30'52"N;132°28'20"W	8 625750E 7045200N
9.	C-300497	Llanvirn-Caradoc, Ordovician	*Periodon* sp., *Protopanderodus* sp., *Pygodus* sp. cf. *P. serra* Hadding 1913	63°31'07"N;132°24'16"W	8 629100E 7045800N
10.	C-300498	Arenig-Caradoc, Ordovician	*Periodon* sp.	63°30'53"N;132°23'56"W	8 629400E 7045375N
11.	C-300494	Arenig, Early Ordovician	*Drepanoistodus* sp., *Oepikodus*? sp., *Walliserodus* sp.	63°30'17"N;132°23'05"W	8 630150E 7044300N
12.	C-300473	Cambrian-Ordovician	Protoconodont?	63°38'35"N;132°02'45"W	8 646300E 7060450N

LEGEND

Geological contact (defined, assumed)

Fault, thrust fault (teeth on upper plate)

Anticline, overturned syncline

Strike and dip of bedding, foliation

① Fossil locality; see Table 1

Km

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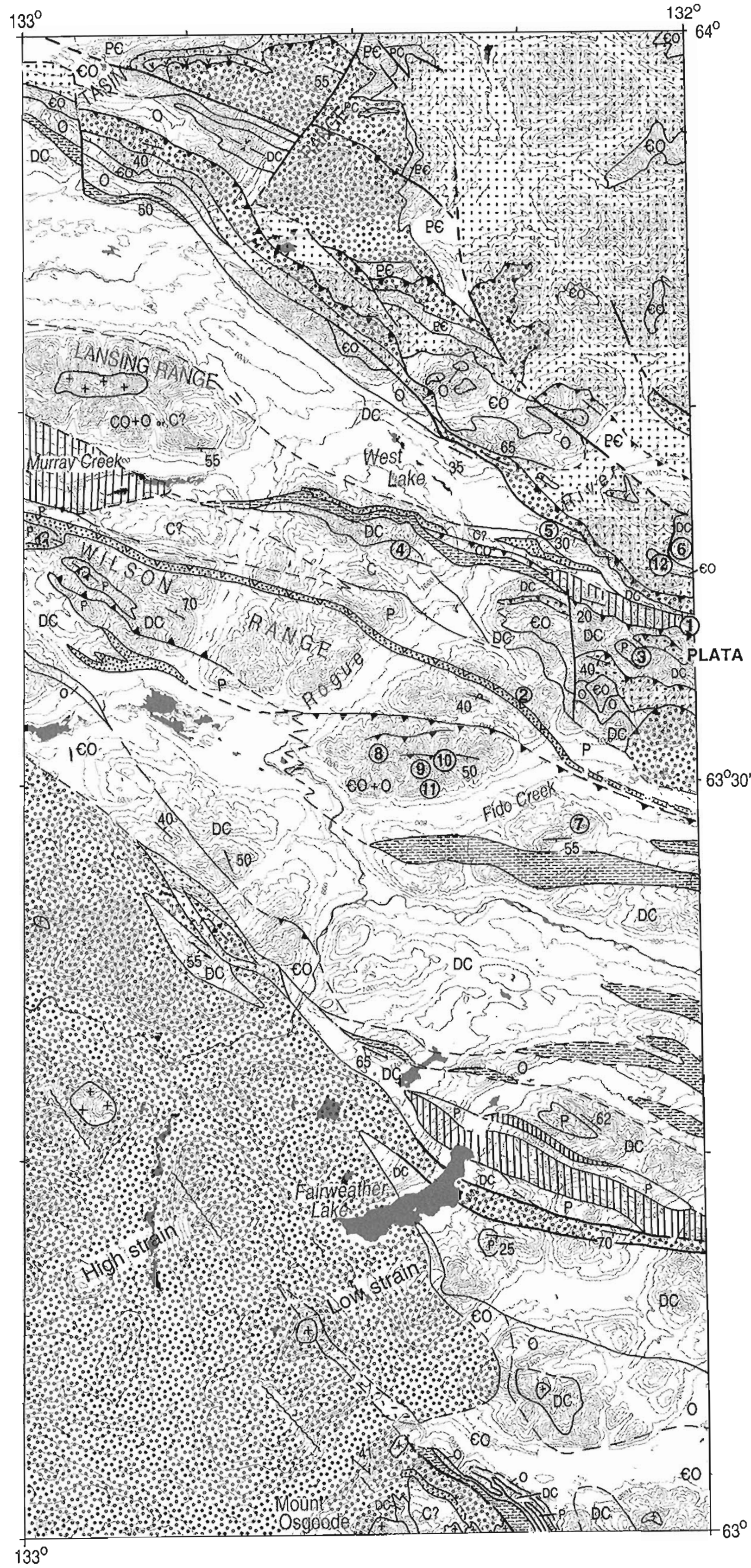

Figure 2.

East half of Lansing map area showing generalized stratigraphic units.

Triassic strata were recognized at three localities in Lansing map area. The rocks are brown weathering, ripple cross-laminated fine sandstone and shale of the Jones Lake Formation (Gordey and Anderson, 1993). Generally this unit is recessive and its extent southeast of the Lansing Range is unknown. North of the Plata property it appears to be an interbedded succession of sandstone and shale up to 600 m thick but is truncated by a thrust.

The age of the large area of recessive shale and siltstone north of the Permian and Triassic units in the broad valleys surrounding the Lansing Range is uncertain. The rocks weather brown and consist of green to dark grey shale and slate, locally wispy laminated and containing thin sandstone interbeds and lenses of white weathering grey chert. In many places the chert resembles that of the Ordovician unit (suggesting the surrounding shale is Gull Lake Formation), but similar chert is known in the Permian successions. At least some of the rocks are upper Paleozoic as indicated by a Viséan conodont in the dark grey platy siltstone succession near the east end of the Wilson Range. The assignment of these rocks awaits the processing of additional microfossil samples and inferences based upon detailed structural interpretation.

STRUCTURE

The eastern Lansing map area contains mostly southwest- and south-dipping rock units which are repeated by thrusts and open folds. Most thrust faults dip south and indicate northward transport, with vertical adjustments on north and northeast trending faults. In the northeast part of the area a single, relatively flat-lying panel of Yusezyu at least 25 km^2 overlies maroon argillite on a presumed shallow thrust.

The southwest quadrant of the area contains extensive Hyland Group with younger strata along its northeast contact. This contact, previously depicted as the southeastern end of

Figure 2. Legend

CRETACEOUS

+ + — White weathering, fine grained hornblende ± biotite quartz monzonite and granite

TRIASSIC

Jones Lake Formation: Brown and grey weathering fine grained calcareous sandstone and siltstone, ripple cross laminated; with interbedded brown shale; thin bedded blue-grey platy limestone

PERMIAN

P — **Mount Christie Formation:** Orange-brown and pink weathering green argillite, chert and silicious siltstone, laminated and burrowed, with barite nodules, minor limy sandstone, brown weathering black siliceous siltstone with deep orange weathering sandstone and dolostone

LOWER CARBONIFEROUS

C — Orange-brown and dark grey weathering weakly calcareous bioturbated micaceous black siltstone and fine brown sandstone, minor beds of dark grey sandy limestone

Grey-brown weathering fine grained dark grey quartzite with thin black shale and silicious argillite interbeds

DEVONIAN to LOWER CARBONIFEROUS

Earn Group

DC — Dark blue grey weathering silicious carbonaceous siltstone, thin bedded black chert, conglomerate and grit with chert fragments, brown weathering dull black mudstone; minor white weathering limestone and amygdaloidal mafic flows

SILURIAN

Road River Group

Steel Formation: Orange brown weathering green silicious argillite; orange weathering, thin bedded beige dolostone

ORDOVICIAN (to LOWER DEVONIAN?)

O — White weathering, thick bedded grey, green and blue chert; grey and brown shale

UPPER CAMBRIAN to LOWER ORDOVICIAN

ЄO — **Gull Lake Formation:** Brown and olive laminated mudstone and siltstone; dark brown shale with minor sandstone and grit, white weathering limestone. Locally mafic flows and tuffs and basal green grey laminated sandstone and nodular silty limestone

UPPER PROTEROZOIC to MIDDLE CAMBRIAN

Hyland Group

Narchilla Formation: Maroon, brown, black and green argillite and siltstone; locally single thick beds of grey weathering laminated quartz sandstone, mafic flows

PЄ — Light brown weathering grit, sandstone, thin-bedded sandstone interbedded with grey and brown shale in fining upward succession; thin sandy limestone and white weathering, thick bedded grey white limestone

Yusezyu Formation: Brown weathering calcareous sandstone, quartz grit with interbedded dark grey shale and siltstone, locally graded; includes single thick bed of yellow-brown limestone the Robert Service Thrust (Wheeler and McFeely, 1991; Roots and Brent, 1994a) lacks the structural relationship of a thrust. Instead the northeast contact of the Hyland Group with younger sedimentary rocks is an irregular trace that appears the result of folds intersecting with topography. Along the boundary maroon argillite of the upper Hyland Group is locally present, indicating that the boundary is at the top of the Hyland Group. Also unlike a thrust is a strip up to 2 km wide of Hyland Group quartz sandstone with limestone pods that trends eastward and is in part faulted between Devonian and Permian rocks.

The Hyland Group in the southwestern corner of the area consists of strained Yusezyu grit, schist and phyllite with a northeast-dipping foliation and gently northwest plunging mineral stretching lineation. The northeast limit of the strained rocks extends northwest from Mount Osgoode for over 50 km. It is a zone containing mylonitic shears and imbricated competent beds, as well as the trend of four granitic plugs and a chloritic mafic dyke. On the northeast side of the shear zone the degree of internal strain is much less; bedding and detrital grains, as well as worm burrows and colour laminations are visible. These observations suggest that Hyland Group in the southwest part of the area dips northeast and underlies much or all of the Lansing map area. We suspect that the highly strained rocks in the southwest are the uplifted exposure of a deeper structural level. One possibility is that the Robert Service Thrust panel moved northwestward on the underlying Tombstone Thrust (documented in the Keno Hill area, Abbott, 1990; Roots and Murphy, 1992a) and the shear zone is a dextral tear fault with a component of vertical displacement. An alternative possibility is that the highly strained rocks represent exposure of strain related to movement of the underlying Tombstone Thrust reaching into the hanging wall of the Robert Service Thrust.

The Tombstone Thrust (Abbott, 1990; Mortensen and Thompson, 1990; Wheeler and McFeely, 1991) has been traced eastward over 250 km to the northeast corner of Mayo map area. Whether it extends eastward into Lansing map area is uncertain (Roots and Brent, 1994a). If present this thrust should lie north of the Keno Hill quartzite (i.e. carry this unit in its hanging wall), which occurs in a thin belt from Murray Creek to Fido Creek (Fig. 2). Although the northern contact of the quartzite unit is poorly exposed and the age of the adjacent rocks largely unknown, we have observed changes in structural orientation along this strike length.

In the northern Wilson Range (west limit of mapping) the Keno Hill quartzite forms a south-verging syncline cored by probable Permian rocks, and has a steeply north-dipping limb with gentle north-dipping cleavage which indicates an overturned fold. Southeast of Rogue River the quartzite dips moderately south with conformable strata of known Permian age structurally beneath it on the north side, and Devonian strata above the quartzite to the south. Thus the quartzite is likely overturned. A thrust itself overturned along strike to the west may underlie the overturned strata. The relatively linear outcrop trace and large area over which the unit is overturned suggest a regional fault, probably the Tombstone Thrust.

In adjacent Niddery Lake map area (105O) the Hess and Macmillan faults appear to be dextral transpression faults related to the Macmillan fold and thrust belt (Abbott and Turner, 1990). The Hess Fault may extend through Lansing map area which connects the southward displacement in the Macmillan Fold belt with the northwestward translation of the Tombstone thrust panel in the Keno Hill area.

CONCLUSIONS

Eastern Lansing map area constitutes a transect across a little known part of the Selwyn Basin at the northeast edge of the Robert Service Thrust panel. It contains upper Paleozoic and Triassic strata on both sides of a break that could be the extension of the Tombstone Thrust. In the southwestern corner the mapping has revealed a strain gradient within Hyland Group rocks. Completion of mapping of the Lansing region will elucidate links between major structures traced eastward from the Mayo area, and westward from the Macmillan Pass district.

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Stratigraphy and structure of the Driftpile stratiform Ba-Zn-Pb deposit, Gataga area, northeastern British Columbia

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Abstract: The Driftpile deposit is hosted by fine-grained carbonaceous black siliciclastic rocks of the Middle to Late Devonian Lower Earn Group, in a series of imbricate northeastward-verging thrust panels.

Mineral assemblages consist of two distinct facies occupying two different stratigraphic positions – a localized, lower sulphide-carbonate facies and a laterally extensive upper barite (±sulphides) facies. The footwall is massive mudstone or siliceous to cherty carbonaceous argillite interbedded with radiolarian-rich chert and locally with cryptic pyrite laminated mudstone. Hanging wall rocks are turbidite lithofacies overlain by concretionary shale and nodular mudstone.

This stratigraphic sequence with two cycles of mineralization, each preceded by a period of anoxic, starved sedimentation and succeeded by an influx of turbidites, represents an interplay of mineralizing and tectonic events, with discrete episodes of faulting triggered the release of metalliferous basinal brines into anoxic bottom waters and then the sudden and rapid influx of turbidites.

Résumé: Le gîte de Driftpile est encaissé dans des roches silicoclastiques carbonatées à grain fin du Groupe d'Earn inférieur (Dévonien moyen – supérieur). La structure comporte une série de panneaux de charriage à vergence nord-est.

La minéralisation comporte deux faciès distincts occupant deux positions stratigraphiques différentes – un faciès inférieur à sulfure-carbonate, à distribution limitée, et un faciès supérieur à barytine (±sulfures) à distribution latérale étendue. Le plancher comporte un mudstone massif ou une argilite carbonatée ou cherteuse à interlits de chert à radiolaires et de mudstone pyriteux. Le toit comporte un lithofaciès à turbidite surmonté d'un shale riche en concrétions et d'un mudstone nodulaire.

Cette séquence stratigraphique à deux cycles de minéralisation, chacun précédé d'une période anoxique pauvre en sédiments et succédé d'un influx de turbidites, représente l'effet réciproque d'événements minéralisateurs et tectoniques durant lesquels des épisodes tectoniques relâchent des fluides de bassin riches en métaux, puis amènent un apport soudain et rapide de turbidites.

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INTRODUCTION

One of the world's largest concentrations of sediment-hosted stratiform lead-zinc-silver-barite and barite deposits is located in the Kechika Trough of northeastern British Columbia. One of these deposits, the Driftpile deposit, lies within the Gataga area of the Kechika Trough (Fig. 1). It consists of stratiform Ba-Zn-Pb minerals hosted by fine-grained carbonaceous and siliciclastic marine strata of the Middle to Late Devonian Lower Earn Group.

This paper summarizes preliminary results of a study of the stratigraphy and mineralization of the Driftpile deposit. The main objective of the project is to establish a detailed stratigraphic section of the deposit, incorporating biostratigraphy based on conodonts. Because of the chaotic sedimentation, metamorphic overprint, and high degree of tectonic deformation in the deposit area, it is difficult to relate local stratigraphy to regional stratigraphy and to tightly constrain stratigraphic position of the stratiform barite (± sulphides) and sulphide-carbonate mineralized units without fossil control.

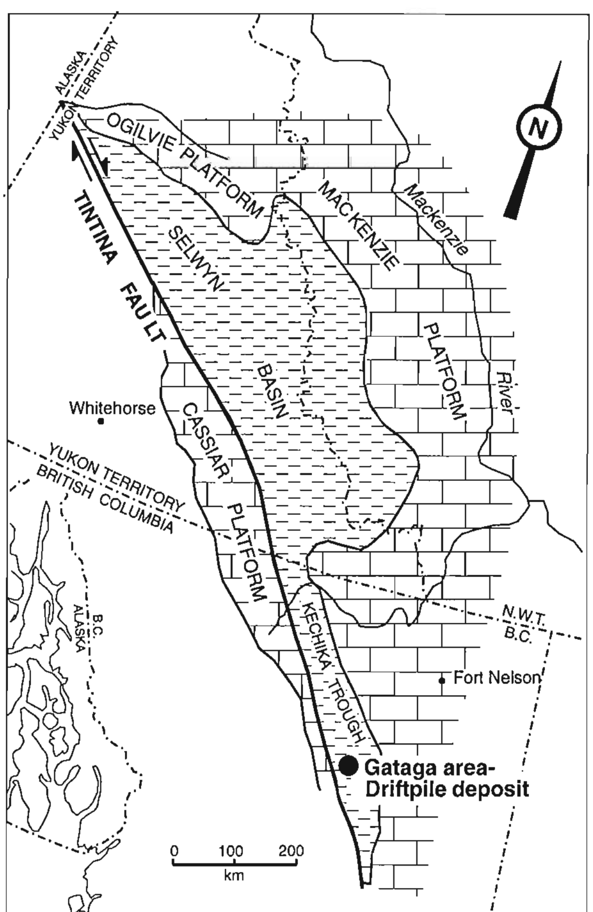

Figure 1. Location of the Gataga area and the Driftpile deposit within the Kechika Trough, northeastern British Columbia (modified from Carne and Cathro, 1982).

This study is part of a two-year multidisciplinary program in the Gataga area that includes regional mapping and geochemical studies (see Nelson et al., in press; Ferri et al., in press; Lett and Jackaman, in press) carried out in cooperation with the Geological Survey Branch of the British Columbia Ministry of Energy, Mines, and Petroleum Resources.

METHODOLOGY

A one-month field season (1994) was undertaken to examine and sample cores from 25 drillholes. Geological mapping along Driftpile Creek was done at a scale of 1:2000.

Over 80 samples of carbonaceous shale and siltstone, and chert were collected and submitted to S.E.B. Irwin and M.J. Orchard (GSC) for biostratigraphic and taxonomic studies.

TECTONO-STRATIGRAPHIC SETTING

The Gataga area lies within the Muskwa Ranges of the northern Rocky Mountains, directly east of the Northern Rocky Mountain Trench-Kechika strike slip fault system (Gabrielse, 1985; Gabrielse et al., 1992). It includes part of the Kechika Trough, which forms the southern extension of the Lower to Mid-Paleozoic Selwyn Basin (Fig. 1). The Kechika Trough has been interpreted variously as a two-sided trough (McClay et al., 1989; MacIntyre, 1992) or as a west-facing open continental slope (Gabrielse, 1985, pers. comm., 1994) during early Paleozoic time.

Several episodes of sedimentary exhalative barite and barite-sulphide mineralization occurred in the Kechika Trough: Middle Ordovician, Early Silurian, and Late Devonian. Clastic rocks of Late Devonian age host the most economically important deposits, such as Driftpile and Cirque.

In the Gataga area, a thick succession of basinal facies clastic and subordinate Late Proterozoic to Mississippian carbonate rocks are exposed within a complex northwest-striking fold-thrust belt (McClay et al., 1988; MacIntyre, 1992). Late Proterozoic rocks include highly deformed phyllite, sandstone, and quartz-pebble conglomerate that outcrop only in highly deformed and cleaved folds cores (Taylor and Stott, 1973; McClay et al., 1988). A thick sequence of Lower to Middle Cambrian shallow water clastic and carbonate rocks (Fritz, 1979; McClay et al., 1987) is overlain conformably by Cambro-Ordovician phyllite and silty limestone of the Kechika Group. Ordovician to Lower Devonian calcareous black shale, mudstone, chert, and limestone of the Road River Group rest conformably above the Kechika Group. Siliciclastic rocks (chert-pebble conglomerate, quartzose wacke, silty shale, siltstone, blue grey weathering shale, siliceous carbonaceous shale, and argillite, and minor pelagic limestone) of the Middle Devonian to Mississippian Earn Group overlie rocks of the Road River Group (MacIntyre, 1992). They represent a change from passive margin sedimentation in Early Cambrian to Middle Devonian time, to a period of starved-basin sedimentation followed by marine transgression during which westerly derived clastics were shed into fault-controlled troughs and basins (Gabrielse, 1976; Gordey, 1988). Earn Group lithologies and their tectonic setting represent a tensional regime that resulted either from continental rifting and strike-slip faulting (Abbott et al., 1986), or from flexural extension during foreland deformation related to the Antler Orogeny (Smith et al., 1993). The Earn Group has been subdivided into three informal formations, Gunsteel, Akie, and Warneford, by Jefferson et al. (1983), Pigage (1986), and MacIntyre (1992). The Middle to Late Devonian Gunsteel formation is equivalent to the Lower Earn Group, which hosts important sedimentary exhalative barite-sulphide deposits such as Driftpile and Cirque in the Gataga area, and Tom and Jason in the Macmillan Pass area of eastern Yukon. The Lower Earn Group is typically carbonaceous and siliceous, and includes cherty argillite, radiolarian chert, siliceous carbonaceous shale, carbonaceous chert, and fine grained turbidite sandstone, siltstone, and mudstone.

Jura-Cretaceous crustal shortening of the ancestral North America continental margin deformed Paleozoic to Mississippian strata of the Gataga area into a complex, mainly northeast-verging fold and thrust belt. McClay et al. (1989) recognize five major thrust panels, defined by rocks of different stratigraphic levels and, in some cases, contrasting facies. Within each major panel, strata are folded and stacked in duplexes. Thrust faults bounding the major panels may be reactivated Paleozoic growth faults. Late Devonian sediment-hosted stratiform barite (±sulphides) and sulphide-carbonate mineralized units are found in the central thrust panel. A stack of upper Road River and Earn Group horsts is exposed within the central thrust panel, bounded by decollements at the base of the upper Road River Group and within the Earn Group. Strata that host the Driftpile deposit are truncated to the west by the major thrust fault that bounds this panel, informally termed the Mt. Waldemar fault.

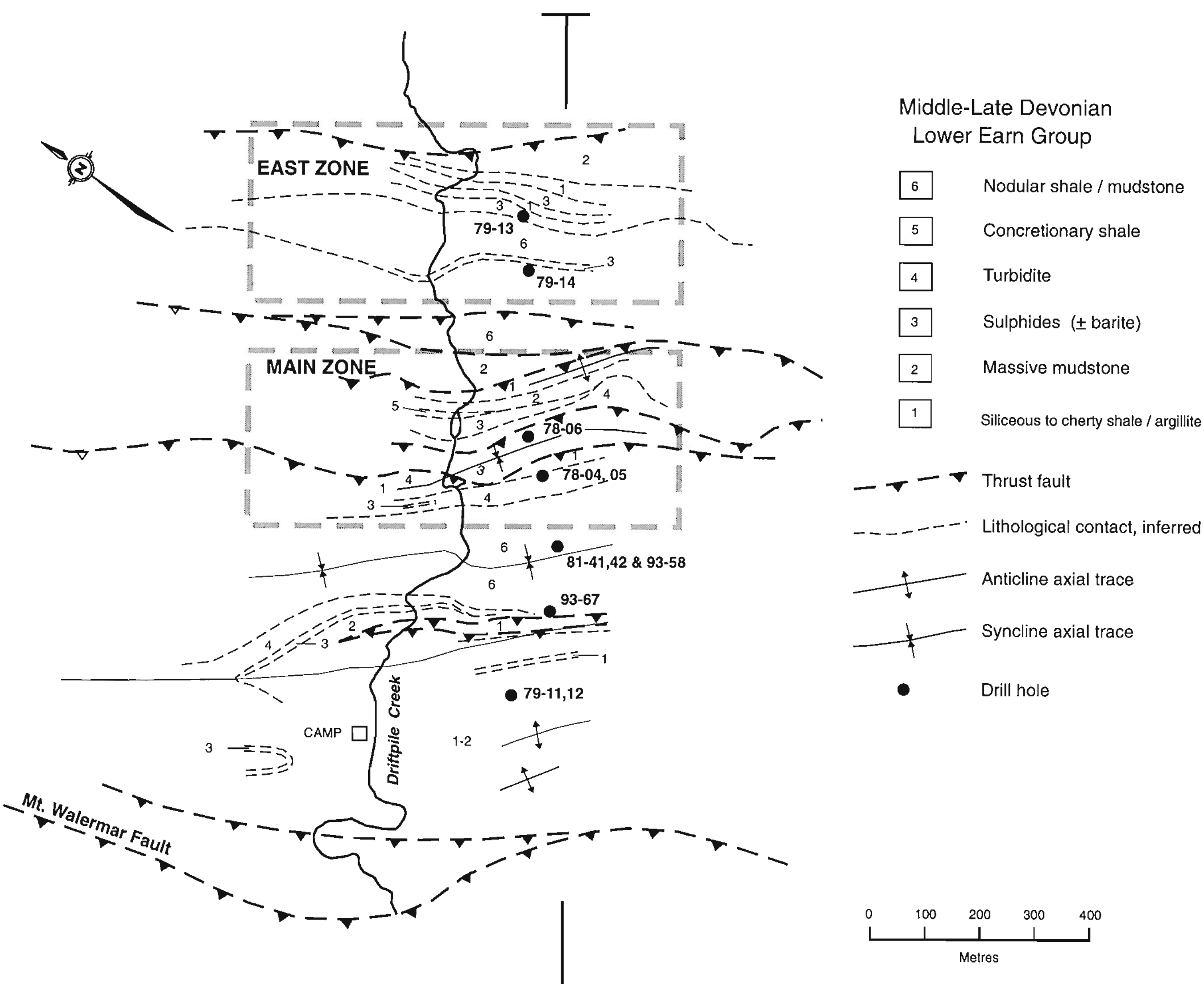

Figure 2. *Geology of the Driftpile Creek area and location of drillholes.*

GEOLOGY OF THE DRIFTPILE DEPOSIT

Stratigraphy

Several occurrences of bedded barite (±sulphides) and one of sulphide-carbonate minerals, all folded and thrusted, occur within carbonaceous and siliceous fine-grained siliciclastic rocks of the Lower Earn Group. Conodont faunas collected within host rocks and mineralized units range from the Middle *Crepida* zone to the Upper *Marginifera* zone, indicating a middle Famennian age (Irwin and Orchard, 1989; Irwin, 1990).

The stratigraphy presented below is characteristic of the "Main Zone" and the "East Zone" of the Driftpile Creek area (Fig. 2, 3).

Massive homogeneous in part siliceous black shale and mudstone (unit 2) locally form the footwall rocks to the sulphide-carbonate assemblage (unit 3A). Unit 2 is a shale/mudstone unit with rare laminations and no concretions, nodules, or distinctive features. It can be siliceous or non siliceous, and occasionally carbonaceous. When it is absent in the immediate footwall to unit 3A, it is replaced by unit 1, a well laminated siliceous to cherty carbonaceous shale and argillite interbedded with centimetre-sized radiolarian-rich chert bands (Fig. 4A). Irregular amoeboid cherty patches are locally present in the siliceous black argillite. Unit 1 also forms the footwall rocks to the barite (±sulphides) assemblage (unit 3B). Contact between units 1 and 2 is gradational over a few metres, and contact between footwall and mineralized rocks is sharp and conformable when not faulted.

Mineral assemblages consist of a sulphide-carbonate facies (unit 3A) and a barite±sulphides (mostly pyrite) facies (unit 3B). Unit 3A consists of massive to laminated pyrite several metres thick (30-75 volume per cent) with minor sphalerite and galena, and recrystallized carbonate concretions and fragments (Fig. 4B). Laminae of black cherty mudstone/shale are scarce within the massive section of the mineral assemblage but increase upwards. In the "Main Zone", unit 3A is gradational upwards into a "transitional unit" that consists of alternating beds of mudstone/shale (50-70 volume per cent) and sulphide-carbonate facies (Fig. 4C). In fact, the "transitional unit" represents the uppermost part of the mineral assemblage and corresponds to the transition from a sulphide-bearing mudstone/shale unit to a sulphide-dominant unit. Unit 3B (Fig. 4D) occurs in the "Main Zone" and the "East Zone" (see Fig. 3). This facies is regionally extensive and can be followed over a strike length of several kilometres. According to Insley (1990), it occurs stratigraphically above unit 3A, but can also flank it. Contacts between mineralized and hanging wall units are gradational over tens of metres.

Hanging wall rocks to both mineralized facies consist of a turbidite lithofacies (unit 4) overlain by concretionary shale (unit 5) and nodular shale/mudstone (unit 6). Contacts between these units are all gradational over a few metres. Unit 4 includes a lower, well laminated pyrite-rich mudstone (Fig. 5A) grading upward into a well laminated pyrite-poor mudstone (Fig. 5B). Carbonate concretions are common. They increase in size but diminish in frequency up section, and their degree of recrystallization diminishes away from the mineralized units. We also noted that laminae (pyrite and/or barite, siltite, and shale) frequency diminishes up section.

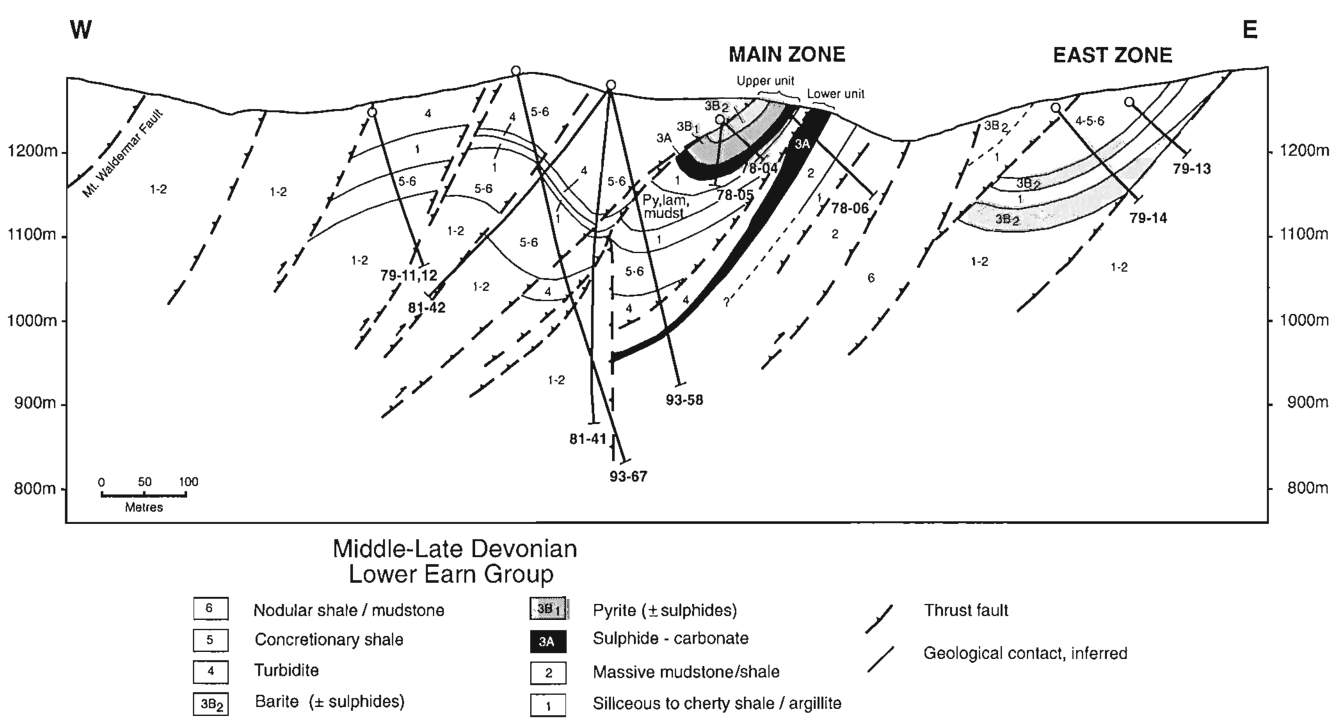

Figure 3. *Stratigraphic section along the Driftpile Creek area (for location see Fig. 2). Abbreviations: Py = pyrite, lam = laminated, mudst = mudstone.*

Unit 5 consists of a moderately laminated grey to black shale with abundant (10-30 volume per cent) carbonate concretions (Fig. 5C).

Unit 6 is a grey to black, thick bedded, poorly to moderately laminated lithofacies, which contains millimetre-sized mudstone (±calcite, barite, pyrite) nodules and sporadically distributed carbonate concretions (< 1 volume per cent) (Fig. 5D). It is the most common lithology in the Driftpile Creek area.

Structure

The structure at Driftpile is dominated by northeastward-thrust panels and a series of thrust faults and related folds trending northwest and plunging gently (10-20°) northwest (see Fig. 2, 3). The fold and thrust patterns result in intense imbrication and repetition of strata in a northeast direction, whereas units, notably the sulphide and barite deposits, tend to be laterally continuous in a northwest direction, parallel to the structural grain. Thrust faults dip moderately southwest. Within the panels, strata are folded into open folds with moderately southwest-dipping to vertical axial planes and gently northwest-plunging fold axes. Overturned folds occur proximally to thrust faults. Small-scale upright chevron folds are developed in the cores of larger folds. This feature renders vergence data from sparse outcrops in the Driftpile Creek valley of equivocal value in interpretation of large-scale structures. In the central part of the section (see Fig. 3), a southwest-dipping reverse fault offsets the western limb of a syncline. In drill cores, this fault and others with demonstrable offsets are expressed as crush zones several metres wide. Such obvious correlation between offset and mechanical expression casts some doubt on the existence of cryptic unrecognizable bedding-parallel thrust faults in core, although they cannot be ruled out.

Several cleavages are observed in outcrop and in core, but one is clearly dominant. It strikes northwesterly and is steeply dipping to vertical. It is axial planar to the major folds. Teck Exploration Ltd. geologists used the general consistency of this cleavage as an aid in core orientation. This technique proved useful, except in holes with very steep dips and near major thrust faults. In the argillites, cleavage-bedding angles are typically moderate to large, in keeping with the open

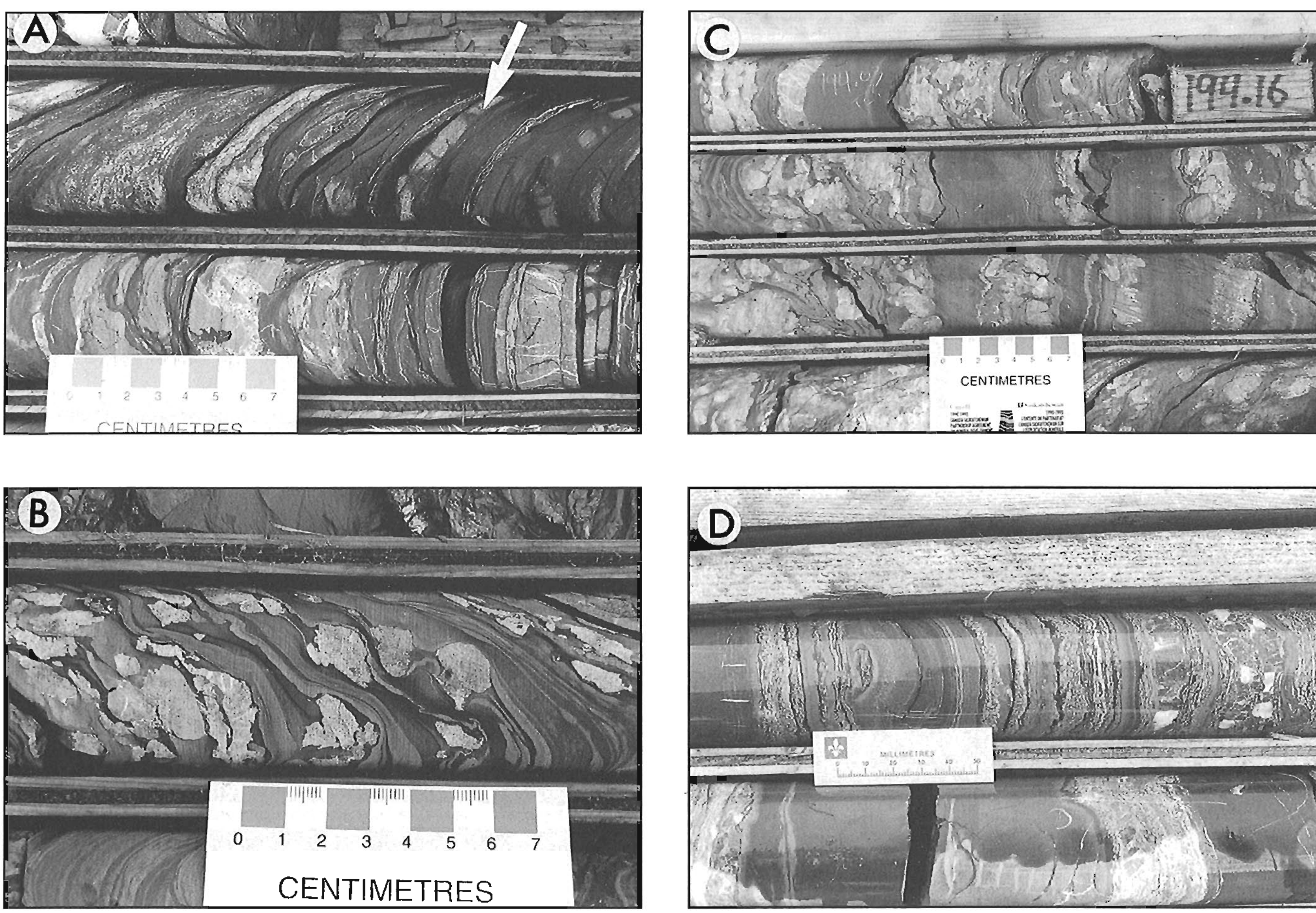

Figure 4. ***A)*** *Unit 1 - well laminated siliceous to cherty argillite interbedded with radiolarian-rich beds (arrow).* ***B)*** *Unit 3A - sulphide-carbonate facies: finely laminated pyrite and mud with laminae disrupted by carbonate concretions.* ***C)*** *Unit 3A - "transitional unit".* ***D)*** *Unit 3B - barite (±sulphides) facies: blebby barite beds interlaminated with thin pyrite and black shale laminae.*

nature of the folds. As the sulphide-carbonate zones are approached, however, bedding is transposed into cleavage and mesoscopic folds are abundant. It is likely that the more ductile sulphides folded disharmonically and were thickened in fold hinges to a greater extent that their host argillites.

Mineralization

At Driftpile, mineral assemblages consist of a sulphide-carbonate facies (unit 3A) and a barite (±sulphides) facies (unit 3B). This bimodal occurrence is also observed elsewhere in the Kechika Trough and the Selwyn Basin (Dawson and Orchard, 1982).

Archer, Cathro, and Associates Ltd (Carne and Cathro, 1982), McClay et al. (1988), and Insley (1990) recognized one main sulphide-rich unit, 3-30 m thick, restricted to the vicinity of Driftpile Creek, and three to five laterally extensive, 2-20 m thick barite-rich units. Some of the barite-rich units can be followed more or less continuously over a strike length of 50 km, and are repeated complexly by folding and thrust faulting (McClay et al., 1988; McClay, 1991).

Our preliminary results indicate two distinct facies of mineralization that possibly occurred in at least two different stratigraphic positions - a localized and lower sulphide-carbonate facies (unit 3A) present in the "Main Zone", and a laterally extensive and upper barite (±sulphides) facies (unit 3B) present in the "Main Zone", in the "East Zone" (Fig. 3), and elsewhere in the Driftpile Creek area.

Unit 3A consists of massive to laminated spheroidal and framboidal pyrite associated with subordinate amounts of fine grained sphalerite and galena, and concretionary carbonate interbedded with carbonaceous siliceous mudstone and chert (Fig. 4B). The laminations are strongly deformed as a result of soft sediment deformation, transposition along cleavage planes, and displacement by recrystallized coarse-grained carbonate concretions (Fig. 6A). The lowest beds of unit 3A tend to be more thickly laminated, banded, or massive and may include fine grained galena and finely laminated sphalerite. Pyrite occurs as either euhedral grains and aggregates of grains or, most commonly, framboidal/spheroidal recrystallized clusters. The latter forms the massive pyritic ore facies. Sphalerite occurs as intergrowths and interstitial grains

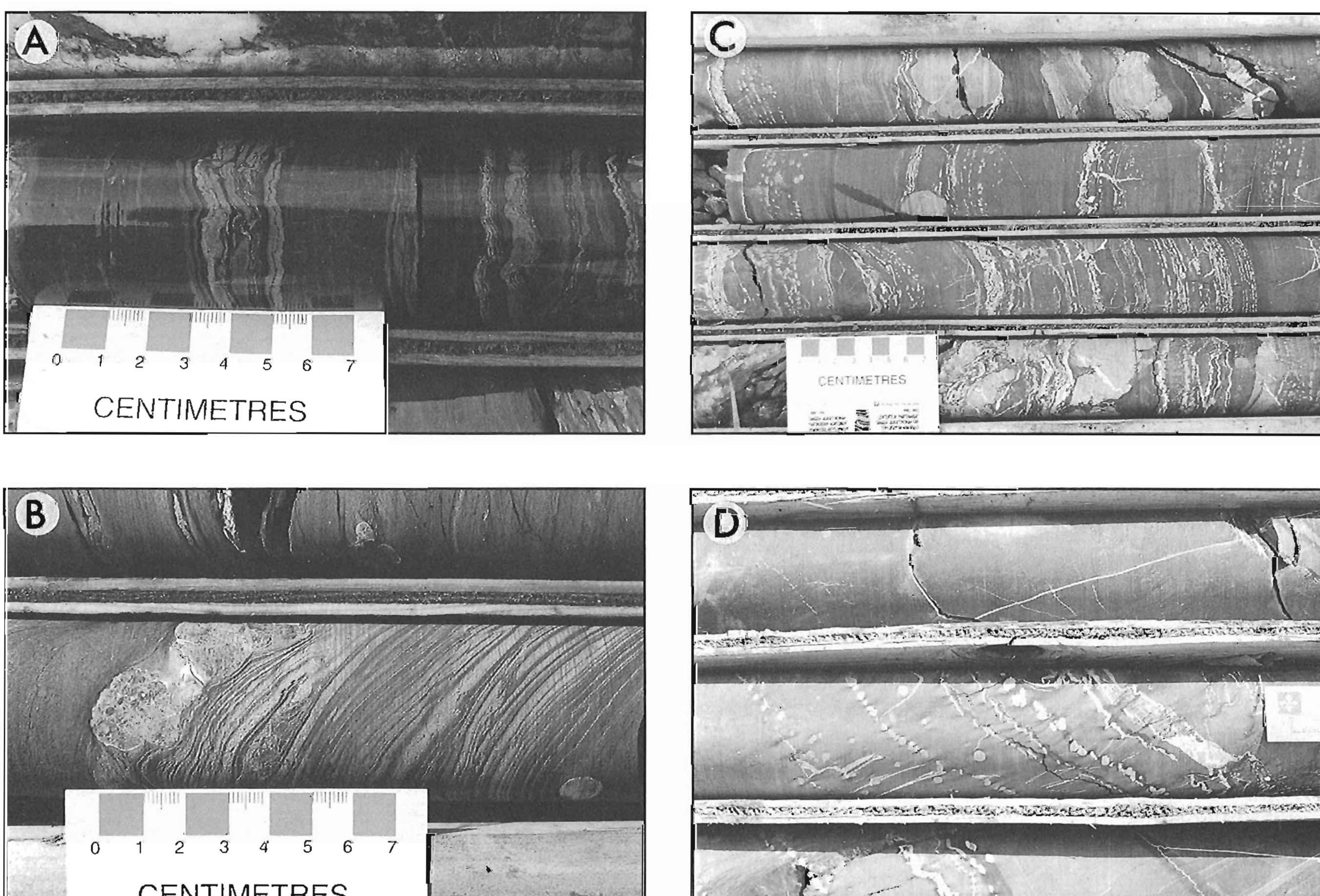

Figure 5. *A) Unit 4 - fine pyrite laminae in massive black mudstone.* ***B)*** *Unit 4 - fine siltite beds in greyish massive mudstone; note the presence of carbonate concretions.* ***C)*** *Unit 5 - laminated mudstone with abundant carbonate concretions and beds of carbonate nodules.* ***D)*** *Unit 6 - grey mudstone with beds of nodules.*

and as fine-grained laminations. Galena occurs principally as coarse-grained recrystallized aggregates and as remobilized fracture and veinlet fillings.

Unit 3B has been divided into two subfacies - a lower pyrite-rich (±barite) ($3B_1$) and an upper barite-rich (±pyrite) ($3B_2$). Both consist of variable proportions of rhythmically interbedded massive, laminated, and blebby barite, laminated pyrite, and black siliceous argillite and chert. Unit $3B_1$ consists of up to 70 volume per cent of pyrite laminae interbedded with beds of blebby barite and siliceous black argillite. Massive, poorly laminated pyrite (and minor sphalerite and galena) occurs at the base this unit. Unit $3B_1$ is gradational upwards into unit $3B_2$. The latter consists of beds of blebby and laminated barite (up to 60 volume per cent) and variable amounts of laminated pyrite (minor sphalerite and galena) interbedded with massive siliceous black argillite (Fig. 6B, 6C). The surface expression of this unit is an orange to pale brown weathering gossan called the barite "kill-zone".

The "East Zone" is well exposed on the south bank of Driftpile Creek and is similar to unit $3B_2$. The base consists of a less than 1 m thick zone of semi-massive laminated pyrite (±galena and sphalerite) interbedded with barite laminae. It is overlain by a bedded sequence of blebby and laminated barite and variable amounts of laminated pyrite interbedded with siliceous black argillite and beds rich in carbonate concretions. Barite amounts increase and pyrite amounts decrease towards the top of the mineralized units. Well laminated siliceous to cherty carbonaceous shale and argillite interbedded with centimetre-sized radiolarian-rich chert bands form the footwall to the "East Zone". This unit may also be interbedded with mineralized units, although possible repetition due to faulting has to be considered.

The mineralized units have undergone intense deformation due to folding, cleavage development, and imbrication (McClay, 1984, 1991; McClay and Insley, 1986; Insley, 1990, 1991), and they appear strongly cleaved with transposition of fabrics (Fig. 6A, 6B).

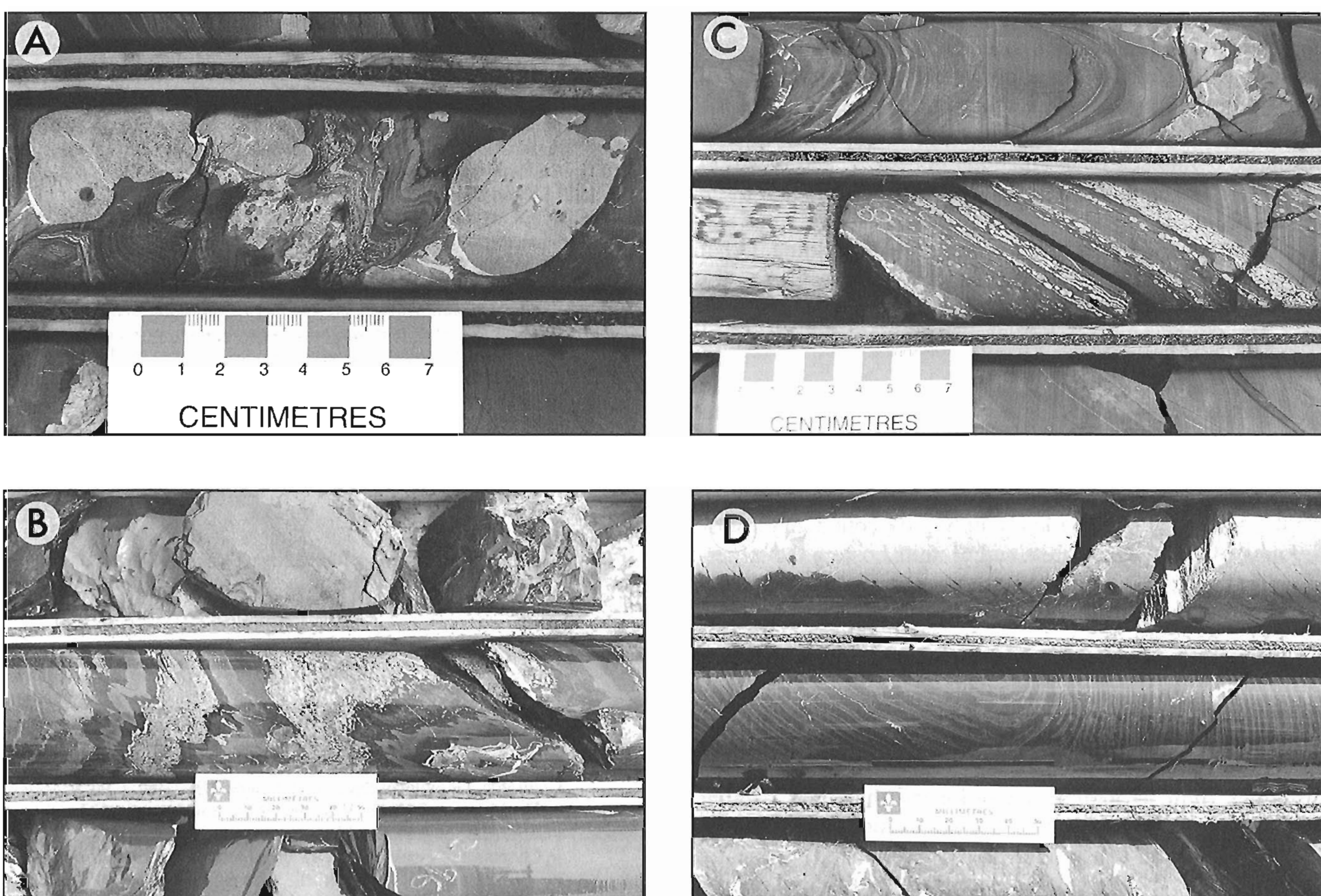

Figure 6. *A) Highly sheared and transposed pyrite and mudstone laminae within the sulphide-carbonate facies unit (unit 3A). **B)** Unit $3B_2$ - Highly sheared and transposed barite, pyrite, and siliceous argillite beds. **C)** Fine beds of blebby barite in laminated mudstone. **D)** Cryptic pyrite laminae in massive siliceous black mudstone.*

DISCUSSION AND CONCLUSIONS

Core logging, construction of a east-west stratigraphic section, and mapping in the Driftpile Creek area have led to the establishment of a preliminary stratigraphic and structural framework of the stratiform barite (± sulphides) and sulphide-carbonate mineralization. Parallel structural panels bounded by thrust faults and related folding dominate the structure at Driftpile (Fig. 3). The stratigraphic section in Figure 3 illustrates broad synclinal structures that form the cores of the "Main Zone" and the "East Zone", and occupy separate thrust panels. Two distinct facies of mineral assemblages are identified: a sulphide-carbonate facies (unit 3A) present only in the "Main Zone", and a barite (± sulphides) facies (unit 3B) present in the "Main Zone" and in the "East Zone". They may represent two distinct mineralized horizons - a stratigraphically lower and restricted sulphide-carbonate facies (unit 3A), and an upper, regionally extensive barite (± sulphides) facies (unit 3B). The latter may also form an apron outboard of unit 3A, but this has yet to be proven. Unit 3B probably stratigraphically overlies unit 3A because of their slightly different stratigraphy (see Fig. 3). The footwall of unit 3A is massive, poorly laminated, moderately siliceous, and rarely radiolarian bearing. A well laminated turbidite sequence with abundant pyrite laminae and carbonate concretions forms the hanging wall. Unit 3B has a footwall sequence composed of siliceous radiolarian-bearing shale interbedded with cryptic pyrite laminated mudstone (Fig. 6D) and an hanging wall sequence composed of interbedded units 4, 5, and 6. Moreover, Ba-Zn-Pb mineral assemblages of unit $3B_2$ in the "Main Zone" and in the "East Zone" are similar, probably correlative, and separated as a result of faulting and folding. Other barite (± sulphides) facies units that crop out few kilometres north of the Driftpile Creek area could also correlate with unit 3B. Some of them have been sampled for conodont biostratigraphic study.

The number of distinct mineralized units and their stratigraphic positions are difficult to establish because of the structural complexity and the lack of reliable marker horizons. Moreover, existing conodont data are too limited to resolve the problem. Detailed sampling for conodont biostratigraphy of mineralized units and of units hosting the mineralization should enable better correlation between mineralized units, lithofacies, and thrust panels.

In summary, the stratigraphy of the Lower Earn Group at Driftpile outlines at least two cycles of mineralization, each preceded by a period of anoxic, starved sedimentation (as evidenced by the footwall siliceous shale and radiolarian chert; unit 1) and succeeded by an influx of fine grained distal turbidites (unit 4) and/or concretionary shale and nodular shale/mudstone (units 5 and 6). Turbidites (unit 4) separate the sulphide-carbonate facies (unit 3A) from the stratigraphically higher, laterally extensive barite (± sulphides) facies (unit 3B). This stratigraphic sequence represents an interplay of mineralizing and tectonic events during which discrete episodes of faulting first triggered the release of metalliferous basinal brines into anoxic bottom waters, then, shortly afterwards, the sudden and rapid influx of turbidites possibly accompanied by convective overturning of the water column. Ideally, detailed conodont biostratigraphy should resolve the relative passage of time between footwall rocks, various mineralized units, and hanging wall rocks.

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Quaternary geology and terrain inventory, Eastern Cordillera NATMAP Project. Report 1: regional landslide characterization[1]

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Jackson, L.E., Jr., 1995: Quaternary geology and terrain inventory, Eastern Cordillera NATMAP Project. Report 1: regional landslide characterization; in Current Research 1995-A; Geological Survey of Canada, p. 159-166.

Abstract: The Foothills area in southwestern Alberta can be subdivided into four provinces, each with distinctive types or suites of landslides. The Porcupine Hills/Interior Plains are characterized by slumps and earthflows on slopes steepened by fluvial activity. The Foothills have a low frequency of landsliding since slope angles are generally smaller than bedding plane dips. Glaciolacustrine valley fills form the floors of Foothills valleys and fail as rotational slumps and flows. Rockslides cluster along major thrust faults in the Rocky Mountains where Mesozoic clastics have been sheared. Rock avalanches are largely confined to Paleozoic units. Slope instability apparently does not decrease with time in the Rocky Mountains. Ongoing mass wasting processes appear to be steepening slopes below massive, mountain-forming Paleozoic carbonates and quartzites.

Résumé : Le secteur des Foothills dans le sud-ouest de l'Alberta peut être subdivisé en quatre provinces présentant chacune des suites ou des types distinctifs de glissements de terrain. Les collines Porcupine/plaines de l'Intérieur sont caractérisées par des décrochements et des coulées de terre sur des pentes que l'activité fluviale a rendues plus escarpées. Dans les Foothills, la fréquence des glissements de terrain est faible, car l'angle des pentes est en général moins prononcé que le pendage des plans de stratification. Les matériaux de remplissage des vallées glaciolacustres forment le fond des vallées dans les Foothills et subissent une rupture par glissements rotationnels et coulées. Les éboulements sont fréquents le long des failles de chevauchement majeures dans les montagnes Rocheuses, là où les roches clastiques du Mésozoïque ont été cisaillées. Les avalanches de pierres sont en grande partie limitées aux unités du Paléozoïque. Il semble que l'instabilité des pentes ne décroisse pas avec le temps dans les montagnes Rocheuses. Les processus actuels de mouvement de masse semblent accentuer l'escarpement des pentes sous les roches carbonatées et les quartzites massifs paléozoïques qui forment des montagnes.

[1] Contribution to Eastern Cordillera NATMAP Project

INTRODUCTION

A second season of field mapping of the surficial geology of the Foothills of southwestern Alberta and contiguous areas of the Interior Plains and Rocky Mountain Front Ranges was undertaken in May-August 1994 as part of the National Geoscience Mapping Program (NATMAP) aimed at bedrock and surficial geology of this region (Jackson, 1994). At the conclusion of the 1994 field season, preliminary digital maps of 8 of the 12 1:50 000 areas to be mapped were completed and all field notes were entered into a relational database (FIELDLOG). Two other map areas were partly covered (Fig. 1). As an adjunct to this systematic and detailed mapping program, five young geoscientists acting as junior and senior field assistants received training in surficial geology mapping, Quaternary stratigraphic investigation, and computer methods. This paper and two companion papers by senior assistants (Leboe, 1995; Little, 1995) summarize highlights of fieldwork and preliminary findings during the past field season.

REGIONAL LANDSLIDE CHARACTERIZATION

Accurate mapping of surficial geology has wide-ranging applications in land use planning and natural resource protection and conservation, including the identification of geological hazards. The surficial geology of the region has been mapped at 1:250 000 or 1:125 000 (Stalker, 1957, 1959, 1962; Jackson, 1986). Landsliding was either not recorded or incompletely recorded on these maps. Ongoing mapping has shown that landslides are common within the study area and are arguably the most widespread geological hazard. An inventory of all landslides (resolvable at 1:50 000) will be included with surficial geology maps at the completion of this project. This data will be combined with bedrock mapping data within a geographic information system (GIS) as one of the overall products of this Eastern Cordillera NATMAP project.

Landslides have been mapped in over 75 per cent of the project area. The distribution and type of failures have led to preliminary observations about relationships between occurrence and type of landsliding and underlying bedrock and surficial geology. A description and discussion of these relationships are given in this paper.

General geological framework and landsliding styles

Bedrock and surficial geology naturally divide the project area into four provinces with respect to landslide activity. These are the Interior Plains and Porcupine Hills, the Foothills, the Rocky Mountains, and glaciolacustrine valley fills. Each province is characterized by a distinctive type or suite of slope failures. These are described below and their distribution is presented in Figure 2 with particularly notable areas of occurrence shown in Figures 3-5. Landslides are classified using the system of Varnes (1958). Landslides were mapped whether they appeared active or inactive. Any reconnaissance classification of landslides is by necessity subjective and many of these failures are complexes of several failure types. A brief description of categories recognized is presented below. Smaller scale mass wasting phenomena such as small scale rockfalls, in which the area covered by deposits is too small to be resolved at 1:50 000 scale, are not addressed in this summary.

Landslide categories

Rotational slump/earthflow

Slope failures were identified as rotational slumps where there was clear evidence of the failure having at least partly rotated about an axis so that the upper part of the failure has dropped and the lower part has risen relative to the unfailed slope. The slumping process commonly reworks failing sediments to such an extent that slumps grade into earthflows downslope.

Earthflows are tongue shaped or lobate in plan and slug-like in profile with a bulbous terminus. They may have concentric transverse ridges near the terminus. They give the impression of movement by plastic deformation through creep rather than by sliding as a mass along a failure plane. They typically lack brittle cracking of the surface.

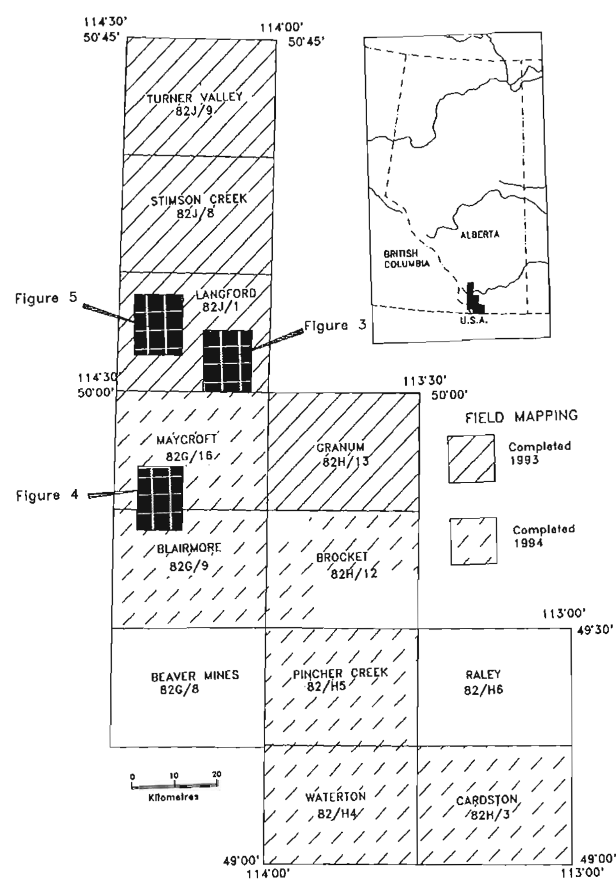

Figure 1. Location map showing 1:50 000 map areas for which surficial geology mapping has been completed.

Rockslide: These failures lack evidence of rotation and display evidence of spreading and breaking up of the slope. A linear failure plane is inferred by morphology. Rockslides may grade into earthflows along their toes. Also included in this category are complex sagging slopes that may be deforming along deep-seated failure planes. Rockslide failure may have occurred by creep or at perceptible rates.

Rock avalanche: These failures are almost entirely confined to Paleozoic and Precambrian bedrock on steep mountain slopes. Former slopes are entirely broken up into bouldery rubble. Failure rates were rapid enough to propel avalanche material partly up adjoining slopes. The largest of these failures are dip slope bedding plane failures. However, notable failures have also occurred by collapse of cliff faces on scarp slopes.

Porcupine Hills and Interior Plains

The Porcupine Hills and adjacent Interior Plains are underlain by latest Cretaceous and Paleocene Porcupine Hills, Willow Creek, and St. Mary River formations. These units are predominantly shale and sandstone with minor coal, conglomerate, and coquinoid limestone (Douglas, 1950). Bedrock is unfaulted and folded over kilometres. Dips range from less than 5° over most of the area to 30° near the western margin of the Porcupine Hills. Shales and some sandstone units, particularly within the Willow Creek and St. Mary River formations, are bentonitic, erodible, and readily form badland topography where exposed. Relief within the Porcupine Hills reaches 350 m along the crest but decreases to less that 100 m along the eastern and western margins.

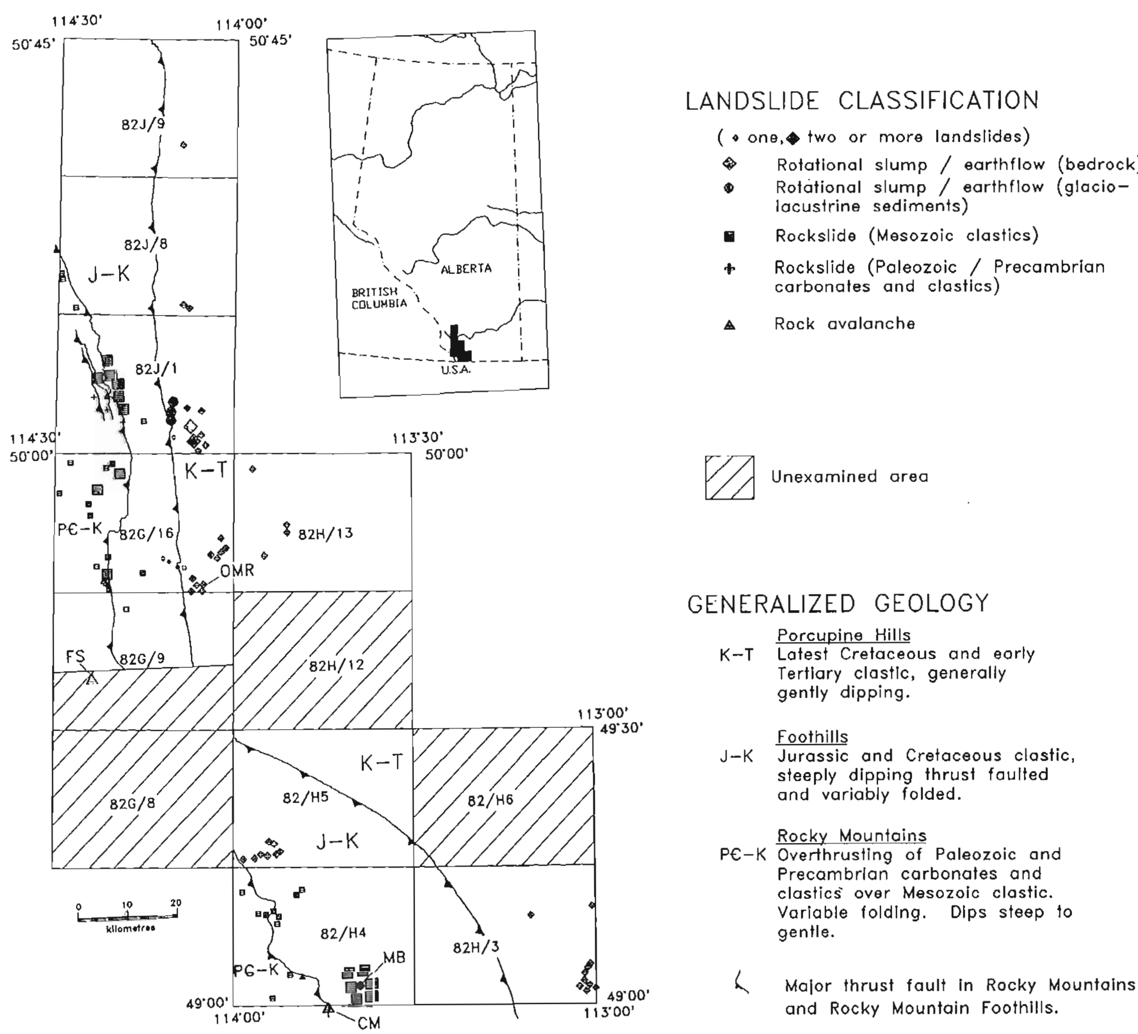

Figure 2. Regional characterization of landslides and associated bedrock/structural provinces. Bedrock geology generalized after Douglas (1950, 1952), Norris (1955, 1958), Okulitch and Woodsworth (1977). Locations: CM- Chief Mountain; FS- Frank Slide; MB- Mokowan Butte; OMR- Oldman River between its confluences with Callum Creek (north end) and Crowsnest River (south end) (particularly active area of slumping involving late Cretaceous bedrock and overlying glaciolacustrine sediments).

Style and occurrence of landslides

Landslides are predominantly rotational slumps and earthflows that originate within shale units (Fig. 2 (indicated by diamonds) and 3). They occur along valleysides within higher relief areas of the Porcupine Hills or on the outside of active contemporary stream bends or the margins of former meltwater channels such as the Callum Creek and Oldman River valleys west of Porcupine Hills. Oversteepening by glaciofluvial or fluvial erosion of valleysides underlain by weak, expansive shales appears to be the underlying cause of slope failures within this geological/structural province.

Foothills

This belt encompasses northwest-southeast trending asymmetric ridges underlain by steeply usually westward dipping Jurassic to late Cretaceous sandstone, shale, conglomerate. These are variably folded and repeated along closely spaced thrust faults (Douglas, 1950).

Style and occurrence of landslides

Landslides originating entirely within bedrock are relatively uncommon within this belt compared to the Porcupine Hills to the east and the Rocky Mountains to the west. The largest

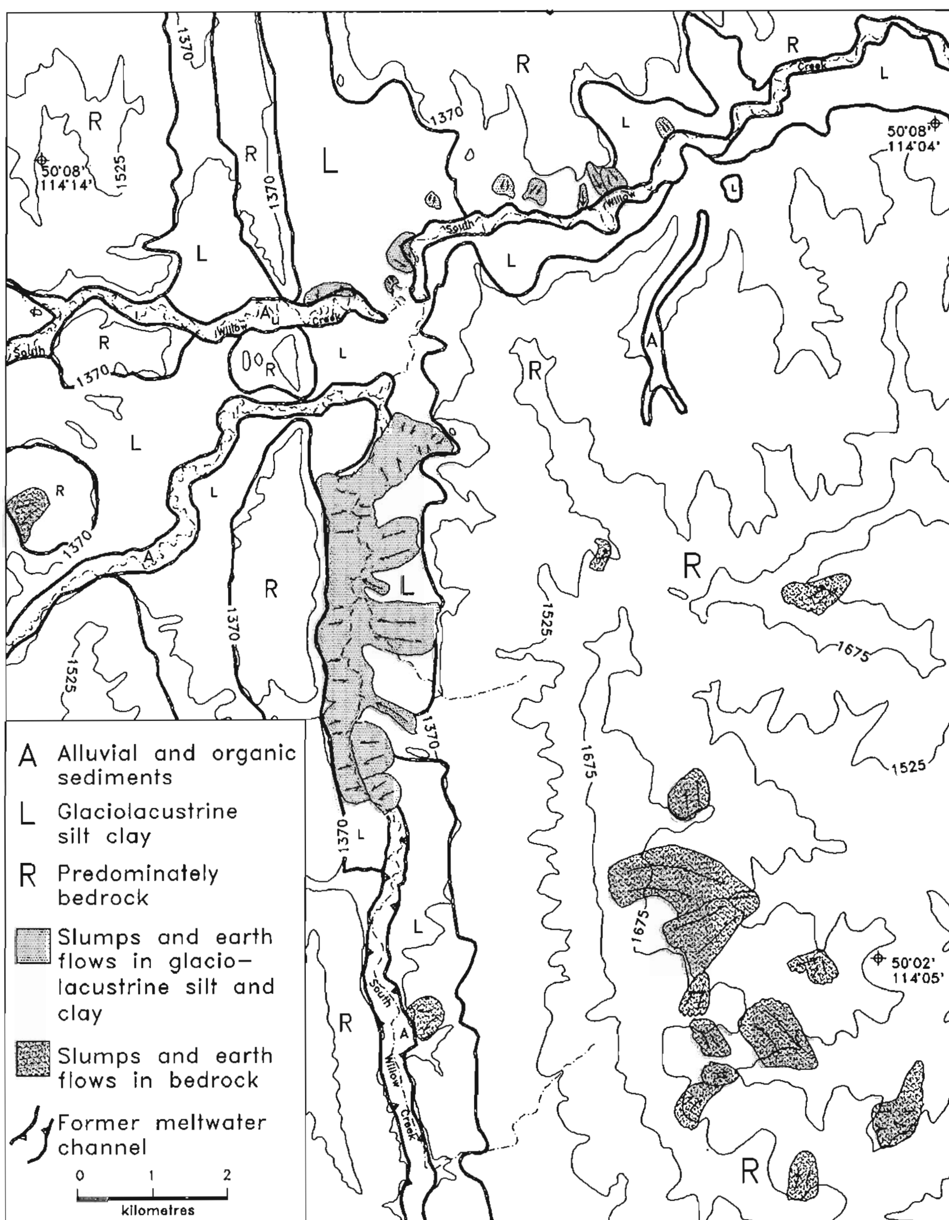

Figure 3. Slumps and earthflows in the Porcupine Hills and extensive failures in glacial lake sediments, upper reaches of South Willow Creek and its major tributaries.

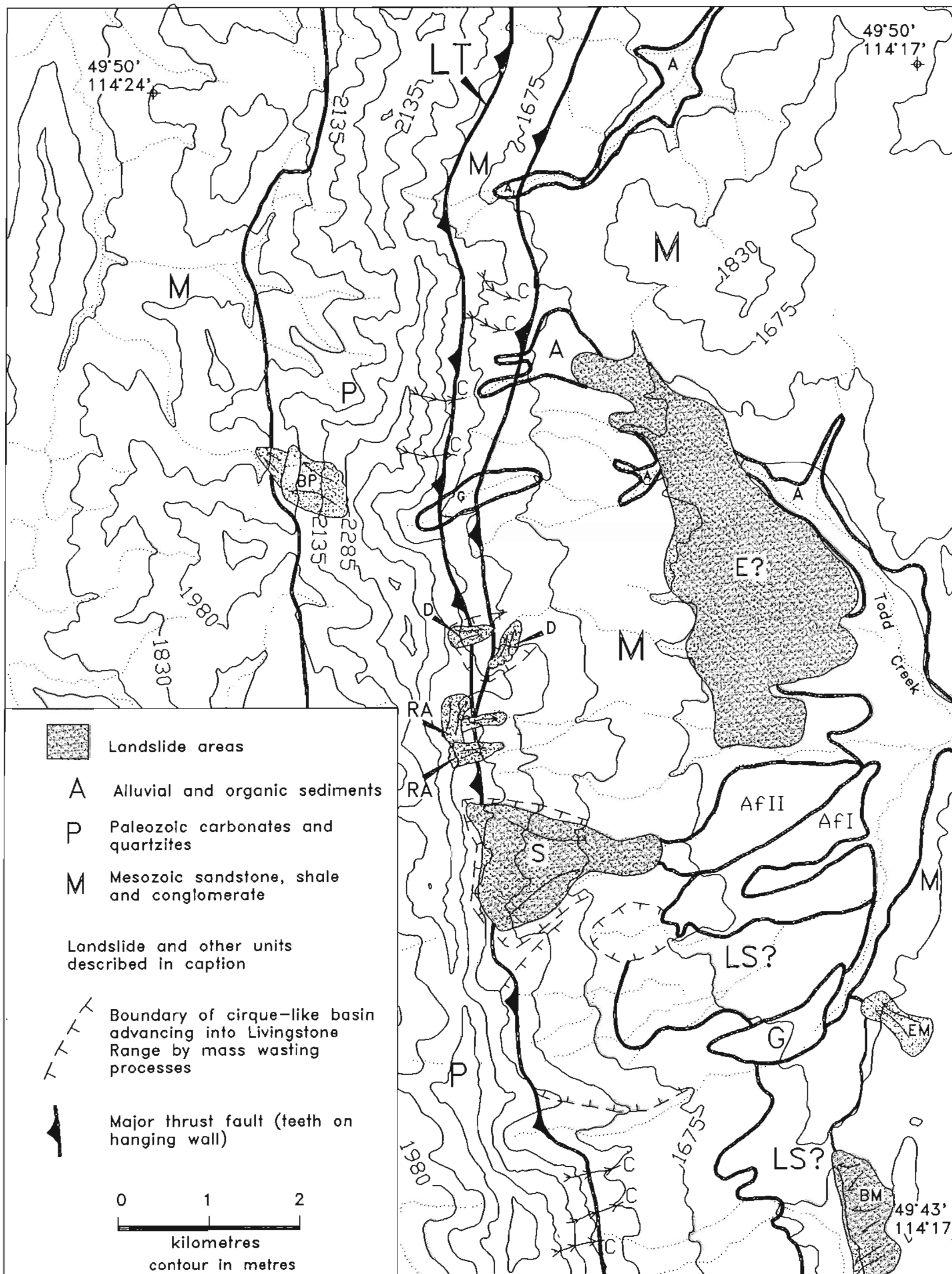

Figure 4. *Landslides and related features along the Livingstone Range near Todd Creek (bedrock geology generalized after Douglas (1950) and Norris (1955)). LT- Livingstone Thrust; BP- bedding plane rockslide in Paleozoic bedrock; RA- rock avalanche in Paleozoic bedrock; S- apparent sagging slope, terminates as an earthflow at toe; D- rockslide in Mesozoic clastics oblique to bedding planes; E?- forested area of lobate topography with no natural or artificial exposures. Possibly a complex of earthflows; EM- earthflow in Mesozoic clastics; BM- bedding plane rockslide in Mesozoic clastics; AfI, AfII- succession of fans built by gravels derived from mass wasting; C- large fresh ravines being incised in the Livingstone Range; LS?- heavily forested area of anomalous landforms, possibly of landslide origin; G- bouldery glaciofluvial gravels.*

landslides in the Foothills are planar rockslides along dip slopes where slope angles and bedding dip angles are similar (Fig. 4, BM). Slides also occur where bedding is intersected obliquely to dip along steep-sided valleys.

The Mokowan Butte area (Fig. 2) is the most extensive area of landsliding in the Foothills. This area will be discussed in the section on the Rocky Mountains because the stress history of rocks has been complicated by shearing along the Lewis Thrust. This exception notwithstanding, several factors account for the relatively low occurrence of landsliding in the Foothills:

1. Rocks are generally more competent in the Foothills than are younger units in the Porcupine Hills and have not been weakened by shearing to the extent that correlative units have within the Rocky Mountains.
2. Dip slope bedding inclinations are predominantly steeper than slope angles. Such underdip slopes have been shown to have a low frequency of landslide activity elsewhere in the Rocky Mountains (Cruden and Hu, 1993).
3. Ridges of Mesozoic clastics are not surmounted by thrust sheets of massive Paleozoic or Proterozoic rocks (as in the Rocky Mountains), which provide significant driving forces.

Glaciolacustrine valley fills

Large valleys in the Foothills, particularly west of the Porcupine Hills, are underlain by extensive fillings of glaciolacustrine clay and clayey silt (Fig. 3). These deposits were laid

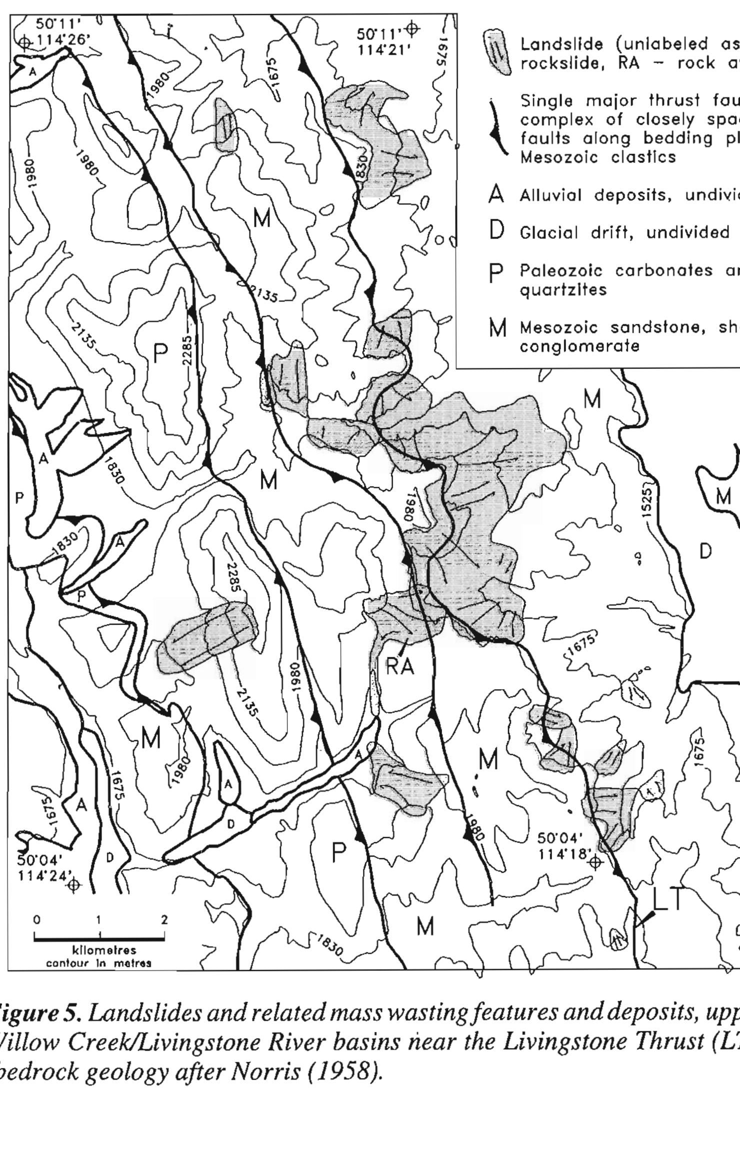

Figure 5. Landslides and related mass wasting features and deposits, upper Willow Creek/Livingstone River basins near the Livingstone Thrust (LT); (bedrock geology after Norris (1958).

down as a result of damming of the valleys by glacial ice during the late Pleistocene (Alley and Harris, 1974; Jackson, 1980; Douglas, 1950).

Style and occurrence of landslides

Glaciolacustrine sediments have mean in situ moisture contents above the plastic limit and values sometimes approach the liquid limit. They creep noticeably in many artificial cuts (Jackson, 1987, p. 20-26) and have failed massively along contemporary stream valleys and former meltwater channels. Failure was apparently initiated as rotational slumping but was transformed into mass flow. The most extensive failures have occurred in the upper reaches of South Willow Creek in the Langford Creek map area (Fig. 3). Here, part of a major north-south trending meltwater channel, estimated to have been 10-20 m deep, has been obliterated by infilling through mass flowage of glaciolacustrine sediment from gentle slopes (less that 10°) to the east and west. This channel was active during the last (Late Wisconsinan) ice age, based on mapping and stratigraphic investigations carried out during the course of the present study. Consequently, landsliding along the meltwater channel occurred during the past 10-15 ka. This type of failure continues to occur. Many new or reactivated failures of this type were noted in glaciolacustrine sediments on the outside of stream bends along South Willow Creek and its tributaries during near-flood conditions in the exceptionally wet summer of 1993.

It is my conclusion that failure occurs in weak plastic glaciolacustrine sediments where resisting forces in slopes are reduced by glaciofluvial or fluvial erosion of slope toes. This mechanism is not unique to glaciolacustrine sediments. Glaciolacustrine sediments and underlying upper Cretaceous bedrock commonly fail together in large slump complexes along steep-sided stream valleys or meltwater channels that have been incised through glaciolacustrine sediments and underlying upper Cretaceous bedrock. Slump complexes along the Oldman River between its confluence with Callum Creek and Crowsnest River are the best examples (Fig. 2, OMR).

Rocky Mountains

Two distinct provinces of the Rocky Mountains are included within the study area. In the Waterton map area, mountain forming Precambrian clastic and carbonate rocks are brought to the surface as low angle thrust sheets. At the mountain front, they are thrust faulted over upper Cretaceous clastic rocks along the Lewis Thrust (Douglas, 1952). North of and including the Maycroft map area (82G/16), mountain forming Paleozoic carbonates are brought to surface along low to moderately dipping northwest-southeast imbricated thrust faults (Douglas, 1950; Norris, 1955, 1958). Triassic to upper Cretaceous rocks are overridden, complexly folded, or sheared along these faults. Thrust faults follow bedding planes within Mesozoic clastics such as the Kootenay, Fernie, and Blairmore groups.

Style and occurrence of landslides

Two styles of failure dominate in the Rocky Mountains: rock avalanches and rockslides. Both tend to cluster above, below, and across major overthrust faults in the Rocky Mountains (Fig. 2, 4-5). Rock avalanches occur predominantly in Paleozoic or Precambrian carbonates and clastics. The best known example in the region is the Frank Slide (Cruden and Krahn, 1973; Fig. 2, FS). The largest in the area mapped to date is a bedding plane failure along the drainage divide between Livingstone River and Willow Creek drainage basins (Fig. 5, RA), which displayed marked mobility characteristic of these failures. However, rock avalanches related to non-bedding failure planes are also common. The two examples indicated by RA in Figure 4 are undated but are thought to be young judging by the absence of vegetation on deposits. The most recent large failure occurred in July 1992 (Dudley, 1992) several kilometres south of the map area in Montana when part of the north face of Chief Mountain (Fig. 2; CM) collapsed in a large mobile rock avalanche estimated to be in the 0.1-1 Mm^3 range based on an overflight during the past field season.

Rockslides occur in Precambrian, Paleozoic, and Mesozoic units but the largest and greatest concentration are in Mesozoic units below and within areas of extensive thrust faulting (Fig. 2, 4, 5). I believe that these failures are due to the combination of sheared incompetent rocks within the zone of overthrusting and driving forces exerted by massive overlying thick carbonates and resistant clastics in the overthrust sheet. The Mokowan Butte area (Fig. 2, MB), previously noted in the section on the Foothills for its anomalous landslide activity within the Foothills landslide province, is the exception that proves the rule in this regard. This upland outlier rises more that 460 m to the east of the Belly River valley and is failing in complex landslides around its entire circumference. Douglas (1952) found residual klippen of Paleozoic rocks above the Lewis Thrust up to 5 km beyond the Rocky Mountain front in this area. Although no Paleozoic rocks have been identified at the summit of Mokowan Butte, it is almost certain that the Lewis Thrust extended over the Mokowan Butte area prior to post thrusting erosion during the Tertiary (M. McMechan, pers. comm., 1994). Study of limited bedrock outcrop by D. Lebel (pers. comm., 1994) indicates that the nearly flat lying late Cretaceous shale underlying Mokowan Butte is sheared. The combination of weak, sheared shale with low dips intersected by steep slopes creates an environment for extensive landsliding at Mokowan Butte.

DISCUSSION

My hypothesis that widespread slope failures along major overthust faults are driven by loading of weak, sheared rocks by overlying thrust sheets offers an explanation for a suite of anomalous features noted along the Livingstone Range in the headwaters of Todd Creek (Fig. 4). Here, large cirque-like basins are eating into overthrust Mesozoic and Paleozoic rocks caught up in the Livingstone Thrust. These features appear to be advancing through a combination of rapid fluvial erosion along steep active ravines in the upper parts of ephemeral streams (C), small and large shallow landslides (D), earthflows (?; based on airphoto interpretation of this heavily forested area), and sagging of large parts of the slopes (E). Glacial deposits (G), which are widespread in all other valley floors in the area, are buried under alluvial and landslide sediments derived from slopes to the west. Fluvial sediments, predominantly gravels, derived from the advance of these cirque-like basins, have been deposited in a succession of fans along the floor of Todd Creek valley. The oldest fans were fed by earlier phases of slope retreat (AfI). Subsequent fluvial erosion and mass wasting have separated these fans from their sediment source areas. Consequently, they are now being incised. Younger fans (AfII) are being constructed around the eroding older fans. The overall result is the undermining of massive Paleozoic carbonates capping the Livingstone Range.

This valley and the upper parts of Willow Creek where similar processes are at work (Fig. 5) were not glaciated during the last ice age. Consequently, slopes in these areas have become less stable over a period of perhaps 100 ka or more. This would seem to contradict recent findings by Cruden and Hu (1993) who concluded that at least one type of landslide (dip slope rock avalanches) become less frequent with time following deglaciation in the Rocky Mountains.

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Geological Survey of Canada Project 930043

Quaternary geology and terrain inventory, Eastern Cordillera NATMAP Project. Report 2: surficial geology and Quaternary stratigraphy, Pincher Creek and Brocket map areas, Alberta[1]

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Leboe, E.R., 1995: Quaternary geology and terrain inventory, Eastern Cordillera NATMAP Project. Report 2: surficial geology and Quaternary stratigraphy, Pincher Creek and Brocket map areas, Alberta; in Current Research 1995-A; Geological Survey of Canada, p. 167-175.

Abstract: The surficial geology of Pincher Creek (82H/5) and Brocket (82H/12) map areas was mapped and sampled and nineteen stratigraphic sections were investigated during the 1994 field season. The upper limit of continuous continental drift cover is at 1360 m on the southern flank of the Porcupine Hills. It is marked by hummocky topography and localized lake plains. Above this limit, the highest Canadian Shield erratic is at 1590 m, and the limit of montane erratics is at 1620 m. Evidence for two glaciations or two major stades of a single glaciation is also found in the southwest corner of 82H/5. No more than two glaciations of the map areas can be supported by continental and montane tills found in stratigraphic sections investigated to date.

Résumé : La géologie des dépôts superficiels dans les régions cartographiques du ruisseau Pincher (82H/5) et de Brocket (82H/12) a été cartographiée, des échantillons ont été prélevés et dix-neuf coupes stratigraphiques ont été étudiées au cours des travaux sur le terrain de 1994. La limite supérieure de la couverture continentale continue de débris glaciaires se trouve à 1 360 m sur le flanc sud des collines Porcupine. Elle est caractérisée par un relief en bosses et creux et par des plaines lacustres localisées. Au-dessus de cette limite, le bloc erratique le plus haut du Bouclier canadien est à 1 590 m, et la limite des blocs erratiques de montagne se situe à 1 620 m. On trouve aussi des indices de deux glaciations ou de deux stades majeurs d'une même glaciation dans le coin sud-ouest de la région cartographique 82H/5. Les tills de montagne et continentaux présents dans les coupes stratigraphiques analysées à ce jour corroborent au plus deux glaciations dans les régions cartographiques en question.

[1] Contribution to Eastern Cordillera NATMAP Project

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INTRODUCTION

Interpretation of the surficial geology of southwestern Alberta has been largely shaped by the work of Horberg (1954), Stalker (1963), Alley (1973), Alley and Harris (1974), and Stalker and Harrison (1977). This work was carried out from the late 1950s to the early 1970s, a period that predates the explosion of research into contemporary glacial sedimentary environments. Furthermore, this work was predominantly based upon interpretation of natural stratigraphic exposures in the absence of detailed regional surficial geology mapping. Also, the region lacks chronological controls on the multiple glaciations documented by these authors. This paper reports on results of the first summer of regional mapping of a part of the Alberta Foothills, which is critical for testing the conclusions of the aforementioned workers, and presents preliminary findings from field investigations of 82H/5 (Pincher Creek), 82H/12 (Brocket), and parts of 82G/9 (Blairmore; Fig. 1).

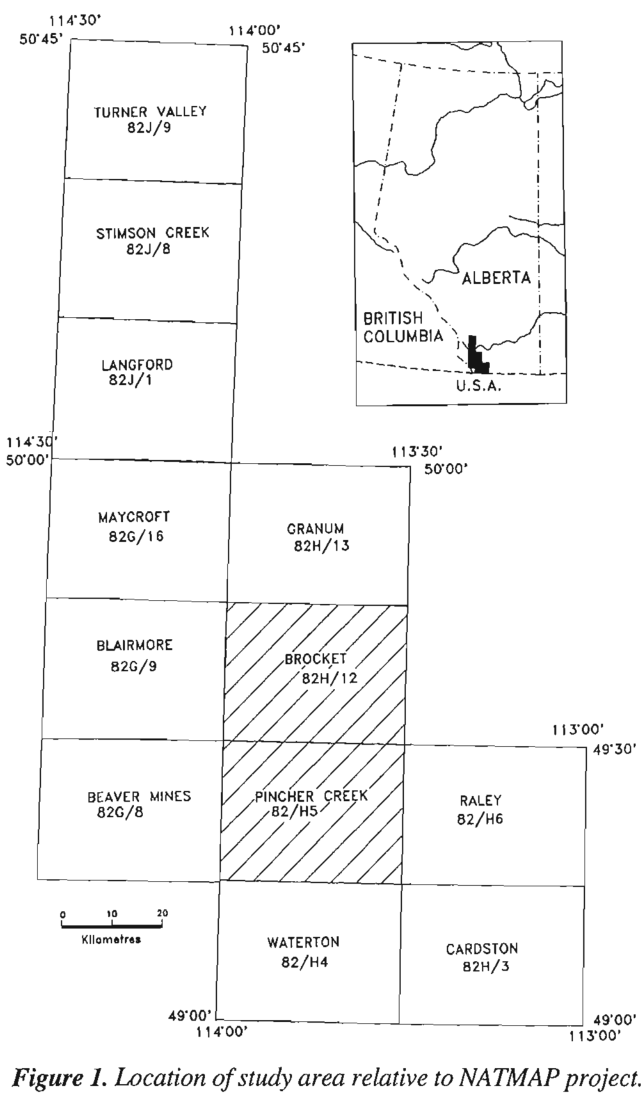

Figure 1. Location of study area relative to NATMAP project.

This research project is part of Canada's National Geoscience Mapping Program (NATMAP); see Jackson (1995). It also will form the basis of the author's Master's thesis at Simon Fraser University (Burnaby, British Columbia). The project will test previous conclusions about the Quaternary history of southwestern Alberta through detailed 1:50 000 mapping and application of contemporary concepts in glacial sedimentology. Significant sections investigated by Stalker (1963) and Alley (1973) will be re-evaluated where exposures still exist. New exposures will also be sought and investigated in the area. The following specific unresolved questions will be addressed as a part of this study:

- Are continental tills and associated sediments in the region the products of two or more glaciations, or are they facies of sedimentary environments representing one glaciation, as proposed by Wagner (1966) and Young et al. (1994)?
- Do distinct glacial limits (from multiple glaciations) exist in the area?
- What is the age of the sedimentary record? Paleomagnetism of sediments will be measured in order to determine if much of the Pleistocene is represented in the sediments in this area, as has been maintained by Stalker (1963) and Alley (1973).

OPERATIONS DURING THE 1994 FIELD SEASON

Airphoto interpretation and field checking

Airphoto interpretation of the surficial geology of the approximately 2000 km^2 of Brocket and Pincher Creek map areas was completed. Field checking of all of Pincher Creek and half of Brocket was completed as well.

Field checking involved east-west and north-south traverses along roads and road allowances, with auger holes driven to 1 m below the surface at approximately 2 km intervals where natural exposures or roadcuts were lacking. Sediments encountered with hand auger or at roadcuts were described. Descriptions included observations of texture, colour, pebble lithology, and other relevant information.

In hillier areas, roadcuts were more common and allowed more direct observation of sediments. Other natural or man-made exposures such as river-cut banks, gravel pits, and slump scarps were described where encountered. In the high relief areas of the Porcupine Hills, where there is little or no road access, traverses were done on foot. The prime objective in ground checking these areas was to establish the limits of the last glaciation and (if there was one) the penultimate glaciation. The highest deposits of glacial diamicton (till) on hillslopes were mapped along with the uppermost occurrences of clasts of Canadian Shield and montane provenances beyond the limits of continuous or discontinuous drift cover.

Of the 350 control points, 200 were done with a hand auger. Thus, overall control point density was approximately one for every 4 km^2. However, the site density was not uniform; density increased with complexity of the landscape as determined by initial airphoto interpretation and initial results of ground checking.

Glacial diamicton (till) and gravel sampling

Exposed diamicton and gravel were sampled at a density of one per township (approximately one per 100 km^2). A sample of 50 randomly selected pebbles was collected from gravel for pebble lithology counts to aid in determining the provenance of the gravel. However, glacially deposited diamicton (till *sensu lato*) was more extensively sampled to determine provenance and genesis, to characterize it as a soil in the engineering sense, and to establish background geochemical values. Specific analyses and their purposes are summarized below.

1. Textural analysis. Weight percentage of sand, silt, clay, and coarser fractions of diamicton will be determined for all samples. This aids quantitative description of diamicton, as well as classification within the Universal Soil Classification Scheme (USCS).
2. Atterberg limits. To be measured on selected samples to classify diamicton according to the USCS.
3. Carbonate content. To be determined by the Chittick method for all samples; this characteristic of diamicton can potentially provide evidence for its provenance, and also its sensitivity to acid precipitation.
4. Pebble lithology. A representative sample of 50 pebbles was obtained from all diamicton sample sites. These will be identified to help to determine diamicton provenance.
5. Heavy minerals. Heavy minerals in the 75 to 150 μm range will be separated and the weight percentage of heavy mineral content determined. Weight percentage of heavy minerals has been shown to be useful for determining provenance of diamicton (continental, montane, mixed; Jackson, 1980). Mineral types will also be determined by optical mineralogy as a further investigation into provenance and as an aid to mineral exploration.
6. Till geochemistry. A broad spectrum of elements will be determined on the clay (< 2 μm) fraction of selected samples. Background geochemical values for this region will be documented through this sampling and analysis.
7. Clay mineralogy. X-ray diffraction analysis of selected samples will identify clay minerals within the clay sized fraction. This has applications in the determination of engineering properties and provenance.

Examination of stratigraphic sections

Nineteen natural sections at 10 different locations were described in detail in order to interpret the age, genesis, and stratigraphic relationships of surficial sediments to help determine the Quaternary history of the region. Of these 10 sections, one was described previously by Stalker (1963) and by Alley (1973). Most well exposed stratigraphic sections are along the western edges of the new Three Rivers reservoir (Fig. 2).

At each section, the steepest accessible cliff faces were cleared of slump debris, then described and measured. Where possible, photomosaics were made of the section. Detailed descriptions of texture, colour, lithologies, sedimentary structures, and clast angularity were made for each unit. Contacts between units were described as depositional, abrupt, sheared, erosional, or gradational, and any structures evident at these contacts were photographed, sketched, or described. Gravel and diamicton units were sampled for textural and provenance studies to aid in interpreting the age and genesis of surficial sediments. Three-dimensional fabric measurements were determined for some diamicton units based on the orientation of 50 rod-shaped pebbles. These were measured to determine ice flow directions and depositional environments, e.g., lodgement, ablation or sediment gravity flow. Samples for paleomagnetic analysis were taken in the two lowest silt and clay units of the Castle River Section. Comparison of results of analysis of these samples to the geopolarity time scale may bracket the deposition of all sediments above the silt and clay.

PRELIMINARY FINDINGS

Glacial limits

Upper limits of glacial deposits occur in two areas: the Porcupine Hills along the northeastern edge of the Brocket map area, and the southwest corner of the Pincher Creek map area.

Porcupine Hills

Glacial diamicton containing erratic pebbles from the Canadian Shield was found to an elevation of 1360 m along the southern margin of the Porcupine Hills in the Brocket map area. Immediately below this limit, hummocky pitted topography can be traced along the slope for several kilometres between 1340 and 1360 m elevation. Some valleys in this area contain local patches of glaciolacustrine sediments at about 1360 m due to continental ice blocking drainage of this part of the Porcupine Hills.

No continuous or discontinuous drift cover exists above 1360 m. However, discontinuous colluvium formed by weathering of local sandstone bedrock contains clasts of erratic lithologies. These lithologies are from both the Canadian Shield and Rocky Mountains. Canadian Shield erratics occur to an elevation of 1590 m, whereas erratics from the Rocky Mountains are present to 1620 m. Although drift cover terminates at about 1360 m, several notches cutting across the ridges at 1420, 1450, and 1490 m have the appearance of meltwater-cut features. Their floors are covered by colluvium containing erratics.

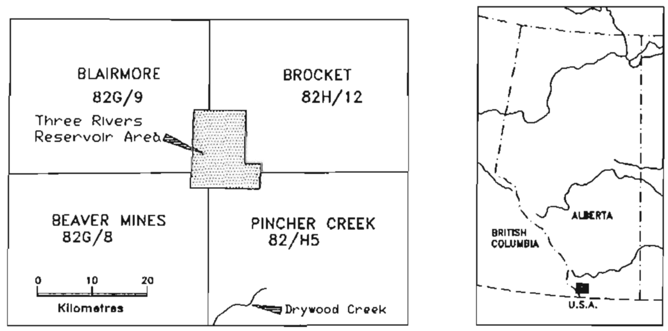

FIGURE 2

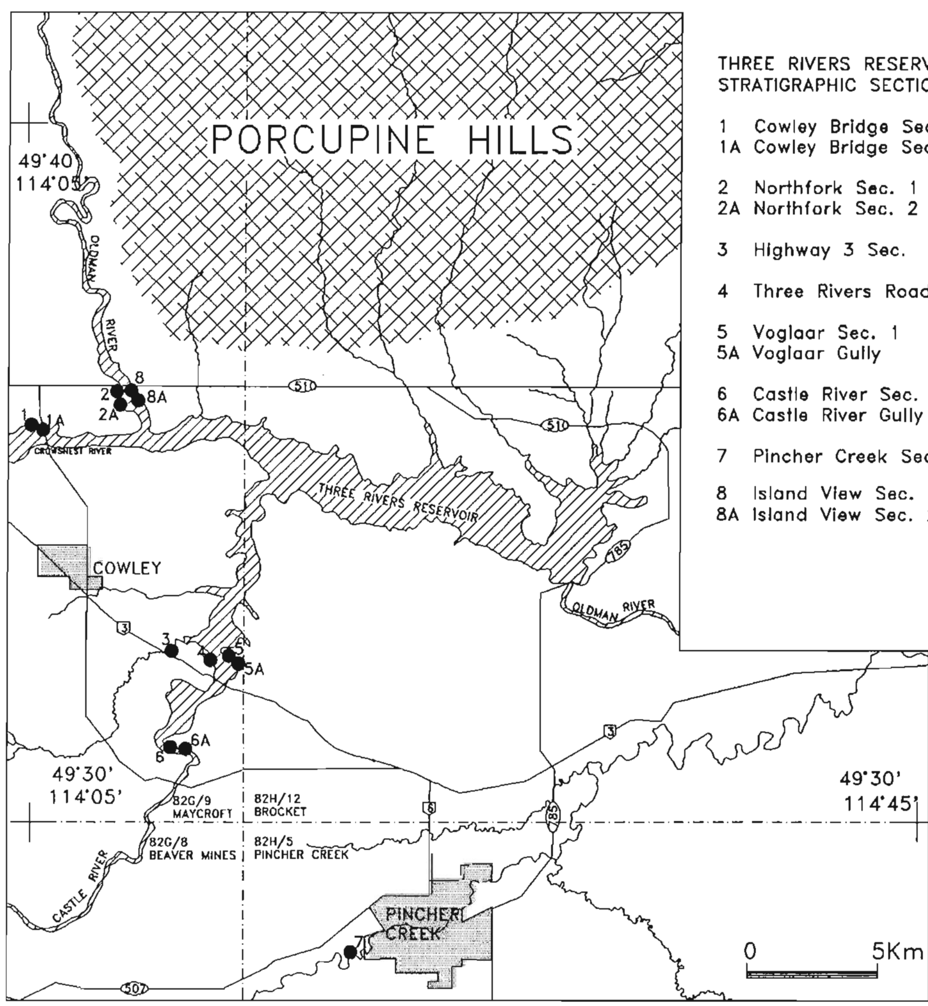

***Figure 2.** Detail of the Three Rivers reservoir area showing locations of most sections and other features.*

Above the 1620 m limit of erratics, bedrock outcrops have a tor-like appearance where they project above weathered sandstone colluvium at their bases.

Southwest corner of 82H/5 (Drywood Creek area)

Recognition of overlapping drift sheets in the southwest corner of Pincher Creek map area (Drywood Creek area, Fig. 2) generally corroborates previous regional mapping by Horberg (1954). Heavy clay and silty clay underlie a glacial lake plain at an elevation of approximately 1520 m. This was formed when continental ice pushed to an elevation of at least 1520 m and blocked drainage, ponding a large, high-elevation lake. Hummocky moraine marking the position of the ice dam lies to the north of this lake plain above 1480 m. The upper limit of this moraine has not been identified. Relief is subdued compared with hummocky moraine to the southeast, and most stones in diamicton units within the moraine are largely disaggregated to depths of 1 m or more. Lake deposits and hummocky continental moraine are in contact with an extremely bouldery moraine of montane provenance.

The montane moraine is located southwest of the aforementioned lacustrine plain deposit. The hillsides are dotted with large erratics of montane (Purcell Group) provenance stromatolitic limestone, vesicular basalt, red and green argillite, and pink quartzite. This landscape is composed of numerous high hills and deep swales, and the relief is approximately 30 m vertically over 100 m horizontally.

The nature of the contact of the belt of high-elevation continental moraine and the belt of montane moraine is apparent along South Drywood Creek. Here, extremely bouldery, poorly sorted ice-contact gravels were deposited off the margin of a lobe of montane ice that advanced out of the local Rocky Mountain valley and deposited the montane moraine. These gravels are inset into the continental-provenance moraine and associated lacustrine sediments.

The bouldery montane gravels and associated moraine and the weathered continental moraine are buried at the limit of a second belt of hummocky and pitted continental moraine trending generally northwest across the southwest corner of Pincher Creek map area. Its upper limit is at approximately 1480 m, marked by a plain of heavy clay deposited in a lake dammed by this moraine at this elevation along South Drywood Creek. At the western edge of this hummocky topography, South Drywood Creek is diverted to the north from its initial path down the southwest regional gradient. Drainage from the Rocky Mountains and meltwater from a stagnating continental ice sheet were blocked by ice to the east and diverted around the ice margin, forming a north-trending channel.

Possible interpretations

Two possible explanations exist for the distribution of moraines and limits of erratics noted in the southwest corner of Pincher Creek map area and the Porcupine Hills in the Brocket map area.

1. A single glaciation characterized by two advances of montane and continental ice, the first stade more extensive than the second. This assumes rapid weathering and erosion of the drift from the first stade.
2. Two glaciations, each with montane ice reaching a maximum extent prior to the climax of the continental advance. This assumes that a significant time period between glaciations was required to remove most or all of the fines in the diamicton on slopes of the Porcupine Hills above 1360 m and to disaggregate stones within diamicton units. After this, a second glaciation (including an advance eastward from the Rocky Mountains and an advance from the Canadian Shield) deposited the topographically lower, fresher moraine and associated lacustrine deposits.

Both hypotheses are equally viable given present data. No unequivocal evidence has been found to document the existence of an interglacial.

Sections

Three natural exposures were studied in detail and were significant enough to merit preliminary description and discussion.

Three Rivers roadcut

Sediments from surface to bedrock were continuously exposed along a roadcut approximately 50 m long and 2 m high (Fig. 3, 4) on the south shore of the Three Rivers reservoir. Distorted and sheared laminated clay rests on bedrock at the bottom of the exposure. Above this is a diamicton of montane origin – no Canadian Shield clasts are present. A three-dimensional clast fabric was determined on the diamicton (Fig. 5-A), showing that the diamicton was deposited by ice flowing from the northwest. It is most likely a lodgement till based upon its sheared basal contact and strongly asymmetric fabric (Dreimanis, 1989).

The lodgement till is succeeded along an erosional contact by partially imbricated gravel. Clasts are mainly of montane provenance, but they contain some granitic and metamorphic lithologies from the Canadian Shield. Their presence leads to the conclusion that although it does not occur in this section, continental drift was deposited in the Castle River basin and eroded, yielding the erratic clasts present in the gravel overlying the montane till.

A silt and silty clay deposit succeeds the gravel and extends to the top of the roadcut. It is at least 20 m thick over much of the area and caps virtually every section near the Three Rivers reservoir.

Voglaar sections

The Voglaar sections are two of a series of gullies cut into a cliff on the Three Rivers reservoir. This 150 m long cliff is 700 m east across a drowned reach of the Castle River valley from the Three Rivers roadcut. Two well exposed gullies

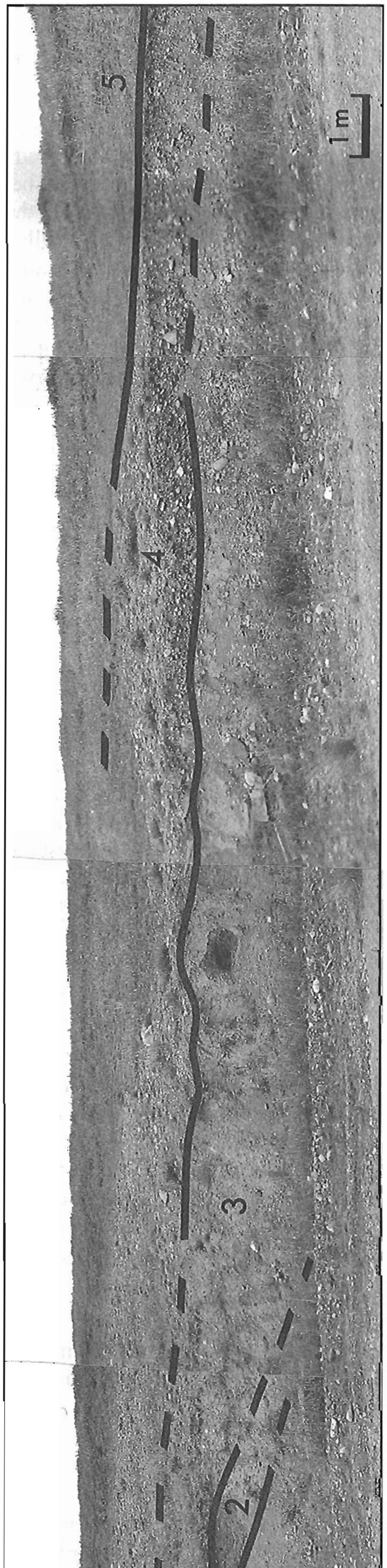

Figure 3. *Photomosaic of the Three Rivers roadcut. Units: 1) bedrock; 2) sheared clay; 3) montane till; 4) montane gravel with some granitic and metamorphic lithologies from the Canadian Shield; 5) silt and clayey silt. Dashed lines are interpreted boundaries.*

through Quaternary sediments and bedrock were described in detail, one near the middle of the section (west Voglaar section) and the other 150 m to the east (east Voglaar; Fig. 4).

In both gullies, diamicton of montane provenance overlies bedrock along a sheared contact. It has a stongly asymmetric fabric (Fig. 5-B), indicating ice flow from the west-southwest. It is most likely a lodgement till based upon its sheared basal contact and strongly asymmetric fabric. Above this montane diamicton is a thin (10-20 cm) unit of sheared silt and clay. The unit above these distorted laminae is a bouldery diamicton, also of montane origin, that has a bimodal fabric with a relatively weak southwest ice flow direction (Fig. 5-C). There is no evidence to suggest that this bouldery diamicton represents another glacial advance. Rather, because of its weak fabric compared with that of the underlying lodgement till and its bouldery texture, it is interpreted as having been deposited as a sediment gravity flow off the surface of a montane glacier into a lake during ice recession.

At this level in the sections, the stratigraphy of the eastern and western gullies differs significantly. In the western gully, the bouldery diamicton is succeeded by rhythmically bedded silts and laminated clays 3 m thick. In the eastern gully, a 1.4 m thick layer of diamicton of continental provenance, containing numerous clasts from the Canadian Shield, occurs between the bouldery montane diamicton and the rhythmically bedded silts and clays. In both sections, rhythmites are overlain along an erosional contact by gravel similar to that at Three Rivers roadcut. Above this contact, most Voglaar cliffs are extensively slumped, and contacts between units are obscured. However, the top units of these sections are massive silt and clayey silt capped by Holocene eolian silt containing *Bison* sp. bones.

Castle River gully

This near-vertical cliff face is approximately 250 m long and between 20 and 60 m high. Parts of this section have been described by Stalker (1963) and Alley (1973), and their interpretations of its stratigraphy have supported their theories of multiple (3 or 4) glaciations in the area. Accessible sections have been redescribed in detail in two places; the second is the gully shown in Figures 4 and 6 and discussed below. The vertical walls of the gully were cleared of all slumped material before starting detailed descriptions.

The lowest Quaternary unit (Fig. 6, unit 2) is a thin, discontinuous layer of stony diamicton, apparently of montane provenance. Due to its location at the lip of a vertical cliff, sedimentary structures within it and its contact with underlying bedrock (unit 1) were not accessible for description although the sediments were sampled. Unit 3 consists of convoluted silt and sand with small pockets of diamicton and coarse sand. Its contact with overlying Unit 4 is sheared. Unit 4 is a diamicton containing only clasts of montane provenance. It is succeeded by a siltier diamicton (unit 5) also of montane provenance, which grades into stratified silts containing angular stones and silty diamicton (unit 6). Over the succeeding 3 m, there is a general fining upwards of sediments, grading into rhythmically bedded silts and laminated clays (unit 7). The top metre of the rhythmites is

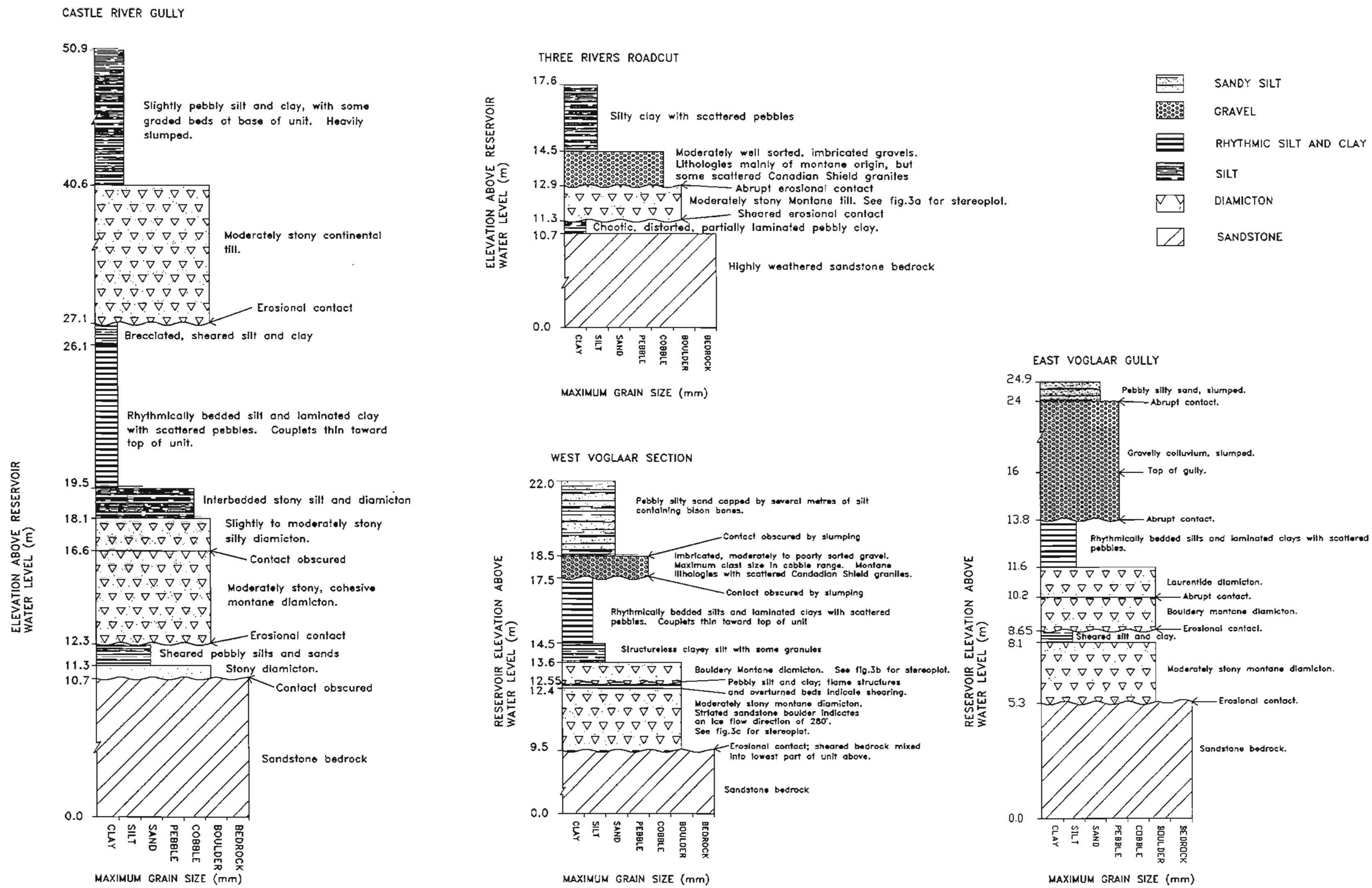

Figure 4. *Schematic diagrams of sections: Three Rivers roadcut (Twp. 7, Rge. 1, Sec. 11, NW 1/4); east and west Voglaar sections (Twp. 7, Rge. 1, Sec. 12, NW 1/4); Castle River gully (Twp. 7, Rge. 1, Sec. 2, SW 1/4).*

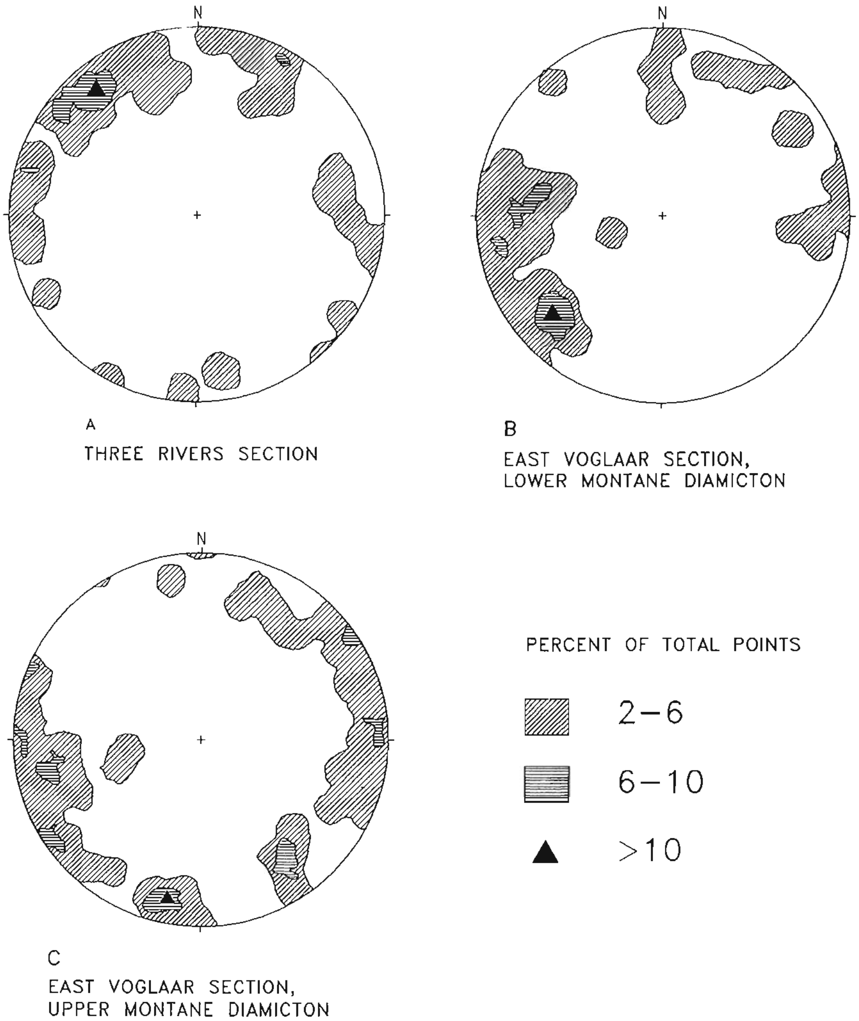

Figure 5.

Stereoplots of diamicton clast fabrics from the Three Rivers roadcut and the west Voglaar section.

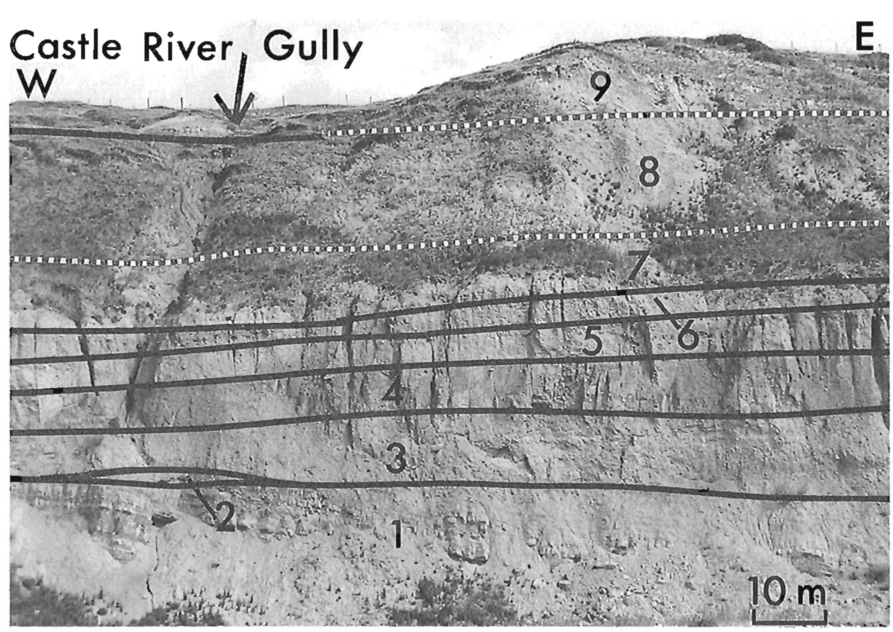

Figure 6.

Castle River gully and near-vertical cliff face to the east. Units: 1) bedrock; 2) sandy diamicton; 3) silt and sand; 4) montane till; 5) silty montane diamicton; 6) stratified stony silt and silty diamicton; 7) rhythmically bedded silt and clay; 8) continental till; 9) silt and clayey silt. Dashed lines are interpreted boundaries.

sheared and brecciated. Unit 8, a glacial diamicton containing clasts from the Canadian Shield, rests upon this sheared and brecciated zone. It is capped by a thick layer of clayey silt (unit 9) which extends to the surface, and correlates with lacustrine sediments seen at the top of the Voglaar section and the Three Rivers roadcut.

Preliminary interpretations of stratigraphic data

Drift from a single continental advance and apparently a single montane advance has been found to date in the Castle River valley. Where two or more montane diamictons are present in a section, no evidence exists to indicate that they represent separate glaciations. Rather, they appear to be facies of a single advance. No more than one glacial diamicton of continental origin is present in any of the sections studied to date. The thick glaciolacustrine silt and clay unit that caps all sections and rests upon an unconformity could have been deposited during a readvance that dammed drainage in the area during final recession of a single glacial advance. Conversely, it could have been deposited during a separate glaciation during which continental ice advanced close to Castle River, forming a proglacial lake. No interglacial sediments or other unequivocal evidence exist to determine the length of time represented by the unconformity at the base of this thick lacustrine sequence. In either case, these findings are dramatically at odds with those of Stalker (1963) and Alley (1973), who found evidence for four glaciations in this region.

ACKNOWLEDGMENTS

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Quaternary geology and terrain inventory, Eastern Cordillera NATMAP Project. Report 3: two continental glacial advances in Waterton and Cardston map areas, Alberta[1]

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Little, E.C., 1995: Quaternary geology and terrain inventory, Eastern Cordillera NATMAP Project. Report 3: two continental glacial advances in Waterton and Cardston map areas, Alberta; in Current Research 1995-A; Geological Survey of Canada, p. 177-182.

Abstract: Glacial limits of two continental advances of decreasing extent were mapped in Waterton (82H/4) and Cardston (83H/3) map areas. These limits are based on clast lithologies, sediment stratigraphy, and geomorphology. Evidence of the older of the two glacial events is present within both map areas, whereas evidence for the recent event is found only within the boundaries of the Waterton map area. Corresponding limits are found on other sheets in the NATMAP study area as well; the general elevation for both of these limits decreases from west to east. The absolute ages of these limits are not known, but relative ages have been determined. Preliminary findings from field stratigraphic and sedimentological investigations are compatible with a single glaciation of southwestern Alberta in the Late Wisconsin.

Résumé : Les limites glaciaires de deux avancées continentales d'étendue décroissante ont été cartographiées dans les régions cartographiques de Waterton (82H/4) et de Cardston (83H/3). Ces limites s'appuient sur la lithologie des clastes, la stratigraphie et la géomorphologie. Des indices sur le plus ancien des deux événements glaciaires sont présents dans les deux régions cartographiques, tandis que des indices de l'événement le plus récent ne s'observent qu'à l'intérieur des limites de la région cartographique de Waterton. On trouve aussi des limites correspondantes sur d'autres cartes de la région d'étude du CART-NAT; l'altitude générale de ces deux limites décroît d'ouest en est. On ignore encore l'âge absolu de ces limites mais des âges relatifs ont été déterminés. Des données préliminaires recueillies lors de recherches stratigraphiques et sédimentologiques sur le terrain sont compatibles avec une unique glaciation dans le sud-ouest de l'Alberta au Wisconsinien tardif.

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INTRODUCTION

Canada's National Geoscience Mapping Program (NATMAP) for the Eastern Cordillera, southwestern Alberta (Fig. 1) was initiated during the summer of 1993 (Jackson, 1994, 1995). In addition to mapping, the surficial geology component of the project involves till sampling for textural analysis, mineralogy of heavy minerals, clay mineralogy, geochemistry, and pebble lithology.

This paper reports on preliminary findings from field mapping and stratigraphic investigations of Waterton (82H/4) and Cardston (82H/3) map areas. In addition to contributing to this NATMAP project, data collected in the course of this study will form the basis of a Master's thesis at the Department of Earth Sciences, University of Western Ontario, London, Ontario.

PREVIOUS WORK

The surficial geology of these map areas has been mapped previously at a small scale (see Shetsen, 1987, for a compilation of recent work), but coverage at larger scales is extremely limited and discontinuous. Harrison (1976) mapped surficial

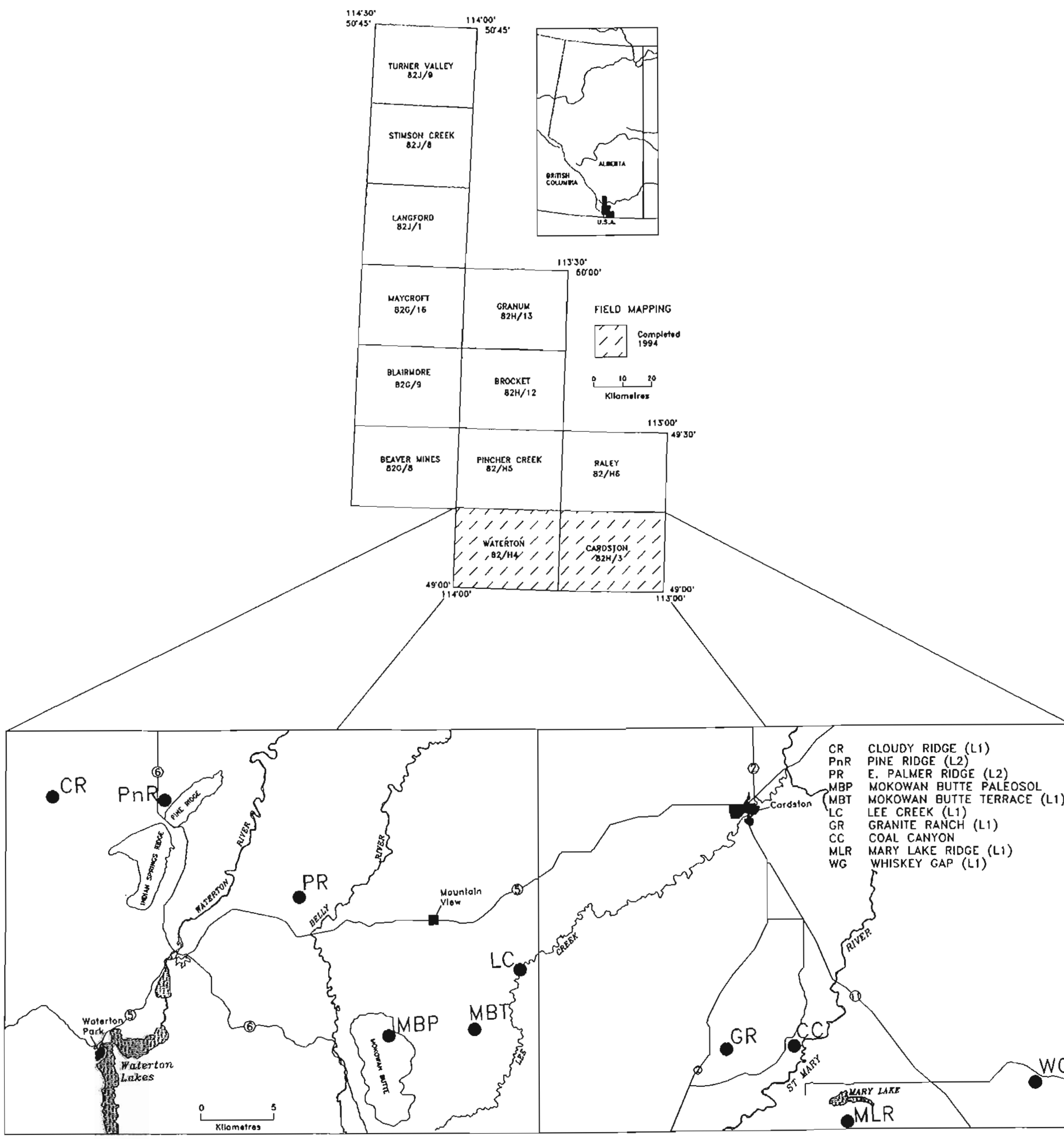

Figure 1. Waterton and Cardston map areas in relation to the Eastern Cordillera NATMAP study area.

deposits in Waterton Lakes National Park and the immediate area surrounding the park; Douglas (1952) mapped bedrock geology in an area including parts of the Lewis and Clark Ranges of the southern Rocky Mountains to the easternmost margin of the disturbed belt to the northeast. Stalker (1963) mapped surficial deposits of the Blood Indian Reserve, which includes parts of the Waterton and Cardston map areas.

The general Quaternary stratigraphic framework of southwestern Alberta has been investigated by various authors (Horberg, 1954; Stalker, 1956; Alley, 1973; Harrison, 1976; Stalker and Harrison, 1977; Jackson, 1980; Jackson et al., 1989; Jackson, 1994). These workers have attempted to decipher the geological record based on a variety of approaches including reconnaissance mapping, regional stratigraphic investigations, and geomorphological studies. Evidence for up to four major glaciations has been recognized in the Waterton Lakes National Park region (Jackson et al., 1989, and references therein).

FIELD METHODS

Preliminary digital surficial geological maps of Waterton and Cardston map areas were completed in the field by the end of the field season. Surficial geology mapping procedures were as follows:

1. The surficial geology of the map area was initially mapped through airphoto interpretation.
2. This was followed by ground truthing using a network of auger holes penetrating the top metre (or until refusal) of sediment and natural and artificial stratigraphic exposures. These stratigraphic data were entered into a computer using the FIELDLOG relational database program.
3. Once airphoto interpretation and data entry were completed, maps were prepared using AUTOCAD release 12. Data and digital maps are currently being readied for open file release.

Investigations of Early Quaternary and Late Tertiary diamictons

In addition to this systematic mapping, some specialized studies of sediments and fossil soils were carried out. The oldest glaciogenic sediments in the region occur at Mokowan Butte, which is located along the eastern margin of Waterton Lakes National Park, and Cloudy Ridge, just north of the northwest corner of the park. Some sediments observed in these sections have been interpreted as tills or possible tills from pre-Wisconsin glaciations (Horberg, 1954; Karlstrom, 1987; Barendregt et al., 1991). Based upon soil analysis and paleomagnetic investigations at Mokowan Butte, diamicton sequences and associated paleosols developed in them are reported to be of glacial (diamictons interpreted as tills) and interglacial (paleosols) origin. The interval represented by this sequence is thought to span as much as 2 Ma (R.W. Barendregt, pers. comm., 1994).

The Cloudy Ridge paleosol and associated sediments have been a topic of controversy for many years (see Williams, 1987, and references therein). Some researchers interpret diamicts in the section to be fan-related deposits whereas others believe them to be glaciogenic deposits. Clast fabrics were measured and paleomagnetic and bulk matrix samples were taken from diamictons at both the Mokowan Butte and Cloudy Ridge sites. Heavy minerals analysis and complete textural analysis, geochemical, clay mineral, pebble lithology, and paleomagnetic studies will be carried out to determine whether any or all of these weathered diamictons are of glacial origin.

PRELIMINARY FINDINGS

Several preliminary observations were made and conclusions reached with respect to the Quaternary stratigraphy of this region. These include definition of approximate glacial limits and evidence of montane/continental ice interaction.

Evidence compiled during mapping suggests that only two glacial events have occurred within the Waterton and Cardston map areas. Although the absolute ages of these events are unclear, their relative ages can be inferred. Evidence supporting only two glacial events also occurs on other map sheets in the NATMAP study area (Jackson, 1994; Leboe, 1995).

Lacustrine sediments and diamictons containing Shield clasts are observed east of the St. Mary River (Fig. 2). They are thought to represent a complex history of interactions between retreating ice and proglacial lake formation. Meltwater from retreating ice in this area would rely on the sill elevation of meltwater channels for proper drainage; depending on the location of the ice front, any number of channels could have been used. Drainage may have been significantly influenced by the Whiskey Gap meltwater channel. Once ice-free, glacial lakes that existed because of restricted flow to the Milk River were drained.

Glacial limits

The provenance of glacial deposits and the limits of glaciation in southwestern Alberta can be determined on the basis of coarse clast lithology, stratigraphy, and geomorphology. The location and lithology of coarse clasts found within drift or deposits derived from drift, and isolated erratics or erratics trains, are of major importance. For example, there are no sources for pink feldspar within igneous and metamorphic rocks in the Lewis Ranges of the Rocky Mountains (Waterton Park and adjacent areas in Montana). The presence of pink feldspar in drift within the Waterton and Cardston map areas indicates a continental (Canadian Shield) provenance for drift. Stratigraphic, topographic, and geomorphological evidence will allow interpretation of possible montane/continental ice interactions and indicate any montane/continental glacial advance sequences that may exist.

Two continental ice limits were identified within the boundaries of the Waterton map area based upon lithological, stratigraphic, and geomorphic evidence (the maximum limits of montane advances in the area were not recognized because they were destroyed or buried by continental ice sheet advances). The best example of the older, more extensive continental ice limit (hereafter referred to as Limit 1) in the Waterton area is found on Cloudy Ridge, at the northernmost tip of Waterton Lakes National Park.

Boulder sized clasts glacially transported from the Canadian Shield (Shield clasts) can be found on Cloudy Ridge at a maximum elevation of approximately 1585 m above sea level. Shield erratics may have reached their present locations on Cloudy Ridge in two ways: 1) they could have been deposited at higher elevations by continental ice then transported to their present location by montane ice or other gravity processes, or 2) they could have been deposited directly at their present locations by continental ice. The latter hypothesis is preferred because of lack of evidence for a continental ice limit higher than Cloudy Ridge. However, Shield clasts were observed only on the surface of the ridge; no clasts with Shield sources were observed within the diamicton that caps the ridge and overlies the diamicton in which the Cloudy Ridge paleosol is developed.

Limit 1 is observed at two other sites in the Waterton map area: on the east slope of Mokowan Butte and approximately 4.25 km northeast of the Mokowan Butte site. The first of these is composed of glaciofluvial terrace gravels containing Shield clasts which are found at a maximum elevation of approximately 1445 m above sea level. Shield clasts observed at the second site have an upper elevation limit of approximately 1440 m above sea level.

Limit 1 can be traced eastward to the Cardston map area where it was documented at three sites. The first site is located approximately 3 km west of Coal Canyon. It consists of hummocky gravel deposits containing abundant Shield clasts to an elevation of approximately 1420 m. Scattered shield erratics continue beyond the limit of continuous drift to 1460 m. where they are found with large montane erratics. These data suggest that continental ice overrode existing montane drift upice in the vicinity and then carried this drift and Shield clasts to their present position.

Ice retreat from Limit 1 is documented by a series of recessional ridges identified in the field and by airphoto interpretation (Fig. 3).

The second ice Limit 1 site is located on Mary Lake Ridge, due south of Mary Lake. This ridge consists of hummocky gravels (Fig. 4) containing abundant Shield clasts, montane gravel, and larger montane erratics; its characteristics are strikingly similar to those observed at the previous site. Shield clasts, montane gravels, and large montane erratics are observed up to the summit of the ridge at approximately 1425 m.

The last Limit 1 site is located on a ridge south of Whiskey Gap. The ridge is bedrock controlled (bedrock outcrops in the area) with overlying gravel deposits containing both montane and Shield clasts to an elevation of approximately 1420 m. The morphology of these deposits is more subdued than at the other two sites. Further evidence of a former glacial margin consists of a series of former ice-walled supraglacial meltwater channels on the east bank of a larger meltwater channel produced during later stages of ice retreat (Fig. 5). Ice margin retreat lowered the floor of the channel(s) producing this

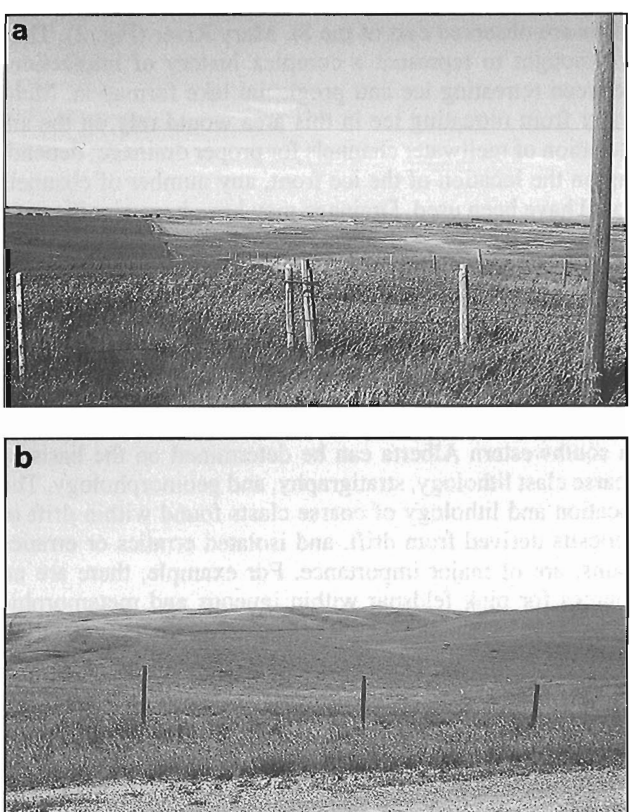

Figure 2. *Views from an approximate proglacial lake shoreline: (**a**) looking north onto a gently rolling lake plain; (**b**) looking south onto glacial drift deposits.*

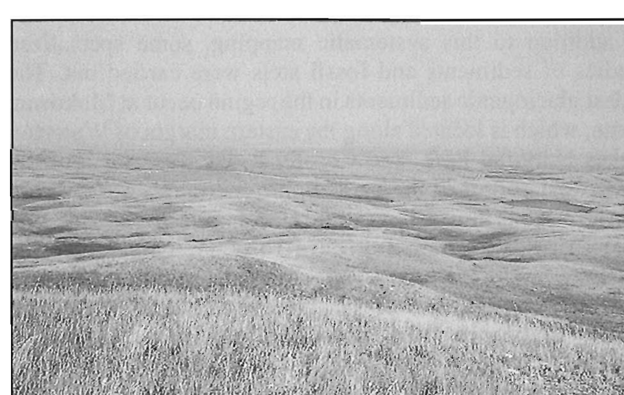

Figure 3. *Parallel recessional ridges below older, more extensive continental ice limits (Limit 1). View to the east.*

Figure 4. Hummocky gravel deposits marking Limit 1 on Mary Lake Ridge. View to the east.

Figure 5. Ridges formed on the landward side of ice-walled meltwater channels. A ridge transverse to the trend of the larger meltwater channel can be seen on the left.

series of descending parallel ridges. The ice front remained stationary for a short period of time, producing an end moraine that is orthogonal to the channel trend (Fig. 5). Once ice began retreating again, meltwater cut the moraine, forming the large meltwater channel that transects the ridge.

Ice limits of the most recent continental advance (hereafter referred to as Limit 2) are evident only in the Waterton map area; no concrete evidence was found for the lower ice limit in the Cardston map area. Unlike supporting evidence for Limit 1, these limits are based primarily on geomorphological observations and the spatial distribution of continental glaciogenic deposits.

Prior to the latest continental advance, a moderate montane advance pushed through Waterton Lakes National Park (Harrison, 1976). Upon retreat of montane ice from this area, eskers and meltwater channels were formed. Water was discharged into a large meltwater channel that parallels Highway 6 and that can be traced to Pine Ridge, where it ends. This, along with the presence of diamictons containing Shield material in the gap between Pine Ridge and Indian Springs Ridge, suggests that at this location, Limit 2 occurs at an elevation of approximately 1400 m.

Another site exists west of Palmer Ridge at approximately 1325 m. It appears clearly in airphotos as a geomorphological transition from strongly kettled (north) to moderately hummocky (south) terrain, the former being deposited by the latest continental advance (corresponding to Limit 2) and the latter, by the penultimate continental advance (corresponding to Limit 1). A major meltwater channel separates these morainic belts. Water in it flowed south along the east side of Palmer Ridge then turned east, and flowed along the front margin of Limit 2 toward Paine Lake.

DISCUSSION

The two drift sheets bounded by limits 1 and 2 have not been dated within the Waterton and Cardston map areas. However, based on the degree of weathering, stratigraphy, and geomorphological evidence, relative ages can be assigned. A difference in the degree of weathering of diamictons is evident in many roadcuts and auger holes. One indication is the appearance of coal at particular depths. In some continental diamictons examined, coal was either completely weathered out or weathered to dark brown down to 70 cm. Below this depth, competent clasts of coal are usually found. However in order to assign relative ages this type of evidence must be used in conjunction with other evidence.

Investigations of stratigraphic sections exposed along streams have also proved to be useful in determining montane/continental ice advance sequences. In those sections where both montane and continental glacially deposited diamictons were observed, a single continental diamicton always overlies montane deposits. No Shield material was observed in any montane drift in any of the sections described. Assuming that continental diamicton is late Wisconsinan in age, the occurrence of a single diamicton is compatible with a single glaciation of southern Alberta as proposed by Young et al. (1994).

This interpretation is contrary to the findings of previous researchers with the exception of Wagner (1966) who also suggested a single glaciation involving a complex relationship between a retreating active ice margin and proglacial lake formation.

CONCLUSIONS

Limits 1 and 2 represent two major episodes of continental ice advance into the Waterton and Cardston map areas. There are currently no absolute ages for these limits in the Waterton and Cardston areas. However, the single continental glacial diamicton recognized in these map areas is compatible with the hypothesis of a single Late Wisconsinan glaciation of southern Alberta proposed by Young et al. (1994).

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