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COMMISSION GÉOLOGIQUE DU CANADA

PAPER / ÉTUDE
90-1E

CURRENT RESEARCH, PART E
CORDILLERA AND PACIFIC MARGIN

RECHERCHES EN COURS, PARTIE E
CORDILLÈRE ET MARGE DU PACIFIQUE

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1990

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ERRATUM

The table shown below was inadvertently ommitted from "Paleomagnetism of the Old Crow Batholith, northern Yukon" by J.K. Park. pages 287-290

Table 1. Magnetic directions and poles

site	sp	type	rcf (mT)	T_{UB} (°C)	D°, I°	R	k	α_{95}°	pole (long°, lat°)
					Old Crow batholith				
2	14 (4)	O_B	<50 (1)	100-400 (3)	000, +72	3.86	21	21	039E, 80N
3	15 (7)	O_B	20-80 (3)	N-450 (4)	323, −79	6.15	7	24	239E, 48N
	15 (5)	O_A	15->200 (2)	N-<600 (3)	318, +25	4.91	45	12	087E, 29N
4	12 (2)	?	<20 (1)	<500 (1)	312, +66[2]	1.46	2	—	—
5	7 (3)	O_B	2.5-35 (2)	<250 (1)	264, −82[2]	2.81	10	—	257E, 64N
6	9 (4)	?	15-100 (2)	N-450 (2)	269, +02	2.47	2	—	—
7	10 (2)	?	—	N-450 (2)	083, +63	1.68	3	—	—
8[1]	4 (1)	?	N-12.5 (1)	—	131, −78	—	—	—	—
9	7 (5)	?	N-80 (2)	N-520 (3)	197, +77	2.87	2	—	—
MEAN		O_B	SITES 2, 3, 5		056, +84	2.94	33	22	262E, 72N
					Granite stocks				
1[1]	4 (4)	O_B	2.5-15 (2)	N-200 (2)	041, +84	3.96	78	11	254E, 75N
10[1]	4 (2)	O_B	—	N-200 (2)	238, +85	1.98	63	—	201E, 62N
11[1]	4 (3)	O_B	2.5-17.5 (1)	N-250 (2)	094, +84	2.97	70	15	248E, 66N
12	15 (11)	?	2.5-30 (5)	N-400 (6)	215, +64	10.56	23	10	195E, 27N
MEAN		O_B	SITES 1, 10, 11		076, +88	2.99	154	10	231E, 69N
MEAN		O_B	SITES 2, 3, 5, 1, 10, 11		061, +87	5.92	64	8	240E, 70N

NOTES: sp = specimens analyzed (specimens yielding magnetic component); D°, I°, = declination, inclination of directions; R = vector resultant; k = precision of direction; α_{95}° = error of 95% probability about mean direction. N = NRM. Components with T_{UB}'s < 100°C and rcfs < 10 mT are not included.
[1] 4 sp analyzed from 10 possible. [2] Normal and reverse directions.

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For description see
Fig. 4, p. 117.

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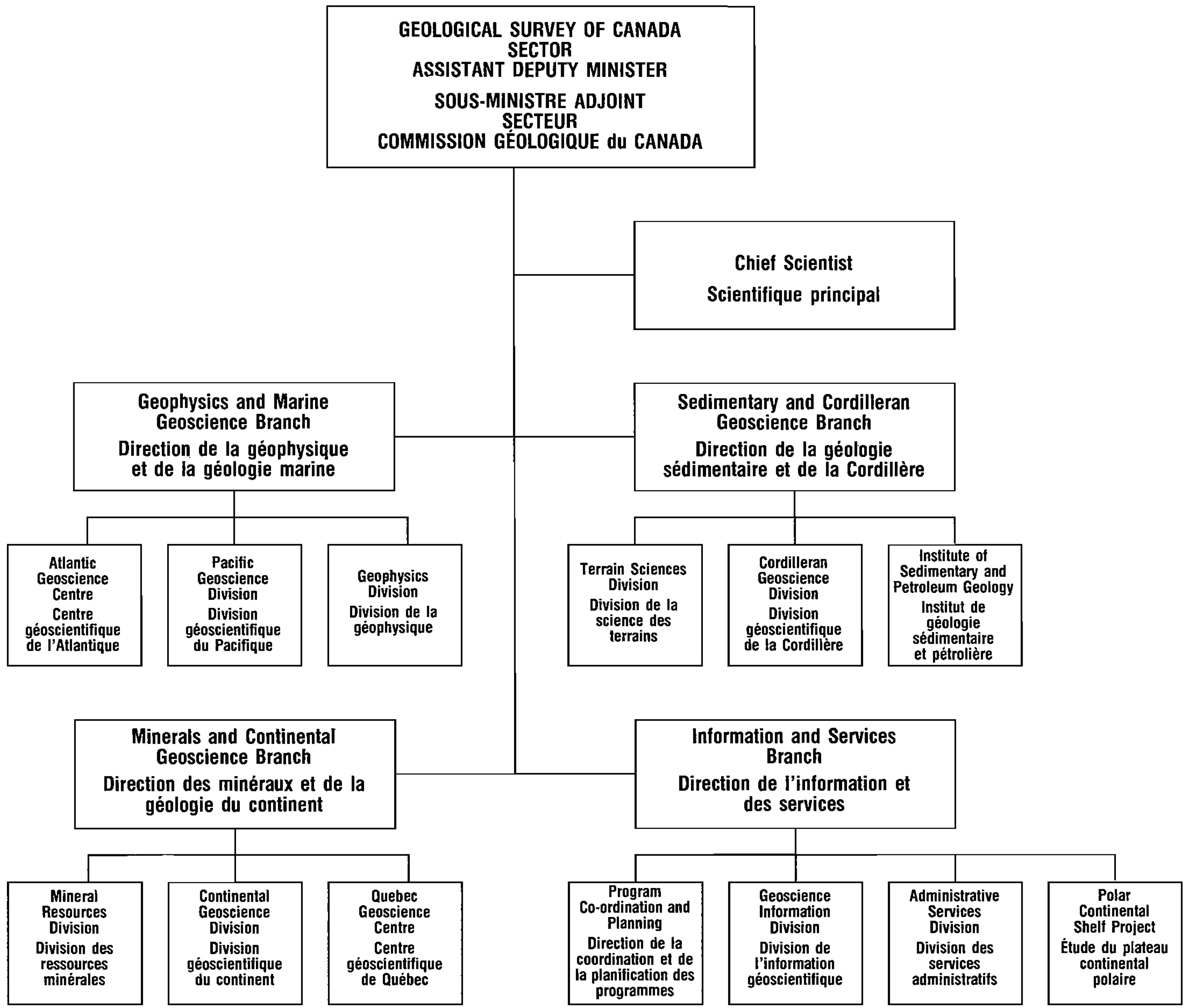
CONTENTS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Selwyn-Wernecke cross-sections, Yukon: a joint Indian and Northern Affairs Canada - Geological Survey of Canada Project

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Cordilleran Division, Vancouver

Abbott, J.G., Gordey, S.P., Roots, C., and Turner, R.J. Selwyn-Wernecke cross-sections, Yukon: a joint Indian and Northern Affairs Canada - Geological Survey of Canada Project; in Current Research, Part E, Geological Survey of Canada, Paper 90-1E, p. 1-3, 1990.

Abstract

The Department of Indian and Northern Affairs and the Geological Survey of Canada are conducting a co-operative one-year project northeast of Mayo, Yukon that includes 1:50 000 scale geological mapping and a study of the Marg volcanogenic massive sulphide occurrence.

Résumé

Le ministère des Affaires indiennes et du Nord et la Commission géologique du Canada exécutent actuellement en commun des travaux qui s'étalent sur une période d'un an au nord-est de Mayo, au Yukon. Ces travaux comprennent l'établissement de cartes géologiques au 1/50 000 et une étude d'une venue de sulfures massifs d'origine volcanique de Marg.

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INTRODUCTION

An awakening interest in base metals in Yukon is reflected in the recent discovery of the Marg Fe-Zn-Pb-Cu-Ag-Au massive sulphide deposit northeast of Mayo, and in rekindled interest in the nearby Blende Pb-Zn-Ag vein property. However, the geology of this economically attractive area is poorly understood. The Department of Indian and Northern Affairs and the Geological Survey are conducting a cooperative one-year study in this region that includes 1:50 000 scale geological mapping as well as a study of the Marg occurrence. The project provides an exceptional opportunity to study the regional setting of an exciting new exploration target (the Marg), and to examine critical stratigraphic and structural relationships in the northern Cordilleran miogeocline where many significant components are exposed. It will provide detailed structural and stratigraphic data in a north-south transect that will assist in the exploration for, and assessment of, other deposits in the region, and act as a framework to guide geological work in poorly understood areas adjacent to the transect. Interaction with industry regarding logistics, geological discussion and access to unpublished data were critical to the project.

This note introduces the project, its location and regional geological setting. Preliminary results are presented in five companion papers that follow by Abbott, Gordey, Turner and Abbott, Roots, and Mustard et al.

LOCATION AND GEOLOGICAL SETTING

The location of the project area is shown in Figure 1. Mapping at 1:50 000 scale has been undertaken in NTS 106D/8 by C. Roots (GSC), in 106D/1 by J.G. Abbott (INAC), and in 105M/16 by S.P. Gordey (GSC). C. Roots and J.G. Abbott jointly mapped 106D/7 east-half. R. Turner (GSC) examined the Marg deposit. In addition, P. Mustard and A. Donaldson (Carleton University) with Roots, examined general stratigraphic features in Proterozoic carbonate of the Gillespie Lake Group at scattered localities in Nash Creek map area (106D).

The area is underlain by Middle Proterozoic to early Mesozoic sedimentary and minor volcanic rocks deposited along the margin of ancient North America. The oldest strata include shallow water siliciclastic and carbonate rocks of the Middle and Upper Proterozoic Wernecke Supergroup and Pinguicula Group that are exposed northeast of the Dawson thrust (Fig. 2). For rocks of Late Precambrian to mid-Devonian age the area straddles a regional boundary that separates "shelf" strata, including shallow water carbonate and clastics (Mackenzie Platform), from equivalent deeper water "offshelf" shale, chert, limestone and turbiditic sandstone (Selwyn Basin). In the project area, preserved shelf strata include lower and middle Paleozoic carbonate, whereas offshelf strata are represented by Upper Proterozoic to Cambrian deep water turbiditic clastic rocks of the Hyland Group. A mid-Devonian to Jurassic shale-dominated clastic succession with quartzite, sandstone, limestone and volcanic rocks overlaps the shelf and offshelf assemblages. Extensive hornblende diorite sills of Triassic(?) age (Abbott, 1990) intrude the succession. Deformation occurred during the Late Jurassic and Early Cretaceous, and was succeeded in the mid-Cretaceous by intrusion of mesozonal to epizonal granitic plutons.

Three significant thrust faults extend eastwards more than 200 km from Dawson into the project area. The Dawson thrust (Tempelman-Kluit, 1982) juxtaposes the offshelf assemblage against the shelf assemblage. The Tombstone thrust has been suggested as a name for a thrust fault near Dawson (R.I. Thompson, pers. comm., 1989) which juxtaposes Mississippian quartzite above Jurassic strata. The Robert Service thrust, documented by Tempelman-Kluit (1970) in the Dawson area, places the Hyland Group onto Jurassic and older strata. In the south (105M/16,

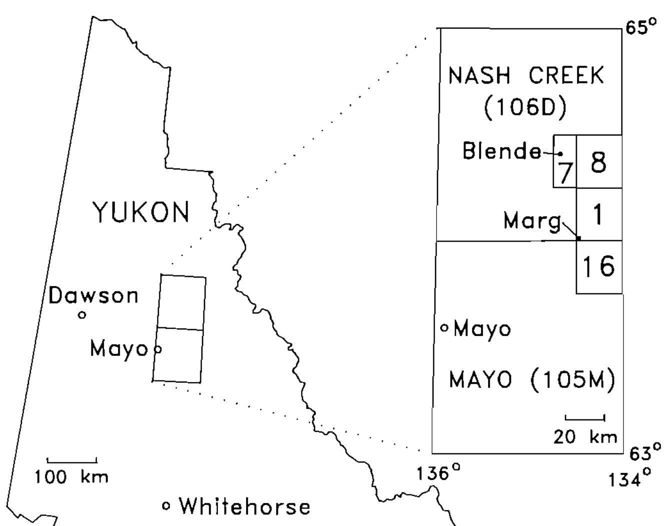

Figure 1. Location of project area showing location of 1:250 000 and 1:50 000 map areas referred to in text.

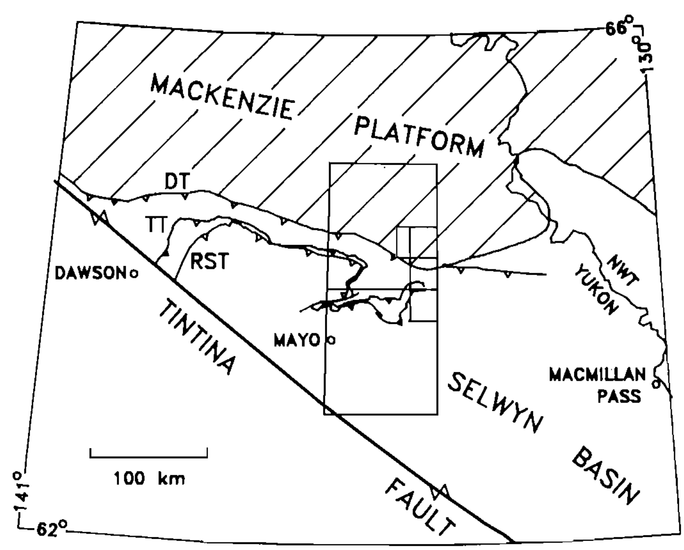

Figure 2. Geological setting of the project area. DT = Dawson thrust, TT = Tombstone thrust, RST = Robert Service thrust.

106D/1) the Robert Service and Tombstone sheets are internally deformed by isoclinal to subisoclinal folds and imbricate splays. This area is dominantly underlain by incompetent latest Proterozoic to Jurassic clastic strata of low metamorphic grade. In the northern part of the area (106D/8, D/7, east half), north of the Dawson thrust, unmetamorphosed late Proterozoic strata and overlying early Paleozoic carbonate are thrust-faulted and gently folded. The Marg occurrence is a Fe-Zn-Pb-Cu-Ag-Au volcanogenic massive sulphide deposit of Mississippian age hosted in folded and thrust faulted meta-pelite and quartzite (the Keno Hill Quartzite). The Blende occurrence is a lead-zinc-silver breccia deposit within carbonate strata of Proterozoic age (Gillespie Lake Group).

PRODUCTS

This project has resulted in significant revisions to stratigraphy and structure that will be portrayed on 1:50 000 scale geological maps, cross-sections, and reports for the areas shown in Figure 1. These will be released as combined INAC-GSC open files in early 1990. A report on the Marg deposit emphasizing geochemistry and petrography of the sulphides and host strata will be released on open file by the fall of 1990.

ACKNOWLEDGMENTS

All project participants express their sincere thanks to Archer, Cathro, and Associates (1981) Ltd., particularly Doug Eaton and Lasha Cymbalisty for their invaluable expediting and logistical assistance. Dave Reid (Trans North Turbo Air) provided superb helicopter support.

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1990: Preliminary results of the stratigraphy and structure of the Mt. Westman map area, central Yukon; *in* Current Research, Part E, Geological Survey of Canada, Paper 90-1E.

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1970: Stratigraphy and structure of the "Keno Hill Quartzite" in Tombstone River-Upper Klondike River map areas, Yukon Territory (116B7/,B/8); Geological Survey of Canada, Bulletin 180, 102 p.
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New geological maps for the southern Wernecke Mountains, Yukon

C.F. Roots
Cordilleran Division, Vancouver

Roots, C.F. New geological maps for the southern Wernecke Mountains, Yukon; in Current Research, Part E, Geological Survey of Canada, Paper 90-1E, p. 5-13, 1990.

Abstract

New geological maps for NTS 106D/7 (east half) and D/8 show the predominance of shallow thrusts in Proterozoic shelf strata that have been correlated eastward over 350 km from the Yukon-Alaska border. Thick sections of Gillespie Lake Group dolostone forms the cores of structural slices thrust over siliciclastic rocks correlated with Middle Proterozoic units in the Ogilvie Mountains, and over lower Paleozoic carbonate strata. Zn-Pb-Ag occurrences are spatially and probably genetically associated with the Middle Proterozoic Gillespie Lake Group.

Résumé

Les nouvelles cartes géologiques 106D/7 (moitié est) et D/8 du SNRC montrent la prédominance de chevauchements peu profonds dans des couches de plate-forme protérozoïques qui ont été corrélées vers l'est, sur plus de 350 km à partir de la frontière entre le Yukon et l'Alaska. Des sections épaisses de dolomie du groupe de Gillespie Lake forment des noyaux d'écailles structurales chevauchant des roches siliciclastiques qui ont été corrélées avec des unités du Protérozoïque moyen dans les monts Ogilvie et avec des couches carbonatées du Paléozoïque inférieur. Des venues de Zn-Pb-Ag dont l'origine est probablement liée au groupe de Gillespie Lake du Protérozoïque moyen, sont associées spatialement à ce dernier.

INTRODUCTION

This report concerns the northern area mapped as part of a joint DINA-GSC Selwyn-Wernecke transect project (Abbott et al., 1990). Exploration interest provided the impetus and opportunity to investigate the geology of eastern Nash Creek map area in detail (e.g. Green, 1972). Current mineral interest centres on the Blende Zn-Pb-Ag property, within the map area. The integrated structural and stratigraphic mapping extends understanding of the Proterozoic geology. It also helps exploration for structurally controlled or stratabound mineralization; there are positive indications in the area for this type of occurrence.

During 1989 the 1:50 000 map areas 106D/8 and the eastern half of adjacent 106D/7 were traversed. This work incorporates and extends unpublished mapping near the Blende property by Grant Abbott of the Department of Indian and Northern Affairs. The geological relations at two other showings (Kathleen and Zap) in the area were also examined.

The south edge of map sheet 106D/8 adjoins the Mount Westman map area (106D/1) described by Abbott (1990). A wide valley along the boundary is underlain by the Kathleen Lakes Fault Zone. No correlation is attempted between the two map areas because only the lower Paleozoic carbonate is common to both, and the degree and style of deformation differ markedly.

REGIONAL FRAMEWORK

The southern Wernecke Mountains are underlain by a Middle and Late Proterozoic shelf assemblage that extends northward and westward beneath lower Paleozoic rocks of the Mackenzie Platform. The shelf assemblage was formed by periodic extension events at the margin of ancestral North America (Thompson et al., 1986). In late Mesozoic time these rocks were folded and thrust northward so that they now form a west-trending fold-thrust belt.

The Proterozoic rocks are regionally exposed in erosional inliers (Fig. 1). Most inliers reveal the Wernecke Supergroup (Delaney, 1981) and a younger, cyclic carbonate-clastic succession. The Wernecke Supergroup is older than intrusive breccia isotopically dated about 1300 Ma (Archer et al., 1986; Parrish and Bell, 1987). The younger succession, here called unit 4, correlates with similar rocks that are capped by a 775 Ma volcanic unit (Roots and Parrish, 1988) in the Ogilvie Mountains.

In each of the inliers the oldest rocks are on the north, and the younger strata are ramped against them on the south. The oldest rocks, subjected to several periods of deformation, are structurally complex. Strata along the southern edge of the Ogilvie Mountains have been useful in distinguishing structures of the Mesozoic fold-thrust belt from those resulting from earlier extension (Thompson et al., 1986).

The area mapped in 1989 occupies a pivotal location in the ancient shelf strata of the northern Cordillera. At this point many stratigraphic units traced from the Ogilvie Mountains undergo lateral facies changes or reflect Proterozoic structural elements (e.g. Eisbacher, 1981). Furthermore, these transitions are reflected by Mesozoic contractional structures, and separated from strata of similar age in the Mackenzie Mountains by the Knorr-Richardson fault system and the lower Paleozoic Misty Creek Embayment (Cecile, 1981). The mapped area is also particularly instructive because it spans the 'foothills' of the rugged central ranges of the Wernecke Mountains. Compared with the geology to the north, the rocks are more diverse and structures are better exposed.

Stratigraphic units in the area (106D/8 and D/7, here called the Mount Good-Mount Williams area; Fig. 2), are described below, each followed by pertinent discussions.

STRATIGRAPHY

Quartet Group

This unit includes dark brown and black siltstone, argillite and minor sandstone which underlie the Gillespie Lake Group. These tough, indurated rocks comprise two large

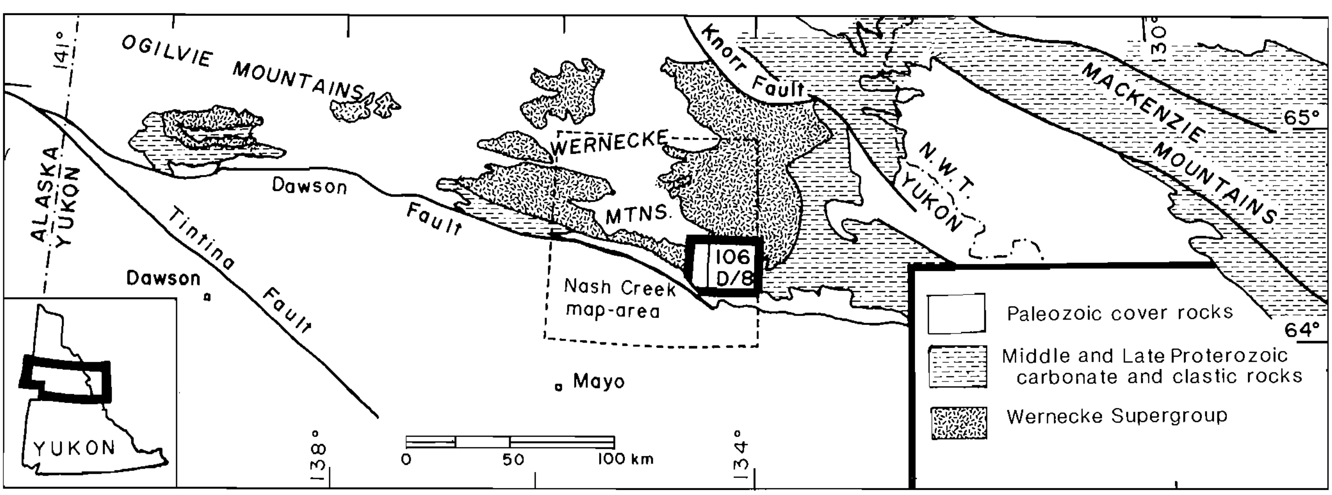


Figure 1. Location of area mapped in 1989. Patterned areas are erosional inliers of Proterozoic rocks, surrounded by the Paleozoic Mackenzie Platform and Selwyn Basin cover rocks.

hanging wall flaps on south-facing mountain slopes, and are exposed in canyons at the base of thrust panels. The base of the unit was not observed, and the top is gradational with the Gillespie Lake Group, marked by an increase in the proportion of interbedded dolostone (Locality 1 in Fig. 2). The preserved thickness is about 250 m.

The fine-grained rocks are uniformly dark and constitute monotonous thin-bedded successions. In structurally complex terrain this lithology is easily confused with dark shales of unit 4: Table 1 outlines the principal characteristics used to separate the two units. Many Quartet beds are normally graded and separated by thin white laminae. In contrast to unit 4, outcrops of the Quartet Group exhibit slaty cleavage (Fig. 3). Low to medium grade metamorphism is indicated by fine muscovite on parting surfaces, and these siliceous rocks do not have conspicuously weathered surfaces.

The Quartet Group represents a homogeneous turbidite succession, as reported by Bell (1986, 30 km northwest of Mount Williams) and by Delaney (1981). Although these authors observed a stratigraphic contact at the top, the Quartet rocks are considerably more deformed than Gillespie Lake strata. About 100 km farther west and in the Ogilvie Mountains, a prominent angular unconformity is present, and in the Rae Creek area large folds of Quartet (C.F. Roots, unpub. data) indicate deformation (and probably the imposition of cleavage) before transition to platform carbonate.

Gillespie Lake Group

Light brown- and orange-weathering dolomite strata form resistant ridges and brightly coloured canyons in a broad band across the Mount Good-Mount Williams area. The unit dominates the northwestern corner of 106D/8 and is virtually continuous northward for 50 km to the type area (Delaney, 1981). Both this region and the central band are interpreted to consist of structurally repeated strata: this

Table 1. Lithological characteristics to distinguish dark clastic rocks of the Quartet Group from those of unit 4.

Quartet group	Unit 4
— monotonous succession of dark grey siltstone	— thick siltstone successions but colour varies; olive and brown
— white 'stringer' alteration along bedding planes, fine laminae	— contains thin sandstone, carbonate interbeds
— pervasive micaceous cleavage; breaks in long indurated splinters	— fissile and soft; breaks in small chips; disturbed pieces 'tinkle'
— surfaces minimally affected by weathering	— silvery grey clay-rich weathered surfaces diagnostic, limonitized
— silica-rich: surfaces covered with characteristic green and black lichen; talus has green cast	— no preferred lichen cover, but developed soil profile

thick and resistant unit between two relatively incompetent clastic layers commonly constitutes a complete thrust slice. In the mapped area these slices dip moderately south, and north-facing cliffs show hundreds of metres of medium bedded dolosiltstone and lutite, with regular dark bands near the top.

Where underlying Quartet Group is preserved, alternating beds of brown-weathering dolosiltstone and darker shales constitute the base. At the stratigraphic top (best exposed at locality 2) are grey siltstone and brown weathering sandstone, but in many places this contact is faulted. Where the top of Gillespie Lake Group is believed nearly intact, this unit is about 1200 m thick (Mustard et al., 1990).

The Gillespie Lake Group is mapped as two divisions (Mustard et al., 1990) because lateral facies changes and structural repetition could not be consistently distinguished.

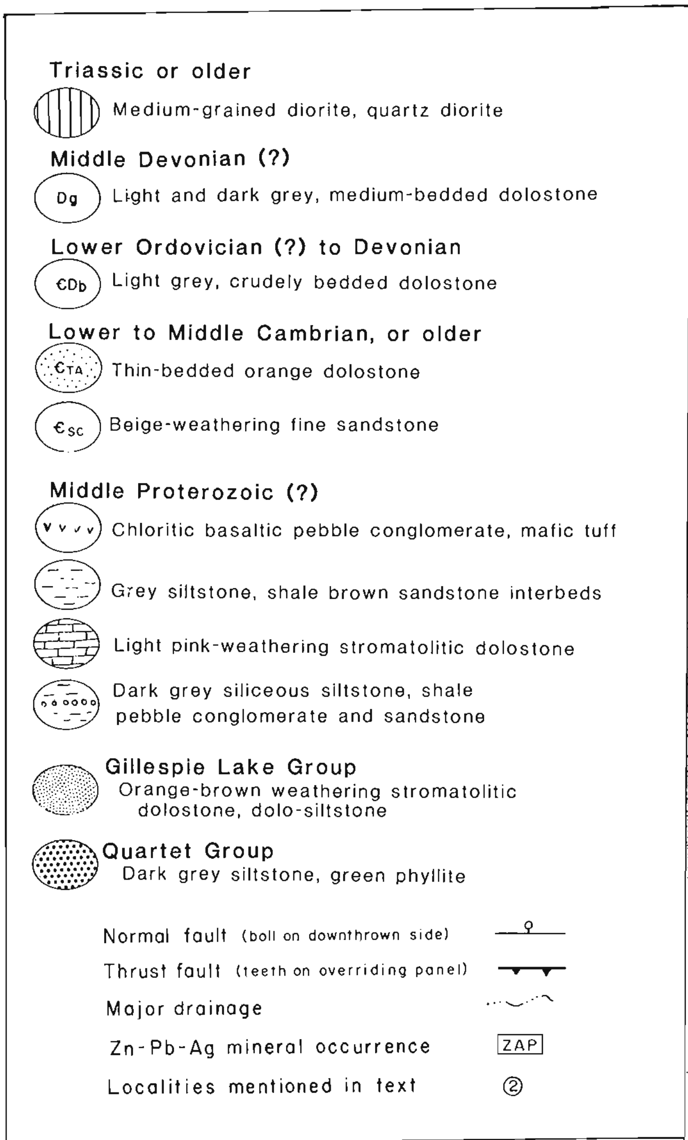
The lower division comprises numerous fining-upward successions 1-5 m thick (Fig. 4). Graded dolomitic sand- to mudstone cycles show abundant crossbeds. Near the top, lenticular rolls of dolomudstone within thinly bedded dolosiltstone predominate, and ribbed weathered surfaces have a 'radiator' texture. Tool marks and minor slump features were also noted.

The upper division consists of thickly bedded dololutite with abundant stromatolite beds. Fine sandstone layers, oolites, dissolution structures, mudcracks, and shaly intraclasts are common. Some extensive beds of columnar stromatolites are up to 4 m thick, with individual heads to 20 cm diameter. Several dark bands of silicified siltstone and mudstone provide markers that can be traced for 10 km east from Mount Good.

A thin green volcanic layer (localities 3-5), exposed over 10 km westward from Mount Good, could be a regional, although discontinuous, marker horizon. It lies near the gradational boundary of the upper and lower divisions. At Mount Good it consists of 60 cm of fine-grained chlorite, with clay amygdules whose structure suggests a former lava flow. Altered phenocrysts, originally equant and up to 3 mm across may have been augite. Occurrences farther west are up to 160 cm thick and internally layered, typically fining upward and intercalated with tan-weathering dolostone. This schistose rock may have been a mafic tuff with a reworked top.

The lower division of the Gillespie Lake Group represents distal turbidites in a deep water environment. In contrast, many features of the upper division indicate shallow water and emergent conditions. Because features representing these depositional environments can be recognized even in small and isolated outcrops, it is expected that this two part classification for the unit can be used regionally. It is, for example, effective at the Blende property in defining the geochemically anomalous zone (D. Eaton and D. Lister, pers. comm., 1989).

Although volcanic rocks are absent from the Gillespie Lake Group in the Ogilvie Mountains, several occurrences are known near the Hart River, about 125 km northwest of Mount Good. Basaltic pillowed flows, tuffs and overlying



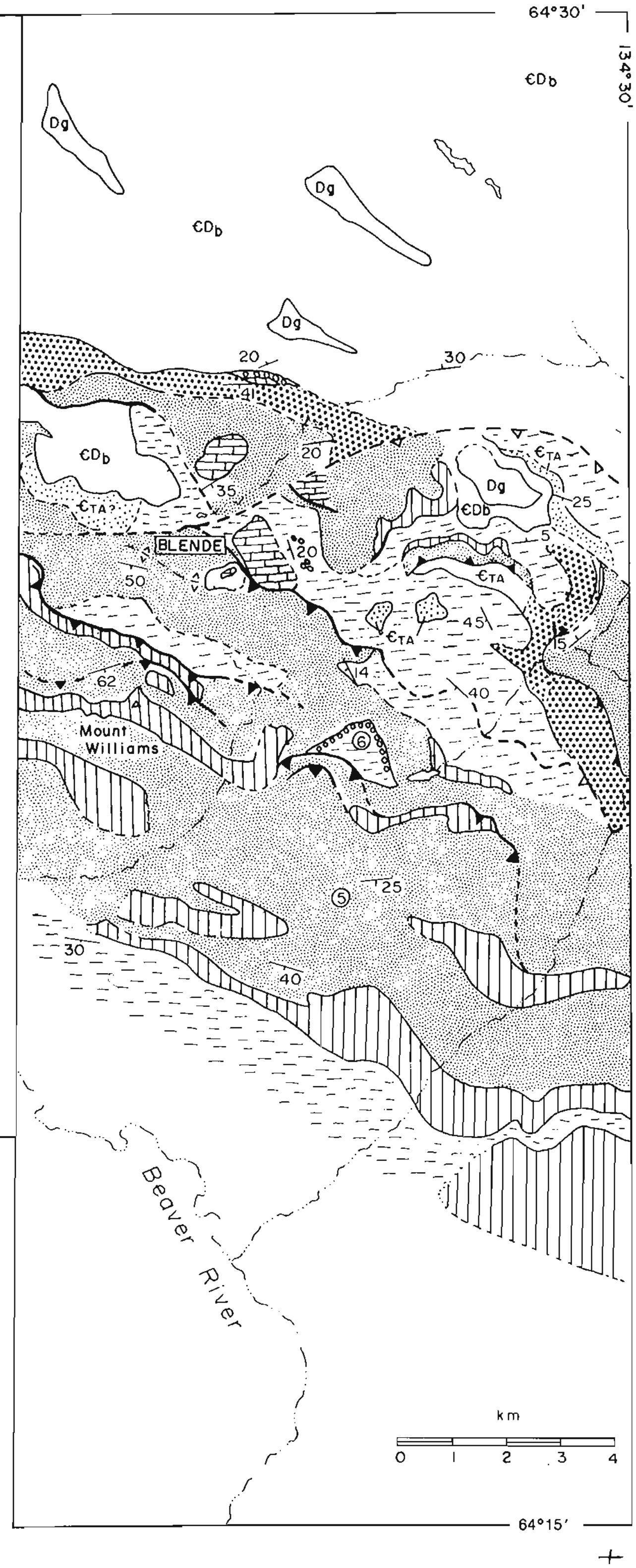
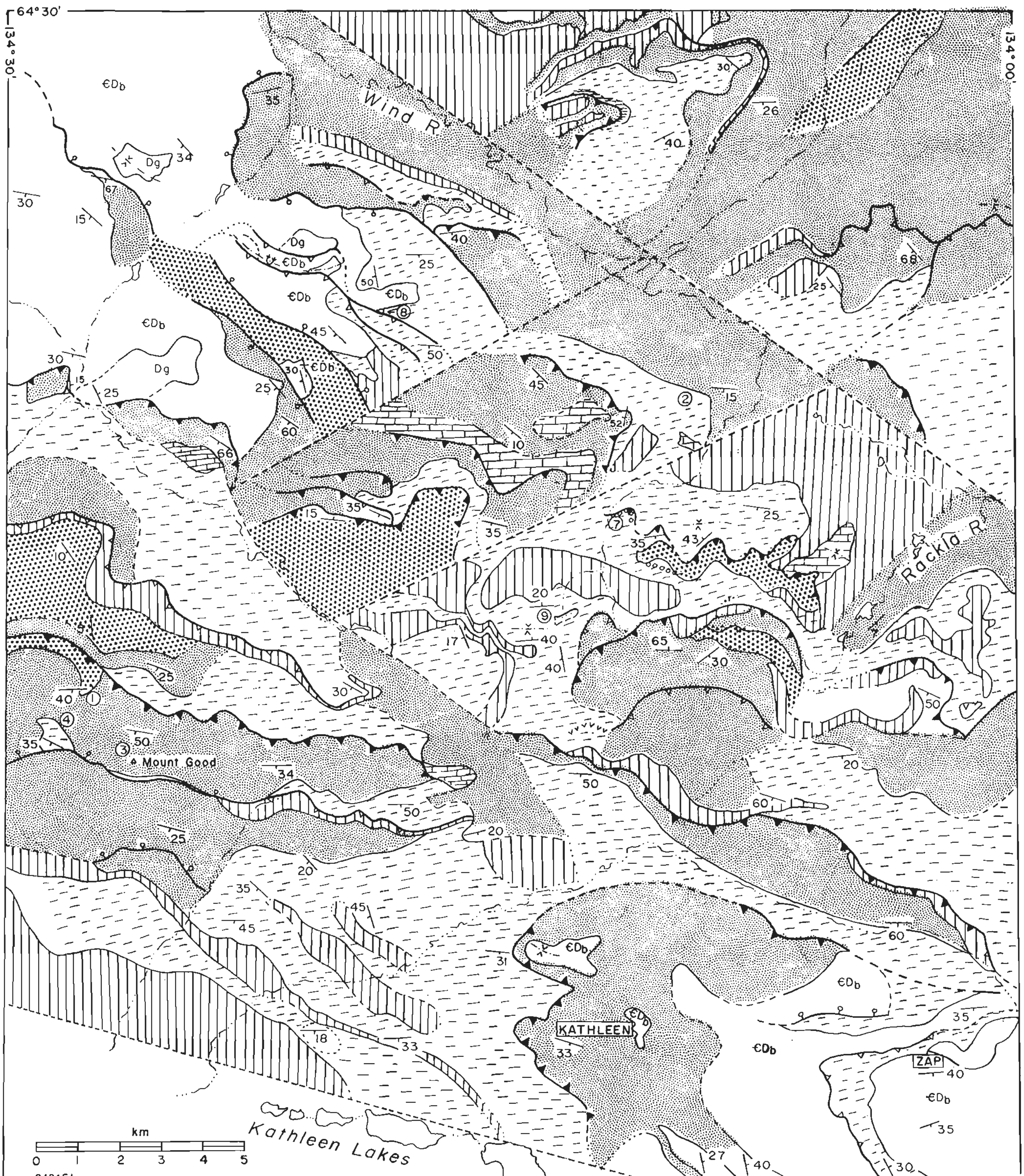


Figure 2. General geology of 106D/7 (left) and 106D/8 (right). Unpatterned area is covered by Paleozoic carbonate strata.


64°30'
134°30'
134°00'
64°15'
Wind R.
Rackla R.
Mount Good
KATHLEEN
ZAP
Kathleen Lakes
km
0 1 2 3 4 5
€Db
Dg

maroon and green shales occur at the Hart River prospect (Morin, 1979) and appear genetically related to massive greenstone sills (may include flows) that extend over 10 km (Green, 1972). A 5 m thick andesite flow was noted by Bell (1986) and some breccias in the eastern Wernecke Mountains are reported to have spatially associated mafic rocks. These additional volcanic occurrences could provide relatively fresh samples for isotopic dating the Gillespie lake Group, and so provide a minimum age for the Wernecke Supergroup.

Unit 4

Rocks between the Gillespie Lake and probable Cambrian strata include dark shales, siltstone, minor sandstone and carbonate, a widely dispersed conglomerate layer, and a single outcrop of volcanic conglomerate. No single stratigraphic succession is representative, and different rock types are exposed at the base. Deposition of unit 4 undoubtedly followed a structural disturbance, and the uneven distribution of basal strata could reflect new structural highs and formation of basins. Furthermore, the stratigraphic position of this unit encompasses at least 600 Ma, most of it as hiatuses. Direct correlation of similar lithology in isolated occurrences up to 40 km apart is not attempted.

Pebble to cobble conglomerate disconformably overlies Gillespie Lake carbonate 4 km east of Mount Williams (locality 6; see Bell, 1986) and at several localities. It is clast-supported, with well-rounded siltstone, sandstone and white quartz pebbles, but no dolostone detritus was found. The 7 m thick bed at locality 7 contains rare leucocratic granitic clasts and about 1 % jasper fragments. In some places the conglomerate is interbedded with sandy horizons with floating pebbles, and may be continuous with pebble-rich sandstone layers that pass upward into dark siltstone characteristic of the unit.

Dark, siliceous fine sandstone and siltstone overlie the upper Gillespie Lake carbonate at the Blende occurrence. This succession, which contains thin beds of fine, cross-laminated dolostone, passes upward into light-coloured platy siltstone, and is overlain by resistant dolostone with a slight pinkish weathering hue. This dolostone, also preserved at scattered localities across the central and eastern part of the area, is characterized by stromatolites different in form and texture from those of the Gillespie Lake Group (Mustard et al., 1990). Typically the algal laminae are very fine, almost hairlike, and small budding stromatolite heads occur atop large columns.

The predominant lithology in unit 4 is dark grey siltstone and shale. It is at least 150 m thick in smooth, rounded ridges in the northeast, and it underlies the broad valley north of Mount Good. The graphitic shales typically weather powdery silver-grey, and are softer and less splintered than similar lithology in the Quartet Group. Unit 4 also contains abundant khaki, brown and maroon beds, but no primary features or trace fossils could be found in them. Three localities revealed delicate cone-in-cone structures, indicative of compaction and dewatering.

Towers of volcanic conglomerate overlie siliceous mudstone along the eastern edge of the mapped area. They are composed of subrounded clasts of chloritized basalt in a chloritic matrix. Up to 6 m of dark green tuff overlies the conglomerate. No source was located, but may occur several kilometres farther east.

Both the succession of dark shales, and the fine-laminated stromatolitic carbonate match unit 4 (Thompson and Roots, 1982) in the Ogilvie Mountains, a unit informally referred to as the "Rackla Assemblage" or "Lower Fifteenmile Group". In that area, these rocks are considered to have been deposited between 1300 and 1000 Ma, and are succeeded by massive carbonate units. Unit 4 in the Mount Good-Mount Williams area, however, contrasts with the sequence and lithology of the Pinguicula Group informally described by Eisbacher (1981) in its type locality about 40 km to the northeast. As indicated by Bell (1982, 1986), "Pinguicula" has been applied by many workers to describe various successions within this stratigraphic interval, and use of the term should be discontinued until a regionally representative section is formally described.

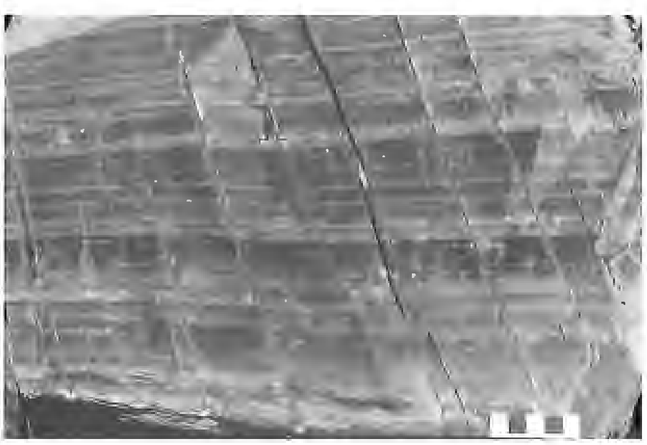


Figure 3. Graded siltstone-mudstone beds of Quartet Group. Note white laminae and cleavage diagnostic of this unit. Scale in centimetres.

Figure 4. Dolosiltstone and dolomudstone beds in the lower division of the Gillespie Lake Group, interpreted as a deep-water distal fan deposit. Scale in centimetres.

Cambrian(?) Units

Light-coloured, unmetamorphosed strata conformable with the base of the early Paleozoic dolostone were noted north of Wind River, at the Zap occurrence in the southwest, and 2 km north of the Blende property. The tan-weathering sandstone with pebbly interbeds is up to 25 m thick. Primary textures are obliterated, but a distinctive hematite banding parallel to bedding is distinctive, where the rock is not fused with silica. This unit resembles the Slats Creek Formation (Early and Middle Cambrian), described by Norris (1982) and Bell (1986) north of Nash Creek map area.

Beige-weathering, medium quartz sandstone, in finely laminated 1 m beds, occurs immediately east of the Zap barite showing. It overlies brown shale and underlies lower Paleozoic CDb dolostone. The well-rounded grains and mature composition does not resemble Proterozoic shelf units.

An orange- to grey-weathering, thin bedded limestone and dolostone occurs near the Zap prospect and 2 km northwest of the Blende occurrence. Over large areas this lithology is pervasively intruded by siderite gash-veinlets, and siderite replaces laminae 2-5 mm apart (Fig. 5), resulting in a 'zebra' texture and pinkish weathering cast. This texture is also referred to as 'presquilite' after its occurrence in the Pine Point lead-zinc district of the Northwest Territories (Aitken, 1989, p. 12). Because this alteration is spatially associated with two of the showings in the Mount Good-Mount Williams area, it may indicate favourable base metal potential.

CDb Dolostone

Light grey-weathering massive dolostone covers the Proterozoic rocks in the northeast corner of the mapped area. Its thick-bedded nature is revealed along ridge crests, but long slopes consist of blocky debris. In the southeast corner of the mapped area, erosional remnants of this unit cap the Gillespie Lake Group. In all places CDb consists of dolomite, commonly sucrosic but locally very fine-grained, with abundant cavities, and most primary textures have been obliterated by dolomitization. Crinoid fragments found at locality 10 indicate the base of the unit is Ordovician or younger.

Figure 5. Siderite veins and replacement of thinly bedded Cambrian (?) dolostone. This breccia extends hundreds of metres in an area 2 km northwest of Blende property.

This dolostone (unit 8 of Green, 1972) extends over most of northern Yukon and has many equivalent names. Because it is diachronous and represents a variety of carbonate depositional environments, the most general term, CDb from the map areas to the north (Norris, 1982) is commonly used.

Middle Devonian(?) Dolostone

Dark grey, brown and light-grey weathering dolostone caps high CDb ridges in the northwestern part of the area. In contrast to the massive CDb dolostone, this unit is distinctively banded on the scale of several metres. Beds are 30-50 cm thick, and contain fine laminae and circular structures that may be organic. From a distance, gentle and isoclinal folds of several hundred metres are distinguished.

Green (1972) mapped this as unit 10, and correlated it with the Middle Devonian Gossage Formation.

INTRUSIVE ROCKS

Resistant, brown-weathering diorite forms broad bands and rugged ridges that trend southeast across the area. Several of the largest exposures are plugs that truncate bedding, but most occurrences are sills, from 1 m to more than 70 m thick. The most continuous sills, traceable 15 km, are paired, although there are gaps, such as the ridge east of Mount Williams, where the intrusions are not exposed (a fault with alteration marks its buried location).

The age of the intrusions is unclear. Many cut unit 4 but none intrude CDb within the mapped area. Northwest of the Blende, however, similarly weathering mafic dykes crosscut folded CDb dolostone. If these intrusions are coeval, the diorite is younger than Devonian, and may be Triassic, the age of sills in the Keno Hill Quartzite (Mortensen and Thompson, in press).

Fresh rocks are typically dark green, medium-grained hornblende diorite, with intersertal texture. Plagioclase is light green, and is more susceptible to alteration than mafic minerals. Two of the larger intrusions contain up to 10 % quartz crystals to 2 mm across.

Carbonate host rocks are bleached white and dedolomitized in an envelope 5-10 m away from the diorite. The limestone is flaky, with greenish talc and scapolite(?) in this zone, but primary features, such as stromatolite shapes are well preserved. In the north-central part of the area, closely spaced sills have resulted in alteration of entire sections. The bleached Gillespie Lake rocks resemble CDb, but strata followed away from the diorite show the characteristic orange-brown weathering hue.

A small breccia pipe (50 m^2) 12.5 km northeast of Mount Good (locality 8) occurs at the base of CDb and probably intrudes siltstone of unit 4. Rusty weathering clasts of carbonate, black and hematitic mudstone and white quartz are densely packed in a fused siliceous matrix. This pipe is probably not related to the Wernecke-type breccias because it occurs at a higher stratigraphic level and contains abundant quartz. No mineralization was noted in the silicified breccia.

STRUCTURE

This area is characterized by southwest-dipping strata, steep-dipping faults and low-angle thrusts. The sequence of faults gives clues to the relative timing of deformation events.

Wide, flat-floored valleys strike northeast and southeast across the map area. They are probably underlain by late faults because they separate ridge systems revealing different stratigraphic successions and fault patterns (Fig. 6). Possibly these straight faults are splays related to the Knorr-Richardson fault array (Norris and Hopkins, 1977) which cut the fold-thrust belt 50 km northeast of the area.

Some thrust faults in areas of high topographic relief are revealed by low-angle truncation of underlying strata but most are indicated by older-over-younger strata relationships. Their extent and magnitude has not been shown on previous maps. The Gillespie Lake rocks mapped north and east of Mount Good are fault-bounded slices, each exposed as a north-facing dolostone scarp 500 m high thrust over siliciclastic rocks of unit 4. Thrusts typically follow shaly layers in unit 4 and Quartet Group; these units must be consistently identified to precisely locate the thrusts (Table 1). Other thrust faults occur at the base of diorite sills and can be identified because rocks beneath these thrusts are unaltered, in contrast to the extensive alteration envelope surrounding most intrusions.

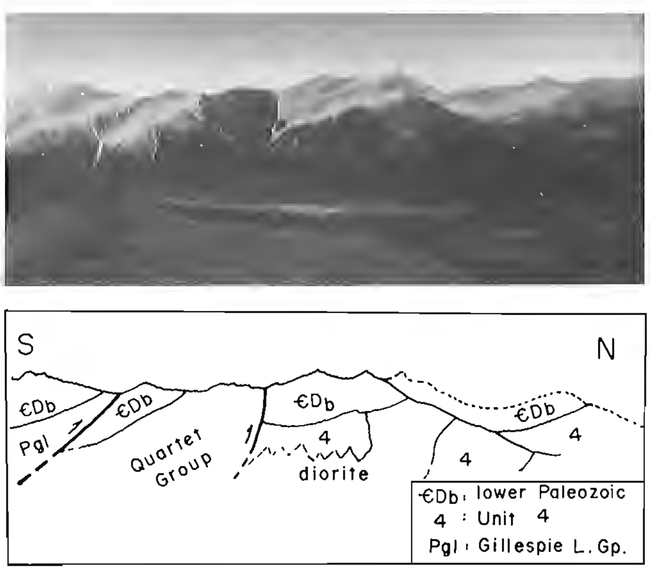


Figure 6. Different units beneath CDb dolostone (white) are uplifted on steep faults. These faults cannot be traced across the wide valley in foreground. Looking northwest at the area near locality 8 in Figure 2.

The southeast corner of the mapped area consists of Gillespie Lake dolostone overlain by erosional outliers of CDb. It appears to be a structural panel overlying the regional southeast-striking units, including unit 4. The postulated overthrust occurred after contraction of the thicker stratigraphic section; this panel also probably reflects more distant tectonic transport. Furthermore, units at a higher stratigraphic level show a different structural style. Some outliers of Devonian strata in the northwest part of the area are isoclinally folded with an amplitude of hundreds of metres, without apparent deformation in the underlying CDb. These and mountain-scale structures farther northwest, about 15 km beyond the map area, indicate that the Paleozoic "cover" has detached from older rocks studied here.

Some structures visible in the Wernecke Supergroup are not reflected in overlying rocks of Unit 4, and may have resulted from a Proterozoic deformation event. Two north-facing cirques (locality 9) expose tilted slabs of Gillespie and Quartet strata several hundred metres thick, and these are disconformably overlain by thick sections of unit 4 shales (Fig. 7). Also the range of rock units that unconformably overlie the Gillespie Lake Group could reflect differential uplift and erosion at the time of deposition.

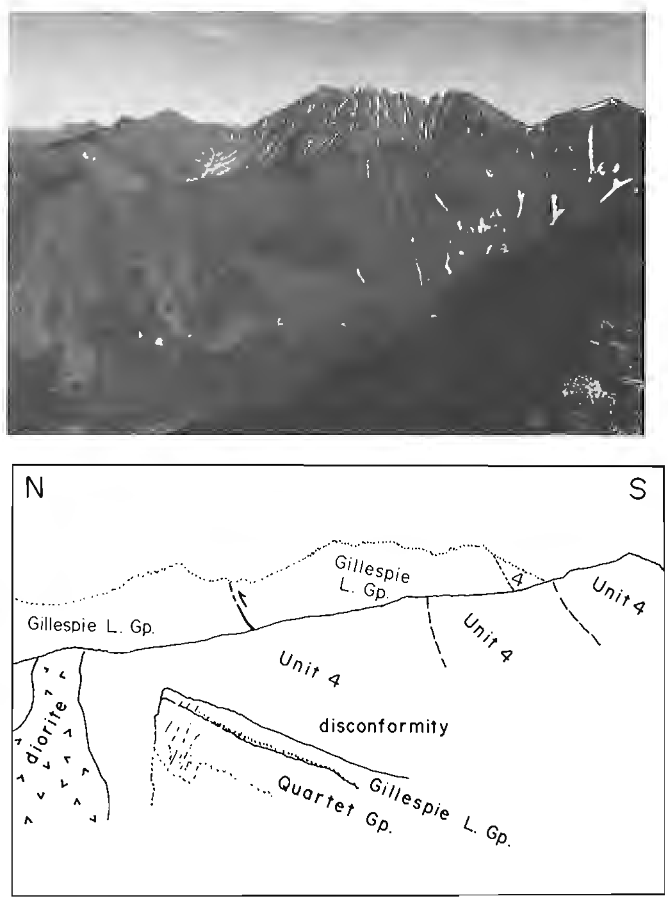


Figure 7. The resistant ramp at base of ridge consists of Gillespie and Quartet rocks. These Wernecke Supergroup rocks have been faulted before deposition of the overlying shale and siltstone on the ridge. View eastward at locality 9 in Figure 2.

The Blende property covers a faulted anticline within a thrust slice of Gillespie Lake dolostone. The anticlinal structure is not reflected by an erosional remnant of dark clastic rocks and pink-weathering stromatolitic dolostone near the axis. This remnant, forming a topographic high point, is identified as unit 4, whose deposition therefore postdated formation of the anticline. Mineralized zones have a parallel trend to the anticline (D.I.N.A., 1984, p. 155), and the overlying unit 4 rocks show no geochemical anomaly (D. Eaton, pers. comm., 1989). The age of Pb-Zn-Ag mineralization may be Middle Proterozoic, prior to deposition of unit 4. Galena from the Blende gives a Helikian Pb-isotopic age (Godwin et al., 1988).

In summary, the mapped area straddles a broad zone of structural transition that reflects the change from middle and upper Paleozoic rocks to the south, to an entire Middle Proterozoic section in the high Wernecke Mountains to the north. The repetition of the Gillespie Lake stratigraphy with siliciclastic strata of unit 4 or Quartet Group between is clear evidence for regional contraction along thrust faults. Sulphide mineralization appears concentrated along pre-Mesozoic structures, whose geometry is elucidated by regional mapping and continued exploration of the Gillespie Lake Group.

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This project received field and moral support from Steve Gordey of the GSC and Grant Abbott of D.I.N.A. The extra effort by D. Eaton, D. Lister and L. Cymbalisty of Archer Cathro and Associates, Ltd., made for healthy and happy working days. Rod Young and Andrew Shobridge gave capable assistance, and Dave Reid of Trans-North Ltd. provided exemplary service. All of us are grateful for a season of unusually fine weather.

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Preliminary results of the stratigraphy and structure of the Mt. Westman map area, central Yukon

Grant Abbott[1]

Abbott, G. Preliminary results of the stratigraphy and structure of the Mt. Westman map area, central Yukon; in Current Research, Part E, Geological Survey of Canada, Paper 90-1E, p. 15-22, 1990.

Abstract

Deep water clastics of the Upper Proterozoic and Lower Cambrian Hyland Group are juxtaposed, across the Dawson Thrust, against mid-Proterozoic to Devonian shelf carbonate rocks. The shelf and offshelf sequences are overlapped by shale with lesser amounts of quartzite, sandstone, limestone, and volcanics that previously included the Cretaceous Keno Hill Quartzite, the Jurassic "Lower Schist", and an unnamed Triassic unit. The "Lower Schist" probably includes Devonian to Jurassic strata. The Keno Hill Quartzite is probably Mississippian and laterally equivalent to parts of the "Lower Schist". As originally mapped, the Keno Hill Quartzite includes black siliceous shale and felsic volcanic rocks (host to the Marg massive sulphide deposit) and quartz grit (possibly equivalent to the Hyland Group). Lack of correlation between two parts of the "shelf sequence" across the Kathleen Lakes Fault and between the "shelf" and "offshelf" sequences across the Dawson Thrust suggest that they represent reactivated Paleozoic faults.

Résumé

Des roches clastiques d'eau profonde du groupe de Hyland du Protérozoïque supérieur et du Cambrien inférieur sont juxtaposées, de part et d'autre de la faille inverse Dawson, à des roches carbonatées de plate-forme du Protérozoïque moyen au Dévonien. Les séries de plate-forme et de la zone externe de plate-forme sont recouverts par chevauchement de schistes argileux associés à de plus faibles quantités de quarzite, de grès, de calcaire et de roches volcaniques qui renfermaient les quartzites crétacés de Keno Hill, les « schistes inférieurs » jurassiques et une unité triasique sans nom. Les « schistes inférieurs » renferment probablement des couches du Dévonien au Jurassique. Les quartzites de Keno Hill appartiennent probablement au Mississippien et représentent un équivalent latéral de certaines parties des « schistes inférieurs ». Comme ils ont été cartographiés pour la première fois, les quartzites de Keno Hill renferment des schistes argileux noirs, siliceux et des roches volcaniques felsiques (hôtes du gisement de sulfures massifs de Marg) et des grès quartzeux (probablement l'équivalent du groupe de Hyland). L'absence de corrélations entre deux parties de la « série de plate-forme » de part et d'autre de la faille Kathleen Lakes et entre les séries de plate-forme et de la zone externe de plate-forme, de part et d'autre de la faille inverse Dawson, laisse penser que ces failles représentent des failles paléozoïque réactivées.

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INTRODUCTION

Until the discovery of the Marg massive sulphide deposit in 1988, the Mt. Westman area (NTS 106D/1), located 50 km east of the nearest settlement and road at Keno City, central Yukon (Fig. 1), was notably barren of significant mineral occurrences. Host rocks to the deposit, mapped as the Keno Hill Quartzite (Green, 1972), were known to contain significant epigenetic, Ag-rich vein deposits at Keno Hill, but were thought to have little or no potential for volcanogenic or sediment-hosted base metal deposits. This project was an exceptional opportunity to not only study the regional setting of this exciting new exploration target, but also to examine some critical stratigraphic and structural relationships in the northern Cordilleran miogeocline, as the area contains most of the significant components.

This study, combined with recent studies in other areas, has resulted in significant revisions to the stratigraphy and structure of the Mt. Westman area, which was only mapped previously at reconnaissance scale by Green (1972) and, in part, by Blusson (1978). The revised interpretation of the regional setting is summarized below.

The area straddles the regional boundary between the "shelf assemblage" of the miogeocline on the north, and the "offshelf assemblage" (Selwyn Basin), on the south (Fig. 1). Middle Proterozoic to Jurassic sedimentary and minor volcanic rocks, Triassic(?) mafic sills, and Cretaceous stocks underlie the area. The "shelf assemblage" includes Middle and Upper(?) Proterozoic shallow water siliciclastic and carbonate rocks of the Wernecke Supergroup (Delaney, 1981) and Pinguicula Group (Eisbacher, 1978), and unconformably overlying Lower and Middle Paleozoic shelf carbonate rocks of Mackenzie Platform. The "off shelf" assemblage here includes Upper Proterozoic(?) to Cambrian deep water clastic rocks of the Hyland Group ("Grit Unit") and younger volcanic rocks. Both assemblages are overlain by a Devonian(?) to Jurassic shale-dominated clastic assemblage with lesser amounts of quartzite, sandstone, limestone, and volcanic rocks, that were originally divided into the Keno Hill Quartzite, "Lower Schist", and an unnamed Triassic unit of shale and limestone (Green, 1972). Until recently, the "Lower Schist" was considered to be Jurassic, and the Keno Hill Quartzite, Cretaceous (Green, 1972; Tempelman-Kluit, 1970). Recent work, including this report, indicates that the "Lower Schist" may range from Devonian to Jurassic (Mortensen and Thompson, in press; Poulton and Tempelman-Kluit, 1982; Blusson, 1978); and that the Keno Hill Quartzite is Mississippian (Mortensen and Thompson, in press; Blusson, 1978). Hornblende diorite sills of probable Triassic age (Mortensen and Thompson, in press) intrude the Lower Schist and Keno Hill Quartzite. Two monzodiorite stocks of probable Cretaceous age were encountered during the course of mapping.

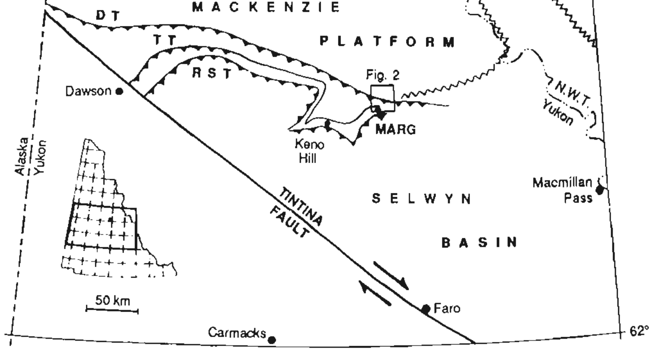


Figure 1: Tectonic setting and location of the Mt. Westman map area. DT-Dawson Thrust; TT-Tombstone Thrust; RST-Robert Service Thrust.

Two and possibly three significant but poorly documented Mesozoic thrust faults probably extend more than 200 km eastward from Dawson into the map area (Fig. 1, 2). The "Dawson Thrust", for many years an informal term and first introduced into the literature by Tempelman-Kluit (1981), juxtaposes the offshelf assemblage onto the shelf assemblage. The "Tombstone Thrust" has not yet been formally introduced, but has been suggested by R.I. Thompson (pers. comm., 1989) as a name for a thrust fault in Dawson map area which places the Mississippian Keno Hill Quartzite onto Jurassic strata of the "Lower Schist". The Robert Service Thrust, documented by Tempelman-Kluit (1970) in Dawson map area, places the Hyland Group onto Jurassic and older strata, and passes about 1 km south of the map area.

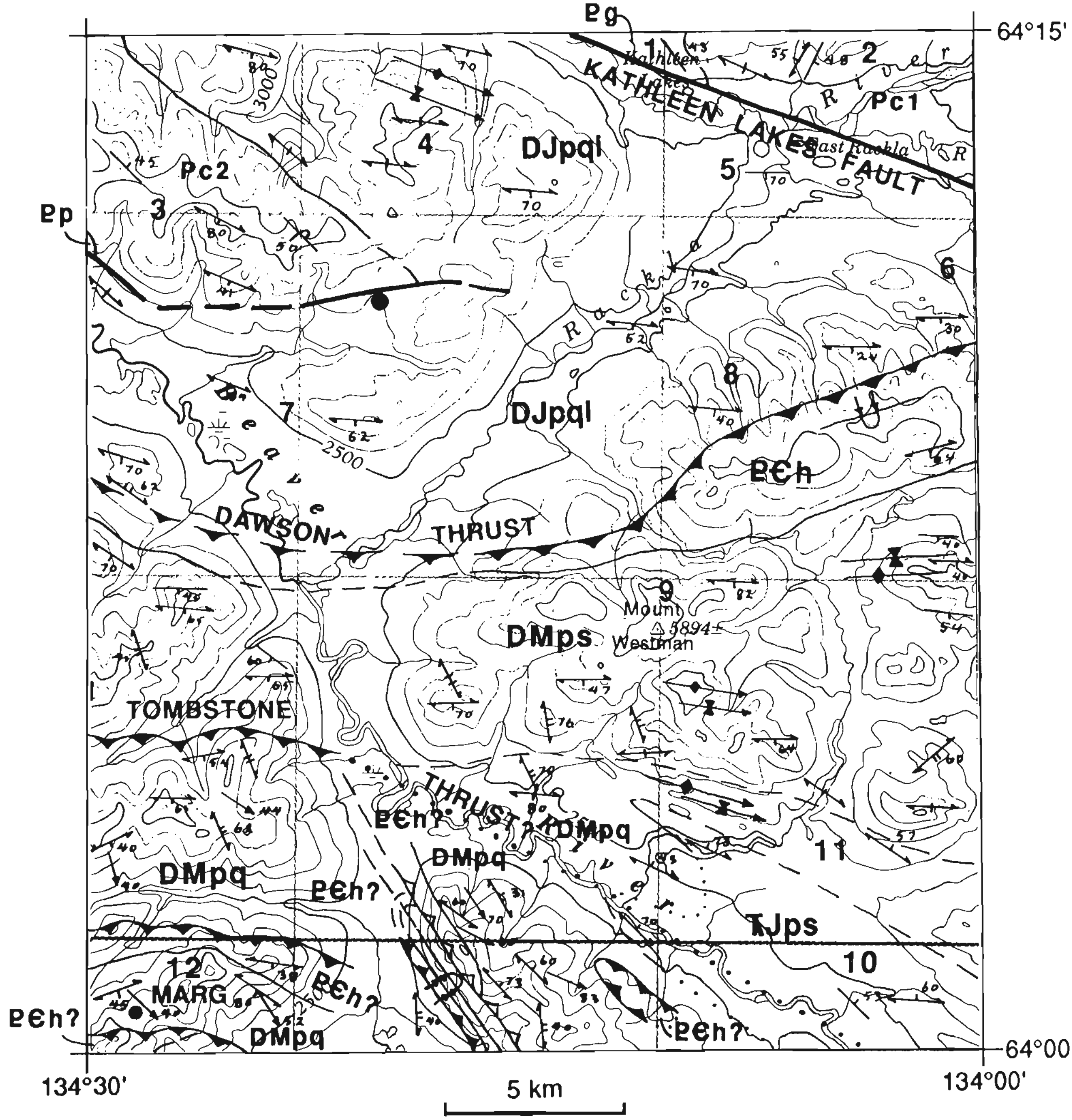
STRATIGRAPHY

The general geology of the area is shown in Figure 2 and the stratigraphy represented in 12 generalized stratigraphic columns (Fig. 3). Complex structure, and lack of stratigraphic markers, and fossil and radiometric age control have made stratigraphic correlations and estimations of thicknesses difficult, particularly in the Devonian to Triassic sequence. As a result, each column is a first order approximation of the stratigraphy of the area where it is located, not a measured section.

Shelf Sequence (Pg, Pp, Pc1, Pc2)

The shelf sequence includes two dissimilar carbonate sequences. North of Kathleen Lakes (Columns 1, 2), orange weathering, greenish grey, thinly laminated to thinly bedded silty limestone, dolomite and buff weathering dolomitic shale (Unit Pg) of the Middle Proterozoic Gillespie Lake Group (uppermost Wernecke Supergroup) is unconformably overlain by Lower and Middle(?) Paleozoic carbonate rocks (Unit Pc1) (C.F. Roots, pers. comm., 1989). Thick, massive white limestone at the base of the Paleozoic sequence may be as old as Early Cambrian. Separated by a covered interval but higher in the sequence, is platy, thinly laminated grey and black dolomite, overlain by massive, medium grey- to pinkish-grey weathering dolomite. These rocks can be no younger than Middle Devonian.

In contrast, southwest of Kathleen Lakes (Column 3), Paleozoic carbonates (Pc2) clearly overlie the Pinguicula Group (Pp) (C.F. Roots, pers. comm., 1989), which com-



LEGEND

TRIASSIC AND JURASSIC

TJps Calcareous black shale, micaceous, calcareous, siltstone and sansdtone, grey, noncalcareous shale

DEVONIAN TO JURASSIC

DJpql Black shale, chert, quartzite, chert conglomerate, argillaceous limestone, grey crinoidal limestone

DEVONIAN(?), MISSISSIPPIAN, AND YOUNGER(?)

DMps Dark grey and rusty weathering black phyllite, thinly laminated to thick bedded micaceous sandstone and quartzite

DMpq Massive quartzite interbedded with black phyllite

LATE PRECAMBRIAN AND LOWER CAMBRIAN

PЄh Hyland Group: Quartz feldspar grit, quartzite, dark grey and green chloritic phyllite, buff and grey weathering limestone, sandy limestone, maroon and green shale

Lithologic contact; defined approximate, assumed

Thrust fault; D1,D2,defined, approximate

S1 foliation

L1 mineral lineation, rodding

F1 fold axis

S2 cleavage

F2 fold axis

Figure 2: a) Simplified geological map of the Mt. Westman area (106D/1). Numbers show location of stratigraphic columns in Figure 3.

prise orange to buff weathering cryptalgal laminated dolomite overlain by a thin interval of rusty-brown weathering, olive green siltstone and shale. The Paleozoic carbonates consist of three main divisions. Monotonous, massive to oncolitic and locally sandy, light grey weathering coarse grained dolomite of the lower division is overlain by recessive grey to buff weathering silty limestone and massive to thick bedded white limestone of the middle division. Two or more mafic volcanic flows and volcaniclastic conglomerate and sandstone up to 50 m thick at and near the base of the middle division probably correlate with Ordovician volcanic rocks mapped to the northwest by Green (1972). The silty limestone grades upward into pale yellow weathering well bedded limestone. The upper division is primarily massive to well bedded light grey weathering dolomite and limestone. At the top, dark to medium grey weathering, fetid limestone contains "two hole" and "star" crinoids, indicating a Middle Devonian age for the top of the sequence. The Paleozoic carbonates were originally interpreted (Green, 1972) to overlie the "Grit Unit" (Hyland Group) southwest of Kathleen Lakes. No Hyland Group strata were seen north of the Dawson Thrust during the course of this mapping.

Off Shelf Sequence (PCh)

The uppermost Proterozoic and Lower Cambrian Hyland Group (Unit PCh) and possibly overlying undated mafic volcanic rocks comprise the off shelf sequence (column 9). Lower and Middle Paleozoic rocks present elsewhere in Selwyn Basin are presumed to have been eroded beneath the unconformity at the base of the "Lower Schist".

The three characteristic subdivisions of the Hyland Group (quartz feldspar grit, quartzite and phyllite; limestone; and maroon and green shales) are exposed in a narrow band above the Dawson thrust. The grit division includes buff weathering quartz sandstone, both calcareous and noncalcareous, massive grey to brown weathering quartzite, and interbedded buff weathering grey phyllite. The top of the grit division contains several bands of orange weathering sandy limestone and massive grey limestone between 10-80 m thick. Maroon and green shale, interbedded with some quartz grit and quartzite, form the upper division. Only the lower division is preserved west of Mt. Westman.

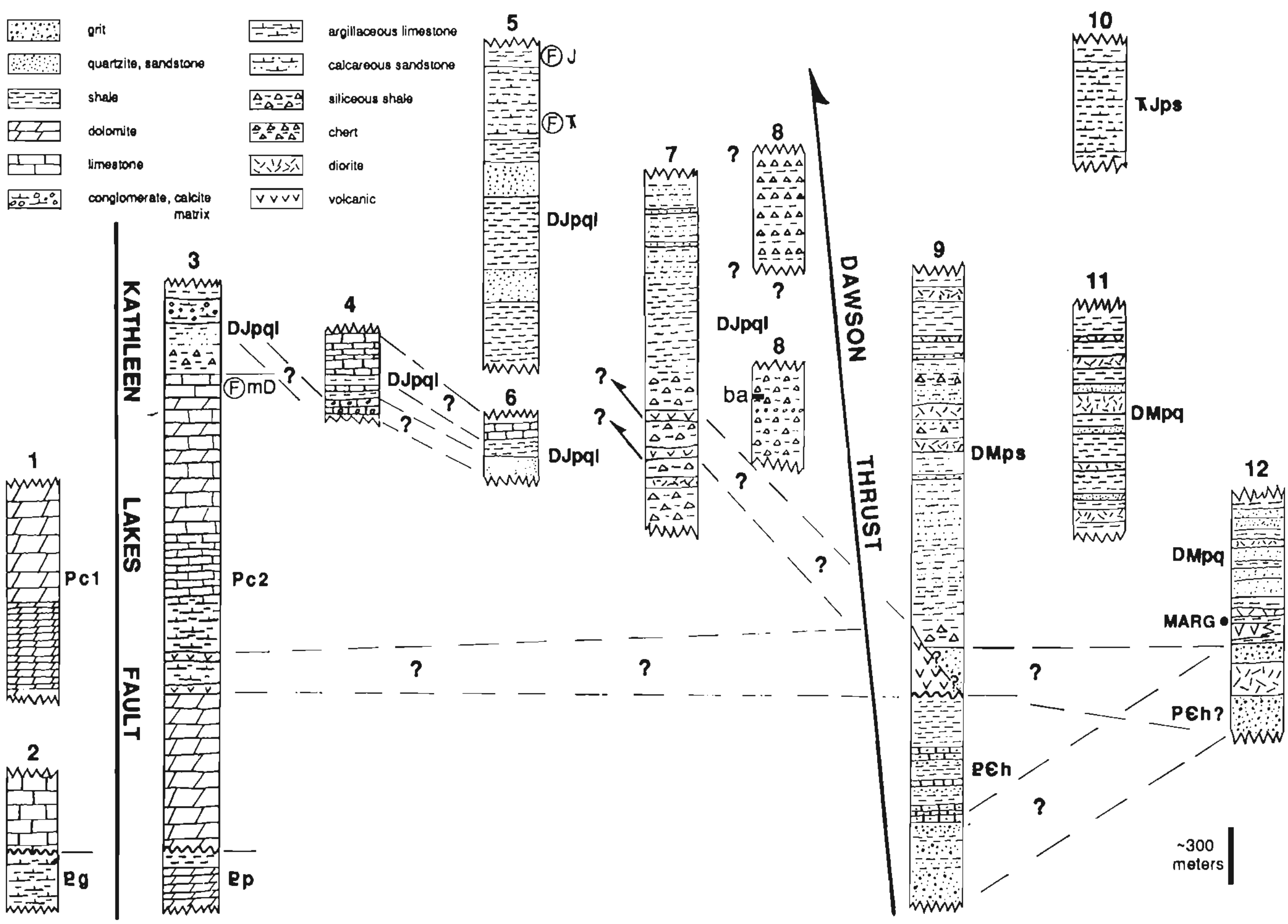


Figure 3. Generalized stratigraphic columns showing the main stratigraphic relationships in the Mt. Westman Area.

In the southwest corner of the map area, buff weathering quartz grit, which appears to form two or more thrust repetitions in Devono-Mississippian shale and quartzite is tentatively correlated with the lower division of the Hyland Group (Turner and Abbott, 1990). The grit forms homogenous sequences up to 200 m thick, with intermittent thin grey and buff weathering marble bands up to several metres thick near the top. Blue and grey quartz grains up to 0.5 cm are supported in a buff to pale green siliceous matrix. An alternative correlation for this unit is discussed below.

Dark green, mafic lapilli tuff up to 25 m thick overlies the maroon and green shale in two narrow imbrications along the Dawson Thrust north of Mt. Westman. Between Mt. Westman and the Beaver River, homogenous, dark green crystal and lapilli tuff at least 300 m thick (its base is not exposed) underlies the "Lower Schist "along the Dawson Thrust. The tuff is massive, blocky, brown and grey weathering and pale green on fresh surfaces. Feldspar and uncommonly pyroxene grains supported in a chlorite matrix display rare sedimentary textures, and porphyritic lapilli in a matrix of similar composition are rarely visible on weathered surfaces. These tuffs resemble Ordovician volcanic rocks which overlie maroon and green shale of the Hyland group in other parts of northwestern Selwyn Basin (Roots, 1988). They could also be as young as Devonian.

Devonian to Jurassic strata north of the Dawson thrust (DJpql)

The Devonian to Jurassic assemblage north of the Dawson thrust contains mainly black to dark grey to blue shale and slate, but also contains conglomerate, quartzite, mafic volcanic rocks, and two limestone members (columns 3-8).

Dark grey weathering slate and siltstone, and silver blue weathering siliceous shale, typical of the Devono-Mississippian Earn Group (Gordey et al., 1982) overlies Middle Devonian limestone north of the Rackla River (column 3). The shale contains an unusual chert pebble conglomerate, in which buff weathering, well rounded, well sorted chert clasts about 1 cm across are cemented in a coarse grained grey limestone matrix. The conglomerate appears to grade laterally into both a silica-cemented chert conglomerate, and into limestone containing scattered chert clasts. The conglomerate appears to form the base of a carbonate sequence (column 4), perhaps 200 m thick which comprises three members, each of roughly equal thickness. The lower consists of beds of limestone up to 2 m thick containing chert clasts; the middle member, limestone in 2 m thick beds, interbedded with dark grey shale in beds of equal thickness; and an upper member, massive to thick bedded crinoidal limestone. East of the Rackla River (column 6), poorly exposed quartzite, shale, and limestone may be equivalent to the three members of the carbonate sequence in column 4.

Immediately south of Kathleen Lakes (column 5) quartzite forms two units up to 100 m thick in a thick sequence of black slate. The quartzite forms medium to thick, sooty, argillaceous beds interbedded with varying amounts of black slate. These rocks grade eastward into massive grey vitreous quartzite, typical of the Keno Hill Quartzite, and form exposures about 50 m thick along the Rackla River. Also along the Rackla River, calcareous shale interbedded with black sooty limestone containing Triassic fossils (Green, 1972), and overlying dark grey slate containing Jurassic fossils (Poulton and Tempelman-Kluit, 1982), were previously thought to be older than the quartzite. The Triassic limestone can also be interpreted to overlie the quartzite, as suggested by recent Mississippian age determinations for the Keno Hill Quartzite in Dawson map area (Mortensen and Thompson, 1989).

If Triassic limestone overlies quartzite and shale along the Rackla River, the limestone, quartzite and conglomerate assemblage shown in columns 3, 4, and 6 is probably older, and quartzite probably forms three or more separate horizons. Near Macmillan Pass (Gordey et al., 1982), three separate Mississippian quartzite horizons in a thick shale and chert sequence are known. There, the shale also undergoes an abrupt northward facies change to Mississippian limestone. Although Tempelman-Kluit (1982) correlated the limestone in column 4 with the Permian Takhandit Formation, a Mississippian age is preferred.

Poorly exposed strata along the lower reaches of the Rackla River (column 7), originally mapped by Green as the "Grit Unit", appear to comprise a lower member of blue to rusty weathering black siliceous shale similar to, but much thicker than that seen in column 3, a middle member of dark grey weathering, dark grey slate with conspicuous rusty rhombs of weathered iron carbonate, and an upper member of thin bedded to thinly laminated quartz siltstone and quartzite interbedded with slate similar to that in the middle member. Quartzite forms both massive exposures up to several metres across, and more commonly, thin beds and laminations. Correlation of these quartzites with the discrete, massive quartzite in columns 5 and 6 is unknown, but possible, judging from the abrupt facies changes apparent between columns 3-6.

Much of the area south of the Rackla River (column 8) is underlain by strata typical of the Earn Group, including blue-grey siliceous shale and chert, and a few beds of chert pebble conglomerate. Several occurrences of bedded witherite or baritocalcite up to 5 m thick, were noted. Dark, chocolate-brown to grey weathering, dark grey to greenish-grey slate, interbedded with grey chert appears to overlie the blue shales, but their stratigraphic position relative to the quartzite and other strata is unknown. These rocks resemble Late Paleozoic strata from other parts of Selwyn Basin.

Massive, vesicular, and locally pillowed basalt flows and tuffs (column 7) as thick as 200 m are intercalated(?) with black siliceous shale in a narrow belt that trends east from the Beaver River, south of Unit Pc2, across the map area. The basalt could be either interbedded with, or thrust onto the black shale. In the latter case, similar rocks which unconformably(?) overlie the Hyland Group south of the Dawson thrust could be equivalent.

Devonian(?) to Jurassic(?) clastic rocks south of the Dawson Thrust (DMps, DMpq, TrJps)

South of the Dawson Thrust (Fig. 2), the clastic assemblage unconformably overlying the Hyland Group includes three units (DMpq, Dmps, and TJps) which together resemble unit DJpql north of the Thrust, but which differ in detail.

Unit DMps (column 9), mapped as Lower Schist by Green, characteristically comprises dark grey weathering black to dark grey slate, with varying amounts of monotonous, rhythmically interbedded, thinly laminated to thick bedded, buff to grey weathering micaceous siltstone, sandstone, and quartzite. The sandstone and quartzite are typically ripple cross-laminated or plane parallel-laminated. Silver-blue weathering, black siliceous shale, containing a few beds of chert grit are intermittently interbedded with the slate. North of Mt. Westman, in the northernmost splay of the Dawson thrust, blue siliceous shale contains bedded witherite or baritocalcite near the base of the unit.

A crude stratigraphic succession is apparent along the northern margin of the belt along the contact with the underlying Hyland Group, but elsewhere the succession appears to be obscured by folds and thrust faults. West of the Beaver River, the base of the unit includes black sooty quartz grit and a thin lens of orange weathering, pale green felsic volcanic rocks which resemble those near the Marg deposit (Turner and Abbott, 1990). Northeast of Mt. Westman, buff- to greenish-grey weathering quartz grit about 200 m thick marks the base of the unit. This grit resembles grit in the Hyland Group but is more homogenous, and contains angular clasts of black shale. Blue-quartz grit mapped as the Hyland Group in the southern parts of the map area is also similar and the two units could be equivalent. Blue weathering siliceous shale, although uncommon everywhere, is more abundant near the base of the unit. East of the Beaver River, quartzite seems to become cleaner and more abundant going upsection, to a point where it predominates over a 30-50 m interval, and includes massive beds up to 2-3 m thick that resemble the quartzite seen north of the Dawson Thrust. Shale with minor siltstone interbeds overlies the quartzite.

Unit DMpq (columns 11 and 12) includes strata mapped by Green as the "Keno Hill Quartzite" in the southwest corner of the map area, and as the "Lower Schist" in the poorly exposed southeast corner. The unit is mainly dark grey slate interbedded with thick sequences of massive quartzite. North of the Beaver River (column 11), tightly folded, massive grey weathering quartzite 1-30 m thick is interbedded with poorly exposed dark grey slate and thinly laminated quartz siltstone. A similar sequence south of the Beaver River includes blue weathering siliceous shale, and a few thin lenses of green, chloritic tuff. The thickness and abundance of quartzite is significantly greater in the southwest corner of the map area near the Marg deposit (column 12), where quartzite predominates over shale and forms thick massive sequences up to 100 m thick in a complex, tectonically interleaved assemblage that also includes a significant component of buff and green weathering tuffaceous or volcanic rocks (Turner and Abbott, 1990).

Unit DMpq could either be laterally equivalent to unit DMps or overlie it. The first possibility is preferred. North of the Beaver River (column 11), the thick quartzites of unit DMpq appear to be folded and in vertical contact with dark grey slate thought to be equivalent to the top of unit DMps (column 9). They could underlie, or be thrust onto that slate, and therefore be equivalent to the relatively thin quartzite horizon in column 9. West of the Beaver River, the contact between units DMps and DMpq is shown as the Tombstone Thrust(?). Strata on both sides of this contact dip moderately south. The similarity of grit and volcanic rocks in unit DMpq in this area to parts of the Hyland group and/or strata at the base of unit DMps farther north suggest a thrust. The main difference between units Dmps and Dmpq is the thickness and amount of quartzite. The abrupt facies changes apparent in the quartzite north of the Dawson Thrust suggests that the facies changes required for the two units to be equivalent are possible.

Constraints on the ages of units DMpq and Dmps are few. The blue weathering siliceous shale, and dark grey shale common to both units are typical of the Devono-Mississippian Earn Group in other parts of Selwyn Basin. Even more characteristic of the Earn Group is the chert conglomerate and witherite in unit DMps. The abundance of micaceous siltstone, sandstone and quartzite, and absence of thick sequences of chert conglomerate are uncharacteristic of the Earn Group. Perhaps the Hyland Group was the source of clastic detritus in unit DMps, rather than the Road River Group which elsewhere was the source of the characteristic chert pebble conglomerate. A Mississippian age for the quartzite of unit Dmps is indicated not only by a Mississippian fossil age for the Keno Hill Quartzite obtained by Thompson (Mortensen and Thompson, 1989) in Dawson map area to the west, but also by similar fossil ages obtained east of the area by Abbott from quartzite near Macmillan Pass and by Gordey (in press) in Nahanni map area. If unit Dmpq overlies unit Dmps, the latter, must be Mississippian or older. If unit Dmpq is laterally equivalent to unit Dmps, parts of the latter could be younger than Mississippian. Hornblende diorite sills that are probably Triassic (Mortensen and Thompson, 1989), apparently intrude all parts of both units, constraining their youngest possible age.

In the southeast corner of the map area (column 10), recessive, black shale, buff weathering calcareous, micaceous siltstone and minor ripple cross-laminated and plane parallel laminated calcareous sandstone, and argillaceous limestone appear to overlie unit DMpq, and are probably Triassic in age. Although somewhat different from the Triassic limestone and calcareous shale north of the Dawson thrust, these rocks strongly resemble Triassic strata elsewhere in central Yukon. Recessive, silvery-buff to grey weathering noncalcareous grey shale is either interbedded with, or overlies the Triassic siltstones, and may be Jurassic.

Hornblende Diorite (Trd)

Numerous hornblende diorite sills from 1-100 m thick intrude units DMpq, DMps, and DJpql, but not unit TrJps, or middle Paleozoic or older strata. Only one sill was seen north of the Dawson Thrust (column 7), but to the south, the sills, in places, comprise more than 50 % of the rock volume and many have been traced for several kilometres. A few dykes cut quartzite beds at a shallow angle, but no sharply crosscutting relationships were observed. Many of the numerous orange gossans and kill zones in the district are attributable to weathering of disseminated pyrite in the diorite and/or pyrrhotite and pyrite in hornfels enclosing the sills.

Mafic sills intrude the Keno Hill Quartzite and Lower Schist in a continuous belt that has been traced by Green (1972) for 200 km to the west where, in the Tombstone Range, Mortensen and Thompson (1989) obtained a U-Pb zircon-baddeleyite age of 232.2 +1.5/-1.2 Ma (latest Middle Triassic). The sills in the Mt. Westman area are also probably Triassic. If so, Units DMps and Dmpq must be older than Middle Triassic, and not Jurassic as originally mapped by Green (1972).

STRUCTURE

Two phases of penetrative deformation have been recognized. The earliest (D1) affects all Jurassic and older rocks, throughout the map area, whereas the younger (D2) has only been recognized in the southern portions.

In most of the area, D1 is characterized by an east to east-southeast trending steep to moderately south dipping cleavage (S1) that is accompanied by occasional small scale, gently to steeply southeast plunging isoclinal folds. From north to south, the plunge of fold axes tends to steepen. An intersection lineation defined by bedding and cleavage is oriented parallel to fold axes, but is generally poorly defined south of the Dawson Thrust, where most bedding is nearly parallel to foliation. South of the Beaver River and the Tombstone Thrust(?), grit quartzite and diorite display an intense penetrative rodding and mineral lineation that plunges moderately southeast, in the plane of S1 foliation, and parallel to F1 fold axes (Turner and Abbott, 1990; Gordey, 1990). Discrete mylonite zones between a millimetre and 2 m with a similar orientation are also developed. The mineral lineation indicates a northwest direction of transport, unlike the north to northeast direction attributed to most structures in east-central Yukon.

On either side of the Beaver River, the style and intensity of deformation differ markedly in the same apparent stratigraphic units. A northwesterly directed thrust sheet that is now eroded away or represented by the klippe shown in Figure 2 may have overridden the sheared rocks south of the river, but not the less deformed rocks to the northeast. Possibly this thrust is the Tombstone Thrust, or alternatively the Robert Service Thrust or one of the imbricates between the two.

Several thrust faults, probably related to late stages of D1 deformation, have been mapped southwest of the Beaver River on the basis of repetitions of quartz grit correlated with the Hyland Group. Some evidence, cited earlier, exists for the Tombstone thrust west of the Beaver River, but along strike to the east, quartzite and diorite sills appear to be deformed into tight upright folds, and could dip north beneath shales of unit DMps.

Younger structures (D2) include south-southeast trending, steeply northeast dipping crenulation cleavage, accompanied by tight small scale folds with southeast plunging axes. D2 structures are most intense in the southwest corner of the map area, and gradually disappear to the northeast.

The only recognized large structures related to D2 deformation are a high angle southwesterly directed reverse fault (open barbed fault in Fig. 2), south of the Beaver River that is accompanied by a syncline that is overturned to the west. S1 foliation and compositional layering above the fault dip to the northeast, rather than to the south, which is typical. The thrusts mapped near the Marg Property and possibly the Tombstone Thrust(?) appear to be cut by this fault and cannot be traced across it to the east. This structure and the Beaver River farther north mark the abrupt boundary between the thick quartzites on the west (column 12) and the thinner sequence on the east (column 11). Perhaps the thick quartzites on the west side are at a higher structural level than those to the east, in a thrust slice, possibly the Tombstone, that has been uplifted and eroded away east of the D2 reverse fault. This possibility would explain why the Tombstone Thrust cannot be traced east of the Beaver River.

Farther north, but south of the Dawson Thrust, diorite dykes in Unit DMps and limestone in the Hyland Group define tight to isoclinal, upright to steeply south-dipping F1 folds with amplitudes of at least 100 m.

Large scale structures north of the Dawson Thrust are poorly defined. Tight upright fold are defined by quartzite south of Kathleen Lakes. The large normal fault marking the southern limit of Paleozoic carbonate rocks west of the Rackla River may be related to uplift about a large, nearby hornfels, associated with a buried Cretaceous intrusion. The Kathleen Lakes Fault juxtaposes strata as old as the mid-Proterozoic Gillespie Group on the north side of Kathleen Lakes against the Mississippian "Keno Hill Quartzite" on the south. Whether movement is normal or reversed to the south, and the age of the fault relative to other structures are unknown.

SIGNIFICANCE OF THE DAWSON THRUST AND THE KATHLEEN LAKES FAULT

The lack of correlation of Middle Paleozoic and older strata, and the rough correlation of Devono(?)-Mississippian units across the Dawson Thrust indicate that it is a reactivated, mid-Paleozoic or older fault. The lack of correlation of Devonian and older carbonate rocks across the Kathleen Lakes Fault suggests that is also has a Paleozoic history. Two possibilities are episodic rifting and Devono-Mississippian strike-slip movement.

Tempelman-Kluit (1981) first suggested this possibility, and speculated that it represented the northern boundary of a long-lived rift system that began in early Paleozoic time and produced Selwyn Basin. The lack of correlation of Devonian and older carbonate rocks across the Kathleen Lakes Fault suggest that it also has a Paleozoic history. Although fossil control is minimal, the results of this study suggest that no stratigraphic links exist across either of these faults in Middle Devonian or older rocks. Nor have any been documented for the entire length of the Dawson Thrust. If the Dawson Thrust reflects long-lived Paleozoic rifting as suggested by Tempelman-Kluit, surely some stratigraphic ties would be evident, particularly in Proterozoic strata, even after telescoping of facies during Mesozoic shortening.

Large scale strike-slip offset in the order of several hundred kilometres is another possibility. The stratigraphy and sedimentology of the Devono-Mississippian Earn Group also suggests such a possibility (Abbott, 1987; Abbott et al., 1986; Gordey et al., 1982), as does the probable Devono-Mississippian age of much of the Lower Schist. However, movement of this magnitude would require a fault that can be traced for much of the length of the northern Cordillera. Such a fault, which should be traceable for much of the length of the northern Cordillera, is not yet obvious beyond the Dawson Thrust. The possibility of large-scale Devono-Mississippian strike-slip movement in the northern Cordillera is therefore speculative at this time.

SUGGESTIONS FOR MINERAL EXPLORATION

With the discovery of the Marg Deposit, the Keno Hill Quartzite, carbonaceous siliceous shale and felsic volcanic assemblage represents a new exploration target in central Yukon. The Keno Hill Quartzite and equivalent strata have been traced intermittently from Dawson to Nahanni map area (Green, 1972; Abbott, 1983; and Gordey, in press). Although the genetic relationship, if any between the quartzite and sulphides is unknown; the quartzite, being thicker and more extensive than the volcanic rocks should serve as a useful exploration guide.

Strata mapped as the Jurassic Lower Schist may be in part equivalent to the Devono-Mississippian Earn Group and in part to the Keno Hill Quartzite and should be explored for both sedimentary exhalative zinc lead silver deposits such as the Tom and Jason at Macmillan Pass and for volcanogenic deposits such as the Marg.

In the Mt. Westman area, and possibly farther west in Nash area, strata equivalent to the Earn Group and Keno Hill Quartzite were previously mapped as the "Grit Unit" and may have been overlooked in previous exploration.

The creek draining the Marg deposit contains a gossan and kill zone; gossans and kill zones are common throughout Units Dmpq and Dmps. Some are clearly related to the weathering of sulphides in diorite dykes or their hornfelsed aureoles, but others have no apparent explanation and cannot be disregarded.

ACKNOWLEDGMENTS

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Geology and mineral potential, Tiny Island Lake map area, Yukon

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Abstract

The Tiny Island Lake area (105M/16) is underlain by three main stratigraphic assemblages that form the hanging wall of the Robert Service thrust. The oldest, of late Proterozoic-Cambrian age, consists of slate, quartzite, and minor limestone of the Hyland Group. This is overlain unconformably by slate, chert, chert pebble conglomerate and felsic volcanics of probable Devono-Mississippian age. The youngest unit consists of Triassic-Jurassic slate, sandstone and carbonate.

These strata are complexly folded under conditions of low metamorphic grade, the complexity of deformation increasing with proximity to the Robert Service thrust. The Devono-Mississippian assemblage is unexplored relative to its potential for stratiform Pb-Zn-Ag and volcanogenic massive sulphide deposits.

Résumé

Le sous-sol de la région du lac Tiny Island est constitué de trois principales séries stratigraphiques qui constituent la lèvre supérieure de la faille inverse de Robert Service. La série la plus ancienne, d'âge protézoroïque supérieur-cambrien, est constituée d'ardoises, de quartzites et de faibles quantités de calcaires du groupe de Hyland. Cette série est recouverte en discordance par des ardoises, des cherts, des conglomérats à galets de chert et de roches volcaniques d'âge dévono-mississippien probable. L'unité la plus jeune est constituée d'ardoises, de grès et de roches carbonatées du Trias-Jurassique.

Ces couches sont plissées d'une façon complexe dans des conditions de faible métamorphisme, la complexité des plis augmentant avec la proximité de la faille inverse de Robert Service. La série dévono-mississippienne n'est pas explorée quant à son potentiel de gisements de Pb-Zn-Ag stratiformes et de sulfures massifs d'origine volcanique.

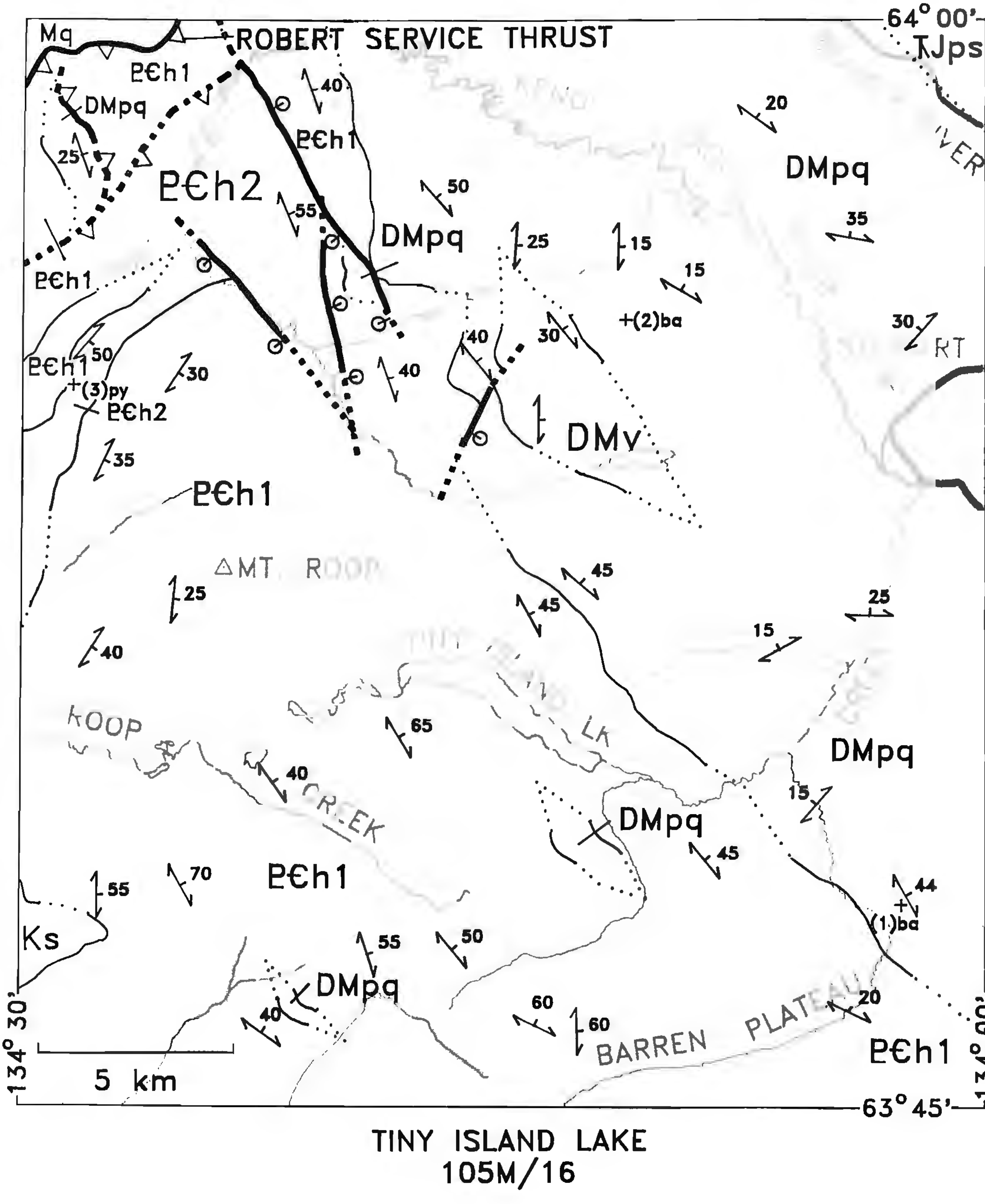

Figure 1. Simplified geological map of the Tiny Island Lake map area (105M/16).

INTRODUCTION

The map area is located 90 km northeast of Mayo, Yukon (see Fig. 1, Abbott et al., 1990). The only previous geological work was by Bostock (1948), who mapped this region at 1:250 000 scale as part of his reconnaissance mapping of the larger Mayo map area (NTS 105M). Green (1971) conducted 1:50 000 scale mapping in the Mayo Lake region that adjoins the study area on the west. Adjacent regions to the north and east have been mapped at 1:250 000 by Blusson (1974; 105N and 106C) and Green (1972; 106D).

Most of the map area lies within the hanging wall of the Robert Service thrust (Fig. 2, Abbott et al., 1990). Strata range in age from Late Proterozoic to Triassic and consist dominantly of sandstone, slate, conglomerate, chert, felsic volcanics and minor limestone (Fig. 1). Jura-Cretaceous deformation was intense, as exemplified by recumbent folds and penetrative planar and linear fabrics. These obscure bedding and other primary features and render impossible the determination of original stratigraphic thicknesses. Metamorphism is low greenschist facies. A small mid-Cretaceous granitic intrusion occurs in the southwest part of the area.

STRATIGRAPHY

The oldest exposed strata are those of the Hyland Group (Gordey, in press; "Grit Unit" of Green, 1972) which underlie much of the western part of the area (Fig. 1). Two mappable units are recognized. In the northwest, dark brown-weathering, black graphitic slate and minor thin bedded, fine grained, dark grey quartz sandstone (PCh2) form a thick member structurally (and stratigraphically(?)) low within the Hyland Group. The upper and more extensive part of the group (PCh1) consists of sandstone, phyllite and minor carbonate. Sandstone and phyllite are of equal abundance, the former occurring as beds from less than a metre to up to 20 m thick. The sandstone ranges from very fine grained to granule sized and is typically bimodal. Large grains of quartz, commonly blue in colour, are surrounded by a fine grained, recrystallized quartzose matrix. Fresh plagioclase and orthoclase constitute less than a few per cent. The amount of matrix carbonate is variable; in some beds sand grains float in a calcareous matrix. The coarsest sandstone occurs at Mt. Roop, where quartz grains reach 13 mm across. Phyllite is pale green to grey on fresh and weathered surfaces, although dark grey to black varieties occur locally. Distinctive purple and green banded phyllite, in intervals from a few to up to 50 m in structural thickness, are restricted to the south part of the area. Relatively pure carbonate is rare within the Hyland Group. Members range from less than a metre to as much as 50 m thick and consist of white to grey, fine to coarse crystalline limestone. The lack of continuity and erratic distribution of the carbonate may result from structural disruption.

Gritty quartzose sandstone and slate of the Hyland Group are widespread in central Yukon. Based on regional correlation, a Late Proterozoic-earliest Cambrian age for those strata in the Tiny Island Lake area is likely (Fritz et al., 1983). The member of black graphitic slate corresponds to the "Upper Schist Division" mapped by Green (1971) in the Mayo Lake area immediately west.

Overlying the Hyland Group is a succession dominated by dark bluish-grey to light grey weathering, jet-black siliceous slate and chert, within which are members of quartzite, chert conglomerate and chert sandstone (DMpq, Fig. 1). The quartzite is fine grained, dark grey to orange weathering, and medium to dark grey on fresh surface. It forms massive unfoliated members up to 30 m thick whose contacts with the enclosing slate are sharp. Members of chert pebble conglomerate are found near the base of the succession and reach 40(?) m in thickness. The conglomerate consists of flattened clasts of dark grey to black chert(?) and argillite to 10 cm in width and less than a centimetre thick. Large, undeformed, equant, well rounded scattered clasts of fine grained quartz sandstone, up to 15 cm across, make up to 10 per cent of the rock. Not typically associated with the conglomerate, members of chert sandstone occur locally in association with dark brown-weathering, medium grey to black slate. The black slate and chert, the chert-bearing clastic rocks and occurrences of barite (described below), are typical of the Earn Group, a regionally extensive unit of Late Devonian and earliest Mississippian age (Gordey et al., 1982; Gordey, 1988). The quartzite members may correlate with mid- to late Mississippian quartzite (e.g. Keno Hill Quartzite) that is also widespread in central Yukon (Green, 1971; Tempelman-Kluit, 1970; Abbott, 1990).

A large unconformity is suspected below the black siliceous slate unit (DMpq) where it rests above the Hyland Group. Chert and argillite of Cambrian to mid-Devonian age that are diagnostic of Selwyn Basin elsewhere (see Abbott et al., 1990, Fig. 2) are absent along this contact. The large quartzite clasts within the conglomerate of the upper unit (DMpq) resemble quartzite of the Hyland Group,

Legend

TRIASSIC AND(?) JURASSIC

- TJps shale, sandstone, limestone

CRETACEOUS

- Ks biotite quartz monzonite

MISSISSIPPIAN

- Mq quartzite, slate

DEVONIAN AND MISSISSIPPIAN

- DMv rhyolite
- DMps slate, chert, quartzite, chert sandstone, conglomerate

LATE PROTEROZOIC TO EARLY CAMBRIAN

- PЄh 1: sandstone, slate, limestone; 2: graphitic slate, sandstone

contact (defined or approx., assumed)

thrust fault (teeth on upper plate)

normal fault (ball on down-thrown side)

average foliation (S1)

mineral occurrence (barite-ba; pyrite-py) (1)ba from which they were likely eroded. Chert and argillite clasts in the conglomerate may derive from the now-missing lower Paleozoic stratigraphy. Unfortunately, the angular relationships of bedding below and above the unconformity are obscured by deformation. The contact may be an angular unconformity or disconformity, but critical data are lacking.

Orange-weathering felsic volcanic rocks form a thick, homogeneous unit (DMv) within black siliceous slate (DMpq) southwest of Keno Ladue River. A strong foliation obscures their volcanic origin and imparts the false appearance of schistose quartz sandstone. Thin section examination, however, reveals volcanic textures for the quartz, features such as large monocrystalline grains, well-preserved euhedral grain boundaries and embayment textures. The rocks are composed of up to 20 per cent quartz and minor feldspar phenocrysts up to 2 mm across in a recrystallized fine-grained matrix. The strong foliation is defined by seams of white mica. In some specimens phenocrysts are rare. Whether the rocks include felsic flows, tuffs or both is uncertain. A Late Devonian or Mississippian age for the volcanics is assumed, based on the most likely age of enclosing strata (DMpq).

The youngest(?) strata within the map area comprise dark brown to light grey weathering slate and dark grey finely crystalline limestone (TJps). These occur as rare infolds(?) within the Devono-Mississippian slate and also underlie the northeast corner of the map area. A Triassic and (?)Jurassic age is likely because shale, siltstone, and limestone like that of this unit typify Upper Triassic strata elsewhere in central Yukon (Gordey and Irwin, 1987).

The succession in the footwall of the Robert Service thrust, in the northwest corner of the map area, is dominated by fine grained, dark grey to black quartzite, and interbedded black, carbonaceous phyllite (Mq). The quartzite occurs in beds from a few centimetres thick to units more than 20 m thick and ranges from clean to impure. The thicker units are typically massive, fine grained, clean quartzite that may have a parting ranging from 10 to several metres in thickness. These quartzites differ from those in the hanging wall of the Robert Service thrust (DMpq) in having more intimately admixed pelite. This unit (Mq) has been mapped regionally as the Keno Hill Quartzite (Green, 1971), and traced westward from the Mayo area nearly to Dawson, Yukon. Conodonts recently recovered from the Keno Hill Quartzite near Dawson indicate a Mississippian (late Visean) age (M.J. Orchard, pers. comm., 1989; Abbott, 1990).

INTRUSIVE ROCKS

In the southwest corner of the map area (Fig. 1) a small body of unfoliated, medium grained, quartz monzonite intrudes the Hyland Group. Fresh biotite is the chief mafic mineral and makes up about 10 per cent of the rock. Hornblende occurs rarely as coarse chloritized grains. Scattered potassium feldspar megacrysts to 2 cm in diameter constitute less than 5 per cent. Surrounding the intrusion is an andalusite-bearing hornfels up to 500 m in width. This pluton is part of the mid-Cretaceous Selwyn Plutonic Suite, composed of numerous plutons of similar type, many of which have yielded mid-Cretaceous ages (Anderson, 1983).

The only other intrusive rocks in the area include rare scattered sills up to 60 m across of altered fine to coarse grained pyroxene gabbro. Their age is uncertain, although the strongly foliated margins of these bodies indicate emplacement before Jura-Cretaceous regional deformation. In Tiny Island Lake map area, they were only observed within strata of the Hyland Group. However, to the north abundant and correlative(?) greenstone bodies of Triassic age intrude the Mississippian Keno Hill quartzite (Green, 1971; Abbott, 1990).

STRUCTURE

Most of the Tiny Island Lake area lies in the immediate hanging wall of one of the largest thrust sheets in the Cordillera, the Robert Service thrust, that cuts through the northwestern part of the map area (Fig. 1). This fault can be traced westward 200 km towards Dawson, Yukon (Fig. 2, Abbott et al., 1990). Its easterly and southerly continuation are uncertain. Along its length it emplaces Upper Proterozoic-Cambrian strata of the Hyland Group against the Mississippian Keno Hill quartzite. In the Tiny Island Lake area, the intensity of deformation increases structurally and stratigraphically downward towards the trace of the Robert Service thrust (Fig. 1). Mesoscopic folds throughout the map area reach about 10 m in amplitude. Larger structures may be present, but their recognition is hindered by poor exposure and lack of marker horizons.

At high structural levels in the east and south part of the area the dominant fabric is a moderate dipping penetrative cleavage (S1, Fig. 1, 2a). Rarely, parts of limbs of large recumbent folds in bedding are observed whose axial planes are concordant to this fabric (Fig. 3a). Rare measured fold axes and cleavage-bedding intersection lineations plunge shallowly to the east and east-northeast.

At lower structural levels and best displayed in Hyland Group strata northwest of Mount Roop, deformation is more complex and additional fabric elements are developed. Gently southeast-plunging recumbent folds that fold the aforementioned cleavage (and bedding) are ubiquitous (Fig. 2b, 3b). Interlimb angles generally range from 60 to 90°. Axial planes trend east to northeast and dip moderately to the south. A variably developed axial plane cleavage (S2, Fig. 2b) crosscuts the pre-existing cleavage (S1) and bedding (S2). Lower structural levels are also characterized by a strong rodding or stretching lineation (L2, Fig. 2b) that is coaxial with the recumbent folds. It is best displayed in competent strata such as coarse gritty sandstone of the Hyland Group, and in spectacular stretched conglomerate (unit DMpq) in the northwest corner of the map area. Phyllitic strata display a coaxial crenulation lineation.

Superimposed on all the above fabric elements are broad upright buckles to local, tight, upright, or northeast-verging folds. Although broadly coaxial with the fold axes and stretching lineation described above, they locally fold, and therefore postdate, these older structures.

Structures in the footwall of the Robert Service thrust are described by Turner and Abbott (1990), and Abbott (1990).

North to northwest-trending normal faults of large stratigraphic throw offset map-unit contacts in the northwest part

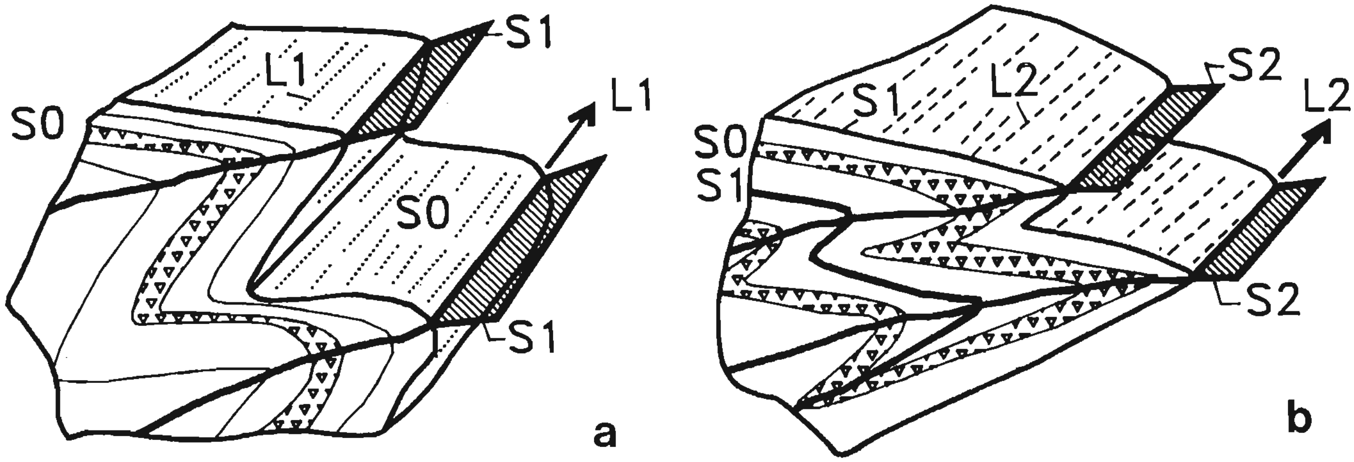

Figure 2. Diagrammatic sketch of megascopic fold style showing main fabric elements. Scale of folds varies up to 10 m in amplitude.
(a) fold style at high structural levels. Folded layering is bedding (S0). A penetrative shallow to moderate dipping cleavage (S1), parallel to the fold axial planes, is the dominant planar fabric in the map area. L1 represents cleavage-bedding intersection lineation and fold hinge-lines.
(b) fold style at low structural levels. Bedding (S0) and a penetrative cleavage (S1) are folded about shallow to moderate dipping axial planes (S2). Locally an axial plane cleavage is developed parallel to S2. L2 represents fold hinge-lines, as well as a rodding or stretching lineation in competent lithologies, and a crenulation lineation in phyllitic strata.

of the area (Fig. 1). Their traces relative to topography indicate near vertical dips. The southeast extent of these faults may be greater than portrayed (Fig. 1) but there are no markers within Hyland Group strata (PCh) with which to gauge offset. Whether these faults represent tear faults confined to the Robert Service sheet, or whether they cut through the sheet and extend northward into the adjacent area is uncertain (Abbott, 1990).

The southeast-plunging stretching lineations and coaxial recumbent folds in the hanging wall of the Robert Service thrust reflect ductile deformation during low-grade metamorphism. The trend of the lineation suggests northwest transport of the thrust sheet during the formation of these fabrics. The intensity of strain diminished upward within the thrust sheet, leading to concurrent(?) development of the simpler fold geometry found at higher levels. The late folds reflect minor amounts of north (?) directed shortening related to final emplacement of the sheet or to post-emplacement shortening. The northwest transport direction inferred from the stretching lineation differs from the north to northeast transport assumed from the general shortening direction of the Cordilleran thrust and fold belt.

METAMORPHISM

The rocks in the area have undergone chlorite zone greenschist facies metamorphism, as indicated by minerals defining cleavage. Cleavage in Hyland Group sandstone, the felsic volcanics and locally chert conglomerate, is defined by fine to medium grained muscovite and chlorite that wraps around primary mineral grains and lends the rock a slightly phyllitic appearance. In originally fine grained strata the micas are very fine grained and cleavage surfaces are smooth and phyllitic. Deformation and metamorphism were probably synchronous because the metamorphic micas are deformed, and because formation of a stretching lineation and local stretched conglomerate indicate ductile deformation mechanisms that require elevated temperature.

Quartz veins, the product of low grade metamorphism and accompanying strain, are ubiquitous at lower structural levels in the Robert Service thrust sheet and diminish in quantity upwards. They are typically a few centimetres or less in width, although pods of bull quartz up to 10-15 cm across are not uncommon. They vary in deformation style from planar and crosscutting to tightly folded, indicating they were likely emplaced during deformation.

GUIDE TO MINERAL EXPLORATION

The eastern part of the map area has high mineral potential despite the lack of exploration it has received. Strata which underlie this area (DMpq) largely correlate with the Earn Group, which elsewhere hosts significant stratiform Pb-Zn-Ag-barite occurrences (Tom, Jason at Macmillan Pass (Bailes et al., 1986; McClay and Bidwell, 1986)). The felsic volcanics (DMv) provide a target for volcanogenic massive sulphide mineralization. Mineralization at the Marg property, although in the footwall of the Robert Service thrust, is hosted in Mississippian felsic volcanics (Turner and Abbott, 1990). The volcanogenic massive sulphide MM deposit in the Pelly Mountains, Yukon is also hosted in felsic volcanics of Mississippian age (Mortensen and Godwin, 1982). Regional geochemical reconnaissance (Hornbrooke and Friske, 1988) has outlined coincident stream sediment anomalies for Hg, Zn, Ba, Cd, Cu and Ag in this area. Two occurrences of stratiform barite discovered during the present work indicate Devono-Mississippian(?) hydrothermal activity (see below).

Hyland Group strata do not hold as much exploration promise as the Devono-Mississippian succession. They are

Figure 3. Photographs of megascopic folds.
(a) siliceous slate of unit DMpq. The folded layering is bedding. Preservation of bedding such as this in unit DMpq is rare. S1 axial plane cleavage, not well displayed on the vertical joint face in this photo, is nevertheless strongly developed (see Fig. 2a).
(b) fine-grained quartz sandstone of the Hyland Group (PCh1). The folded layering is not bedding, but S1 cleavage (see Fig. 2b).

relatively barren of significant mineral occurrences elsewhere in central Yukon, and stream sediment geochemical anomalies for most elements are singularly lacking (Hornbrooke and Friske, 1988).

Three mineral occurrences were found during the present work (Fig. 1). Occurrence (1) (within unit DMpq) consists of relatively pure laminated barite at least 4 m thick that extends along strike at least 100 m. Lack of exposure prevents delimiting its top, base or lateral extent. No associated sulphides were observed. At occurrence (2), country rock of black siliceous slate (DMpq) grades sharply upward to pyritic chloritic phyllite, about 2 m thick, which in turn grades sharply upward to at least 2-3 m of rock composed of barite+quartz+pyrite. The lateral and vertical extent of this baritic horizon are not constrained by outcrop. At occurrence (3), a foliated, fine grained rock consisting of quartz, barite, pyrite, and carbonate occurs as a lens about 3 m thick and at least 25 m long within black, graphitic slate of the Hyland Group (PCh2). Pyrite occurs as both disseminated grains and concentrated in laminae up to 1 cm thick in the "top" 10 cm of the horizon. Other sulphide minerals were not noted.

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Regional setting, structure, and zonation of the Marg volcanogenic massive sulphide deposit, Yukon

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Abstract

The Marg Fe-Zn-Pb-Cu-Ag-Au volcanogenic massive sulphide deposit occurs in a south-dipping imbricate thrust array of black siliceous phyllite, quartz-sericite-chlorite schist, and quartzite of Early Mississippian age. The sulphide body is in a northeast-verging isoclinal fold above a thrust fault (D1) deformed by northwest-trending upright folds and cleavage (D2), and younger faults. The sulphide body and underlying quartz phenocryst-bearing metavolcanic rocks are zoned about a linear paleo-vent complex characterized by a ferroan carbonate-rich massive pyrite with high Cu/Pb and Zn/Pb ratios and 'footwall' ferroan carbonate-quartz-sericite-pyrite schists. Away from the paleo-vent complex, the sulphide body is transitional to quartz-rich massive pyrite and an outermost pyrite body. Major sulphides are pyrite, sphalerite, galena and chalcopyrite with minor tetrahedrite and arsenopyrite. Altered volcanic rocks away from the paleo-vent are sericite-quartz or chlorite-quartz schists.

Résumé

Le gisement de sulfures massifs de Fe-Zn-Pb-Cu-Ag d'origine volcanique, de Marg, se trouve dans une série de chevauchements imbriqués, inclinés vers le sud, constituée de phyllite siliceuse noire, de schiste à quartz-séricite-chlorite et de quartzite du Mississippien inférieur. Le massif de sulfures se trouve dans un pli isoclinal de vergence nord-est, au-dessus d'une faille inverse (D1) déformée par des plis et clivages droits (D2) et des failles plus jeunes. Le massif de sulfures et les roches métavolcaniques sous-jacentes à phénocristaux de quartz forment une zone autour d'un complexe linéaire de paléo-cheminées, caractérisé par une pyrite massive riche en carbonates ferrifères, ayant des rapports élevés de Cu/Pb et de Zn/Pb et des schistes dans le mur à carbonates ferrifères-quartz-séricite-pyrite. En s'éloignant du complexe de paléo-cheminées, le massif de sulfures passe progressivement à une pyrite massive riche en quartz et un massif de pyrite à la périphérie. Les principaux sulfures sont la pyrite, la sphalérite, la galène et la chalcopyrite associées à des quantités mineures de tétraédrite et d'arsénopyrite. Les roches volcaniques altérées, situées loin des paléo-cheminées sont des schistes à séricite-quartz ou à chlorite-quartz.

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INTRODUCTION

The Marg deposit is 42 km northeast of Keno City, Yukon (64° 01'N, 134° 28'W; NTS 106D/1) at 1400 m elevation near the southern margin of the Patterson Range (see Abbott et al., 1990, Fig. 1). The Marg property was first staked in 1965 to cover a geochemical anomaly from a GSC regional geochemical survey (Gleeson, 1965a,b). Drilling in 1988 by Archer Cathro and Associates (1981) Ltd for joint venture partners NDU Resources Ltd. and Cameco - A Canadian Mining & Energy Corporation has delineated indicated and inferred reserves of 1 922 000 tonnes grading 1.97 % Cu, 5.19 % Zn, 2.72 % Pb, 1.97 opt Ag, and 0.03 opt Au making it the largest volcanogenic massive sulphide deposit discovered to date in the Yukon (Cathro, 1988). Drilling in 1989 intersected extensions of the mineralized zones beyond the reserve blocks.

The purpose of this study is to establish a preliminary model for the structure and stratigraphy of the Marg deposit and evaluate evidence for zonation within the sulphide body and associated altered rocks. This work complements 1:50 000 scale mapping underway in this and adjacent areas. Turner spent 8 days on the property. Five drill holes in section 2510E at the eastern end of the sulphide zone were logged for structural relationships, ore facies and alteration types. Cursory evaluation of sulphide facies and alteration character in 23 other intersections in the eastern portion of the deposit established a preliminary zonation within the sulphide horizon and underlying altered rocks. Abbott spent 6 days on the property in 1988 and 4 days in 1989 as part of his mapping of map sheet 106D/1 at 1:50 000 scale (Abbott, 1990).

REGIONAL SETTING

In the southern Patterson Range, a convex arc of south- to southeast-dipping strata about 8 km across, originally mapped by Green (1972) as the Keno Hill Quartzite, are imbricated along northerly directed thrust faults (Fig.1). Three main thrust sheets, here referred to as the northern, central , and southern panels have been recognized between the Robert Service Thrust to the south and the Tombstone Thrust(?) to the north. Each thrust panel contains quartzite, black siliceous phyllite, quartz grit and felsic volcanic rocks, with the Marg massive sulphide deposit in the central panel (Fig. 2). Intense shearing and penetrative deformation, numerous smaller thrusts, and folds disrupt the stratigraphic order within panels. The lack of fossils and only one radiometric age further complicate efforts to reconstruct stratigraphy.

REGIONAL STRATIGRAPHY

Hyland Group (PCh)

The Hyland Group (PCh), above the Robert Service Thrust, comprises mainly massive, grey weathering calcarous and noncalcareous blue and grey quartz (and minor feldspar) grit, interbedded with dark grey phyllite, dark green chloritic phyllite and thin buff weathering limestone. Similar buff weathering siliceous quartz grit, each about 200 to 300 m thick, form the lowest parts of the central and southern panels. Ubiquitous grains of blue and grey quartz up to 0.5 cm across occur scattered in a buff, siliceous matrix. Rounded to angular grains of quartz and minor feldspar grit occur in a fine-grained matrix of recrystallized, monocrystalline quartz, lesser muscovite, and minor feldspar and carbonate. Depositional structures are not obvious and may have been obliterated by deformation. Buff and grey weathering marble forms beds up to several metres thick near the top of the unit. The possiblility of a Devonian-Mississippian age for these rocks is discussed by Abbott (1990).

Phyllite and Volcanic Rocks (DMvs)

In the central panel, the Hyland Group (PCh) is overlain by recessive, black, carbonaceous, siliceous phyllite interbedded with thin quartzite laminations, and lenses of rusty, buff or pale green weathering quartz-muscovite and quartz chlorite phyllite up to 30 m thick (DMvs). An infold or thrust repetition of these rocks within quartzite (Mq) of the central panel hosts the Marg deposit. In the southern panel, black phyllite above the Hyland Group (PCh) is either a few metres thick or absent, but strata like unit DMvs are interbedded with, or tectonically interleaved, with intervals of quartzite. In the northern panel, black phyllite and interbedded quartzite, as well as buff and green tuffaceous (?) phyllite form a thick, structurally complex assemblage in which the stratigraphic order is unknown. Quartz-muscovite and chlorite-muscovite phyllite contain quartz "eyes" . These rocks appear to be of volcanic origin, perhaps tuffs although no primary volcanic textures were observed. Igneous zircons in similar quartz muscovite phyllite from drill core at the Marg deposit have yielded an Early Mississippian radiometric age (J. Mortensen, pers. comm.).

Quartzite (Mq)

Massive, resistant, dark grey weathering vitreous quartzite (Mq)is the dominant rock type in the thrust belt and forms massive units up to 70 m thick. The quartzite lacks marker units and its internal stratigraphy and true thickness are not known. Bed-like partings in the quartzite that appear to be shear planes or mylonite zones obscure the nature and thickness of bedding. Well exposed areas reveal that "beds" of quartzite and phyllite cannot be traced far without structural dismemberment. The quartzite is generally interleaved or interbedded with black siliceous phyllite, and less commonly with buff to pale green tuffaceous phyllite. In the north parcel, the quartzite unit includes interbanded laminated quartz siltstone, phyllite and quartzite units 10 cm to 1m thick.

Black Phyllite (DMps)

Dark grey and rusty weathering phyllite and siliceous siltstone (DMps) forms a homogenous sequence about 500 m thick at the top of the northern panel where it overlies a thick, massive quartzite. Similar strata occur in the footwall of the Tombstone Thrust. The phyllite resembles the black phyllite in units DMvs and Mq, but lacks thick quartzite beds or buff and green tuffaceous phyllite.

Hornblende Diorite (Trd)

Massive, resistant, dark grey-green weathering hornblende diorite forms tectonically dismembered lenses up to 200 m thick, and 3 km long. The diorite is typically medium grained, equigranular, and composed of chlorite after hornblende and plagioclase. Some finer grained massive chloritic greenstone is also included with this unit. Ferroan carbonate alteration is locally intense.

REGIONAL STRUCTURE

Two phases of penetrative deformation are recognized. The first phase (D1) is characterized by rodding and an intense mineral lineation (L1), recumbent isoclinal folds (F1) that fold bedding on a small scale, and a strong axial planar foliation (S1). The second phase (D2) is a steeply dipping, spaced cleavage (S2), accompanied by upright, tight to isoclinal small scale folds (F2).

A pervasive, moderately south to southeast dipping foliation (S1) is the dominant fabric in phyllite and volcanic rocks (Fig. 4c, 4d). Marble and quartzite display small to medium scale recumbent isoclinal folds with axial planes parallel to S1 and fold axes parallel to L1. Diorite sills are structurally dismembered near the Marg deposit but in other parts of 106D/1 deformation is less intense and diorite bodies are continuous for many kilometres. L1 is best developed in quartzite (Fig. 4a) as moderate to strong rodding and in diorite as a mineral lineation that trends southeast with a moderate plunge. S1 is weakly developed in these rocks. Quartzite and diorite also contain widely spaced partings that in quartzite give a first impression of bedding, but instead are shear surfaces or thin mylonite zones. These surfaces cut lineations (L1) at very acute angles and contain lineations that are subparallel to the pervasive lineation. Most beds cannot be traced far, and appear to be dismembered by these shears.

The Robert Service and other large thrusts roughly parallel S1. The orientation of L1 mineral lineations are consistent with northwesterly movement on the thrusts, at right angles to the north to northeasterly direction of transport attributed to most structures in the eastern parts of the Cordilleran orogen in north-central Yukon.

Second Phase (D2) structures consist primarily of small scale, upright tight to isoclinal folds (Fig.4b, 4g) accompanied by a weakly developed axial plane cleavage. Cleavage consistently strikes about 160° and dips steeply northeast. Fold axes of deformed S1 surfaces plunge moderately to the southeast. The broad arc formed by map units and D1 structures may reflect deformation, but otherwise large scale D2 structures have not been recognized.

STRUCTURE OF THE MARG DEPOSIT

The Marg deposit occurs in a 4 km long east-trending fault repetition or recumbent infold of unit DMvs within the central thrust panel (Fig. 1). This "Marg sequence" is overlain by massive quartzite and is in fault contact with similar underlying quartzite.

Several subparallel sulphide horizons of moderate southerly dip have been traced by drilling along an easterly strike for over a kilometre. Glacial till and colluvium is thick and bedrock bedrock exposure in this belt is largely limited to trenches. Most drilling has concentrated on the eastern portion where sulphides occur at two major structural levels that appear to reflect either fold or fault repetition. Cathro (1988) interpreted these sulphide zones to be part of a single sulphide body deformed into an overturned tight to isoclinal fold that verges to the northeast and plunges moderately southeast. Evidence presented in this study (see below) supports such a fold (Fig. 3) because: (1) lithotypes show symmetry across the axial plane; (2) when unfolded, mineralogical and geochemical patterns within the limbs combine to form single linear trends (Fig. 5); and (3) the very pyritic and ferroan carbonate-rich nature of metavolcanic schists in the core of the fold relative to metavolcanic schists above the upper sulphide horizon or below the lower sulphide horizon suggest an altered stratigraphic 'footwall' of the sulphide deposit. This later relationship also indicates the fold is an anticline (e.g. upper limb is an upright stratigraphic sequence). However, results from 1989 drilling in the western part of the Marg deposit suggests a more complex geometry of the sulphide body.

The anticline occurs above a south-dipping thrust fault that juxtaposes the Marg sequence against quartzite (Fig. 3). In section 2510E, the thrust fault is generally subparallel to the lower limb of the anticline but steepens updip and appears to truncate the nose of the folded sulphide. The thrust is a zone several metres wide of intensely foliated rock, tightly contorted bedding, with boudinage of thin quartzite beds and pod-like bodies of quartz (e.g. DH20-282-285m). In some drill intersections, the fault zone also includes zones of rubble and graphitic or chloritic gouge interpreted to represent younger fault movement (e.g. DH13-105-108m).

S1 cleavage is the dominant fabric and trends east, dips moderately south, and is parallel or subparallel to lithological contacts. In schists, the S1 cleavage is defined by the preferred orientation of minerals as well as compositional and textural banding (Fig. 4c). Black phyllites have a micaceous or carbonaceous sheen on S1 surfaces and contain discontinuous quartz lenses and bands parallel to S1 (Fig. 4d). The S1 fabric in the sulphide zones is expressed as textural or compositional banding with quartz and ferroan carbonate augen that suggest a cataclastic or mylonitic orgin (Fig. 4e).

The S1 cleavage is observed in core to be axial planar to small isoclinal folds. These F1 folds are best developed in quartz bands within black phyllites or carbonaceous metacherts and commonly have attenuated limbs or are rootless fold noses detached from limbs (Fig. 4f). Quartz rods within the S1 fabric of quartz-carbonate-sericite schists parallel F2 crenulations; as F2 folds appear coaxial to F1 folds, the quartz rods likely parallel F1 fold axes.

Within schists, F2 kink folds and crenulations of micaceous and carbonaceous foliae (Fig. 4g) are locally developed; in some cases kink bands can obscur S1. Within thicker banded quartzite, F2 folds are open.

Figure 1. Geological map of the Marg deposit area.

Fault zones of non-indurated graphitic or chloritic gouge up to several metres in drill thickness occur throughout the sequence. Based on correlation between drill holes, many of these faults are subparallel to S1 foliation. Several such faults occur within the axial region of the Marg fold. Another cuts the sulphide horizon on the upper fold limb at a shallow angle. Non-indurated fault gouge cuts the thrust fault at the base of the sulphide body (e.g. DH13-105-108m).

LEGEND
(to accompany Figure 1)

TRIASSIC(?)

TRd Hornblende diorite, greenstone

DEVONIAN(?), MISSISSIPPIAN, AND YOUNGER(?)

DMps Dark grey and rusty weathering black siliceous phyllite and siliceous siltstone

Mq Massive grey quartzite, interbedded with black siliceous shale

DMvs Black siliceous phyllite interbedded with thin quartzite laminations and rusty, buff and pale green weathering quartz muscovite and quartz chlorite phyllite.

LATE PRECAMBRIAN AND LOWER CAMBRIAN

PЄh Hyland Group: Quartz feldspar grit, quartzite, dark grey and green chloritic phyllite, and limestone

Lithologic contact; defined approximate, assumed.........

Thrust fault; defined, approximate.................................

S1 foliation.. 45

L1 mineral lineation, rodding... 40

F1 fold axis.. 40

S2 cleavage.. 70

F2 fold axis.. 60

In the following discussion, the terms 'above' and 'below' are used with reference to relative stratigraphic position (assuming an anticline geometry) rather than structural position.

LITHOTYPES IN THE MARG SEQUENCE

Quartz-sericite schist

Quartz-sericite schists are white to pale greenish grey, and composed of up to 60 % quartz, 50 % sericite and 5 % disseminated pyrite (Figs. 4c, 4g). Quartz-sericite schist up to 20 m thick is interbedded with carbonaceous metachert above the ore horizon on both limbs of the fold; minor quartz-sericite schist occurs interleaved with quartz-sericite-carbonate schists below the sulphide horizon in the core of the fold. Dark quartz augen up to1 mm in diameter, some of which are subhedral quartz phenocrysts, occur in quartz-sericite schist below the sulphide body. Zircons from quartz-sericite schists with quartz augen (i.e. DH6-133.20-141.40m) have yielded a U-Pb age of Early Mississippian (J. Mortensen, pers. comm.)and represent altered quartz phenocryst-bearing volcanic or intrusive rocks of intermediate composition.

Quartz-sericite-carbonate schist

Thin beds (10-50 cm) of grey quartz-carbonate-sericite schists occur throughout the black phyllite above the sulphide horizon (e.g. DH1-54.5m). The schist is composed of interbanded quartz-ferroan carbonate (0.1 to 7 mm) and grey green sericite (0.1 to 1 mm) bands and lenses parallel to S1. Quartz, sericite and orange weathering, grey carbonate grains (to 0.2 mm) comprise up to 70, 20 and 30 per cent of the rock respectively. The thin-bedded and probable Si-Ca-K-Mg-rich nature of these schists suggest they are altered meta-tuff.

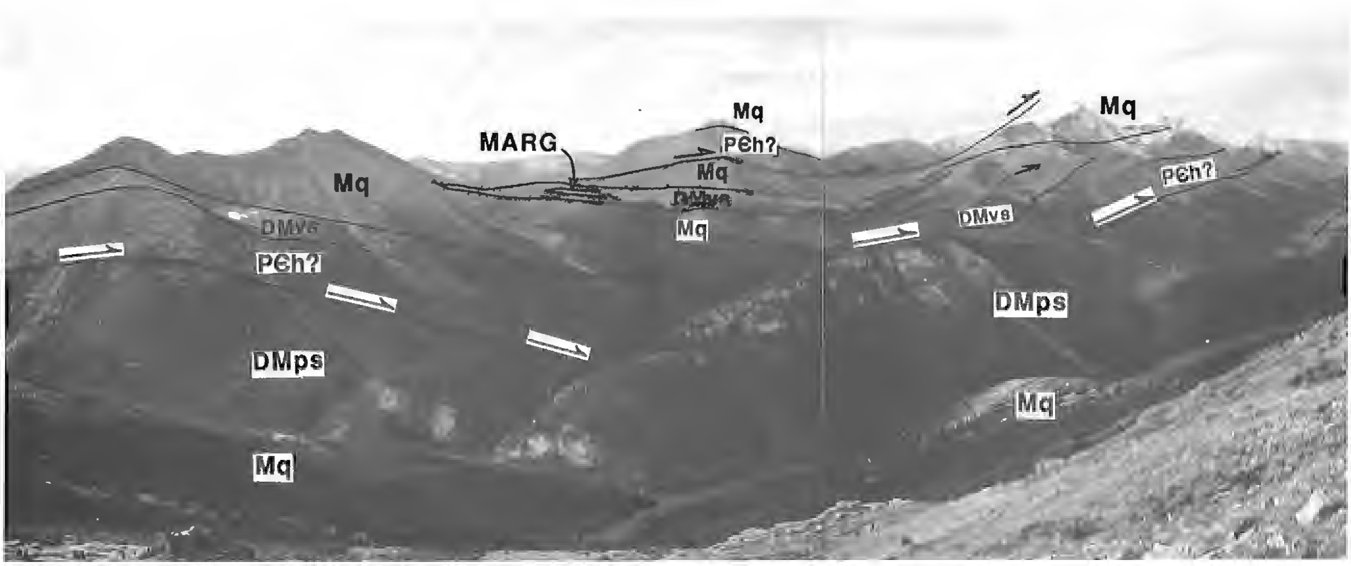

Figure 2. Photograph looking southwest at the Marg deposit and thrust repetitions of strata. Map unit symbols are described in Figure 1.

Quartz-sericite-carbonate schists also occur on the lower fold limb where they are interbedded with carbonaceous pyritic chert and thin sulphide units (Fig. 3). Here, banded quartz-sericite-carbonate units are composed of up to 75 % 3-10 mm lenticular quartz bands, 30 % 1-3 mm sericite, 10 % disseminated buff to orange weathering carbonate grains to 1mm, and 1-2 % disseminated fine grained pyrite (e.g. DH9-120m).

Sericite-carbonate-quartz-pyrite schist

Pale orange weathering, grey to cream sericite-carbonate-quartz-pyrite schist is the major lithology below the ore horizon in the core of the Marg fold (e.g. DH25-210-220m). The sericite-carbonate-pyrite schist consists of up to 60 % sericite (1-2mm), 40 % lenticular domains up to 2mm in thickness of intergrown grey carbonate and quartz, and 20 % very fine-grained pyrite (0.06-0.08mm). Minor dark quartz augen of equant to elongate shape 0.5 to 1mm in diameter occur within these schist bands. Though most augen are anhedral, some preserved crystal faces indicate formation as quartz phenocrysts. Sericite-carbonate-quartz-pyrite schist has both sharp and gradational contact with black phyllite, carbonaceous metacherts and massive and semi-massive sulphide beds.

Carbonate-quartz bands, locally deformed as F1 folds compose up to 70 % of the schist (e.g. DH20-219.5m), are up to 4 cm thick and include subhedral to euhedral grains 0.5 mm across of buff weathering cream carbonate (dolomite?). Similar carbonate-quartz bands occur within the sulphide body (e.g. DH20-227m). Carbonate-quartz bands are interpreted as syn-ore veinlets transposed into S1, though they may also represent synmetamorphic segregation bands. The presence of such veins, the high pyrite and ferroan carbonate content of schist, and proximity to the overlying sulphide horizon suggest the protolith was a stockwork of

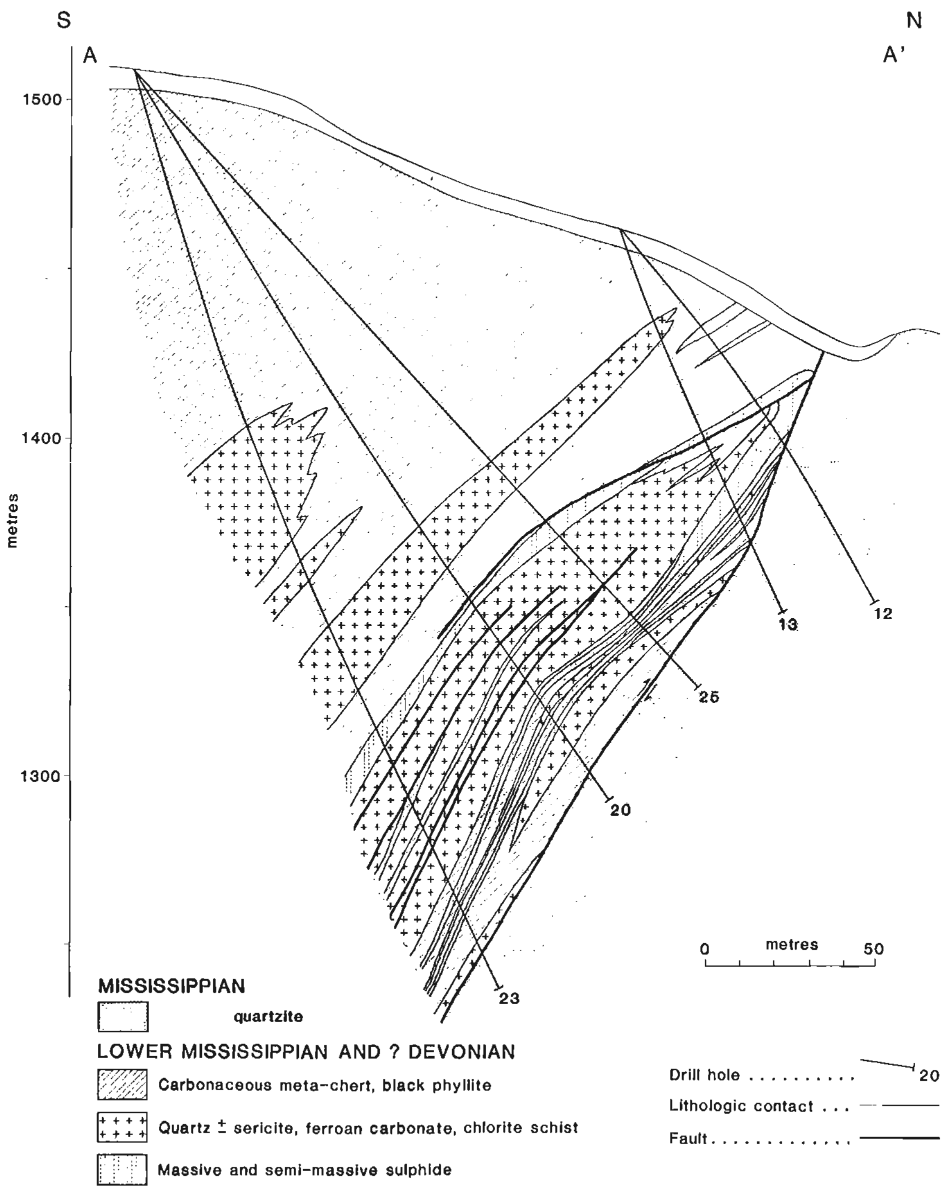

Figure 3. Structural cross-section of Marg deposit (2510E section). Location of section indicated on Figure 1.

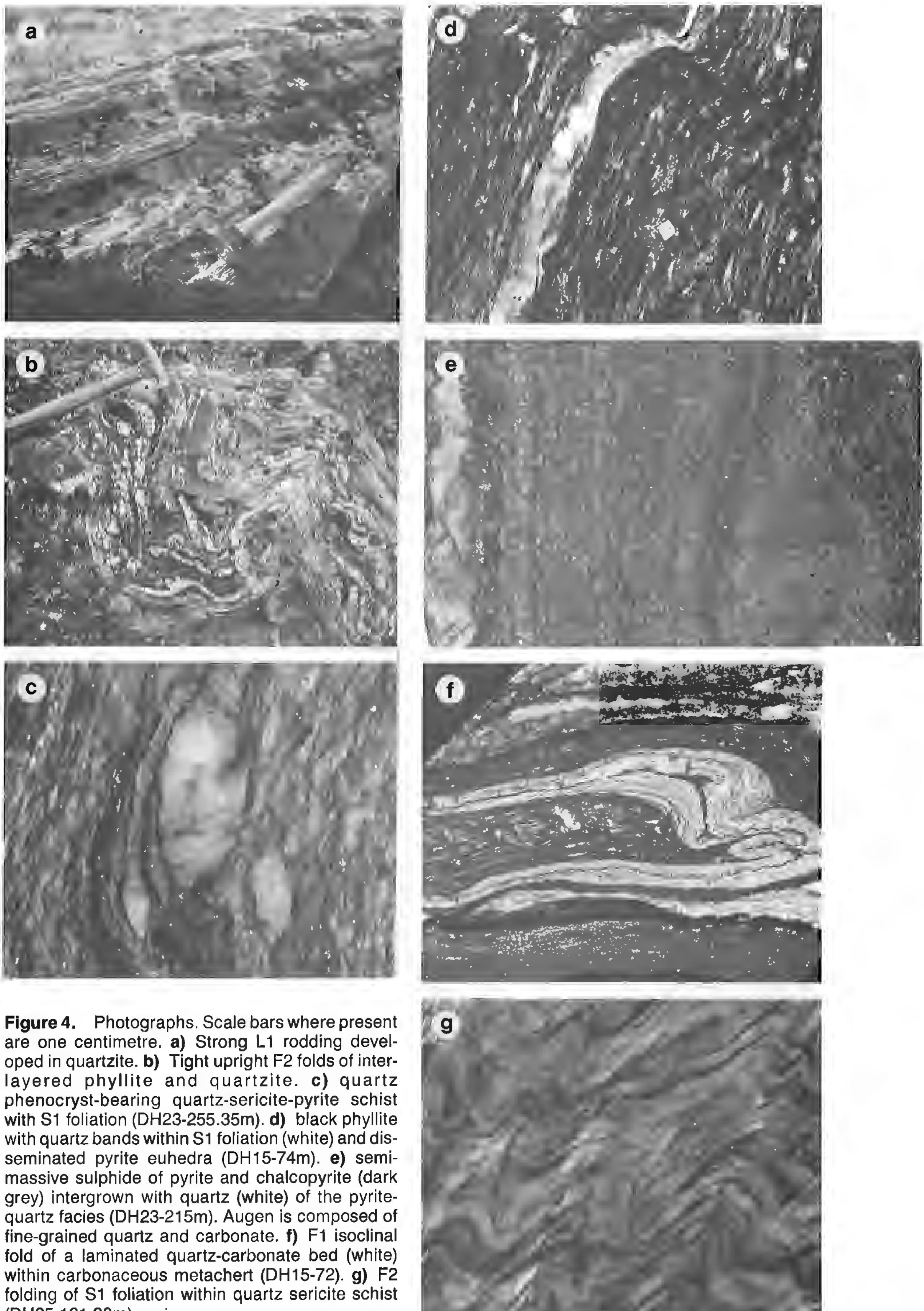

Figure 4. Photographs. Scale bars where present are one centimetre. **a)** Strong L1 rodding developed in quartzite. **b)** Tight upright F2 folds of interlayered phyllite and quartzite. **c)** quartz phenocryst-bearing quartz-sericite-pyrite schist with S1 foliation (DH23-255.35m). **d)** black phyllite with quartz bands within S1 foliation (white) and disseminated pyrite euhedra (DH15-74m). **e)** semi-massive sulphide of pyrite and chalcopyrite (dark grey) intergrown with quartz (white) of the pyrite-quartz facies (DH23-215m). Augen is composed of fine-grained quartz and carbonate. **f)** F1 isoclinal fold of a laminated quartz-carbonate bed (white) within carbonaceous metachert (DH15-72). **g)** F2 folding of S1 foliation within quartz sericite schist (DH25-161.90m).

quartz-carbonate veinlets in sericite-carbonate-pyrite altered volcanic rocks. These altered rocks likely formed in the upflow zone or vent complex of the hydrothermal system that formed the sulphide body.

Quartz-chlorite-carbonate schist

Distinctive green chlorite-bearing schists vary in composition from non-pyritic quartz-chlorite schists to pyritic quartz-chlorite-sericite-carbonate schists and occur on the upper limb of the fold below the sulphide horizon (e.g. DH23-216-225m). Quartz-chlorite-carbonate schists consist of 1 to 2 mm thick bands of quartz with minor orange weathering grey carbonate interleaved with thin bands of pale green to greenish black chlorite-sericite-pyrite schist. Dark quartz augen 1 to 3 mm in diameter, some with a relict hexagonal quartz phenocryst shape, are locally noted within chloritic schists immediately underlying the sulphide horizon. As the quartz-chlorite schist underlies the sulphide horizon at the same stratigraphic position as sericite-carbonate-quartz schist, and both lithotypes contain quartz phenocrysts, both likely represent differing alteration of the same volcanic protolith (Fig. 6).

Carbonaceous pyritic metachert

Dark grey to black sooty siliceous pyritic metachert with thin quartz bands and lenses occurs interbedded with the sulphide horizon (e.g. DH20-220-239m). It also occurs as a thick sequence above the sulphide body, and as interbeds within schist below the sulphide (Fig. 4f). The metachert is composed of microcrystalline quartz and up to 50 % by volume microgranular carbonaceous matter, has a graphitic or phyllitic sheen on S1 cleavage surfaces, and contains 5 to 15 % disseminated pyrite grains (0.05 to 0.4 mm) . Metachert interbedded with sulphide horizons includes lenticular bands of fine-grained ferroan carbonate and quartz, and up to 30 % disseminated and laminated pyrite. Metachert beds display sharp contacts with interbedded sulphides and schists. Deposition of the carbonaceous pyritic chert coincided with massive sulphide deposition and reflects pelagic or hydrothermal silica and organic matter deposition within a reduced water mass.

Black phyllite

Quartz-banded black phyllite over 150 m thick occurs above the sulphide horizon and correlates with the black recessive unit on surface below the quartzite. The black phyllite consists of microscopic to centimetre-scale segregation bands of quartz and sericite with carbonaceous matter that define the S1 foliation (e.g. DH23-260-263m) (Fig. 4d). Fine-grained quartz is the dominant mineral with lessor sericite (10 to 40 %), carbonaceous matter (5 to 40 % by volume), and minor disseminated pyrite. Quartz bands are often contorted, less commonly brecciated. High quartz and carbon, significant sericite, and low pyrite content suggest a carbonaceous siliceous shale protolith. Black phyllite is gradational with, but distinct from, carbonaceous meta-chert by higher luster (due to sericite content), lower pyrite content, non-sooty nature and black rather than dark grey colour.

Quartzite

Quartzite overlies the recessive black phyllite above the sulphide deposit and underlies the thrust fault below the sulphide body. Below the fault, quartzite beds to 2 m thick are interbedded with black phyllite and carbonaceous metachert. The quartzite is pale grey, fine-grained (up to 0.1mm), massive or thin-banded (1 to 3 mm), and contains minor sericite, carbonate, pyrite, chlorite and carbonaceous matter. Locally bands of disseminated orange- or brown-weathering ferroan carbonate parallel the S1 banding and compose 10-20 % of the rock. The interbedded nature of the quartzite with black phyllite suggests deposition within a reduced water mass, likely via turbidity flows.

CHARACTER OF THE MASSIVE SULPHIDE DEPOSIT

The Marg deposit is a folded sheet-like body or series of bodies over a kilometre long. The sulphide body consists of fine-grained massive to semi-massive pyrite intergrown with quartz, ferroan carbonate, sphalerite, chalcopyrite and galena. Contacts between the sulphide body and carbonaceous pyritic metachert and quartz-sericite-carbonate schist tend to be sharp. In section 2510E in the eastern portion of the deposit, the upper limb of the sulphide body is a single horizon 2 to 7 m thick (Fig. 3). On the lower limb, the sulphide body comprises up to seven sulphide layers 30 cm to 2 m thick interbedded with carbonaceous metachert and quartz-sericite schists.

The sulphide body is composed of fine-grained granoblastic pyrite, quartz and ferroan carbonates, lessor sphalerite, chalcopyrite and galena, and minor tetrahedrite and arsenopyrite (Fig. 4e). Pyrite is the dominant sulphide and comprises up to 90 % of the sulphide body. The sulphides are massive to banded with a granulated submylonitic texture in thin section (J. Harris, report to Archer Cathro and Associates (1981) Ltd., 1988). Augen to 3 cm of ferroan carbonate, quartz and sulphides locally occur within banded sulphide rock (Fig. 4e). Milky to pale grey quartz and cream to grey carbonate occur intergrown with sulphides, or as lace-like networks, irregular clots, or bands up to 5 mm thick parallel to S1 foliation. Carbonate minerals include an early medium-grained siderite (?) commonly cut by veinlets of coarse-grained buff ankerite (?) dolomite (?).

Pyrite occurs as aggregates of microscopic euhedral crystals, often with interstitial sphalerite, chalcopyrite and galena (*ibid.*, 1988). Galena and brown to amber sphalerite grains up to 0.175 mm form pods up to 1 mm in diameter. Chalcopyrite occurs as disseminated grains or as irregular clots or veinlets with quartz and carbonate. Minor arsenopyrite occurs in intimate intergrowth with pyrite; tetrahedrite grains are often associated with chalcopyrite and galena (*ibid.*, 1988).

In the eastern portion of the Marg deposit, the sulphide body can be divided into three facies based on the proportions of the dominant minerals pyrite, quartz and ferroan carbonate. A central core of carbonate-rich semi-massive pyrite (pyrite-carbonate facies) is surrounded by a transitional envelope of semi-massive quartz-rich pyrite (pyrite-quartz facies) and distal massive pyrite (pyrite facies) (Fig 5a).

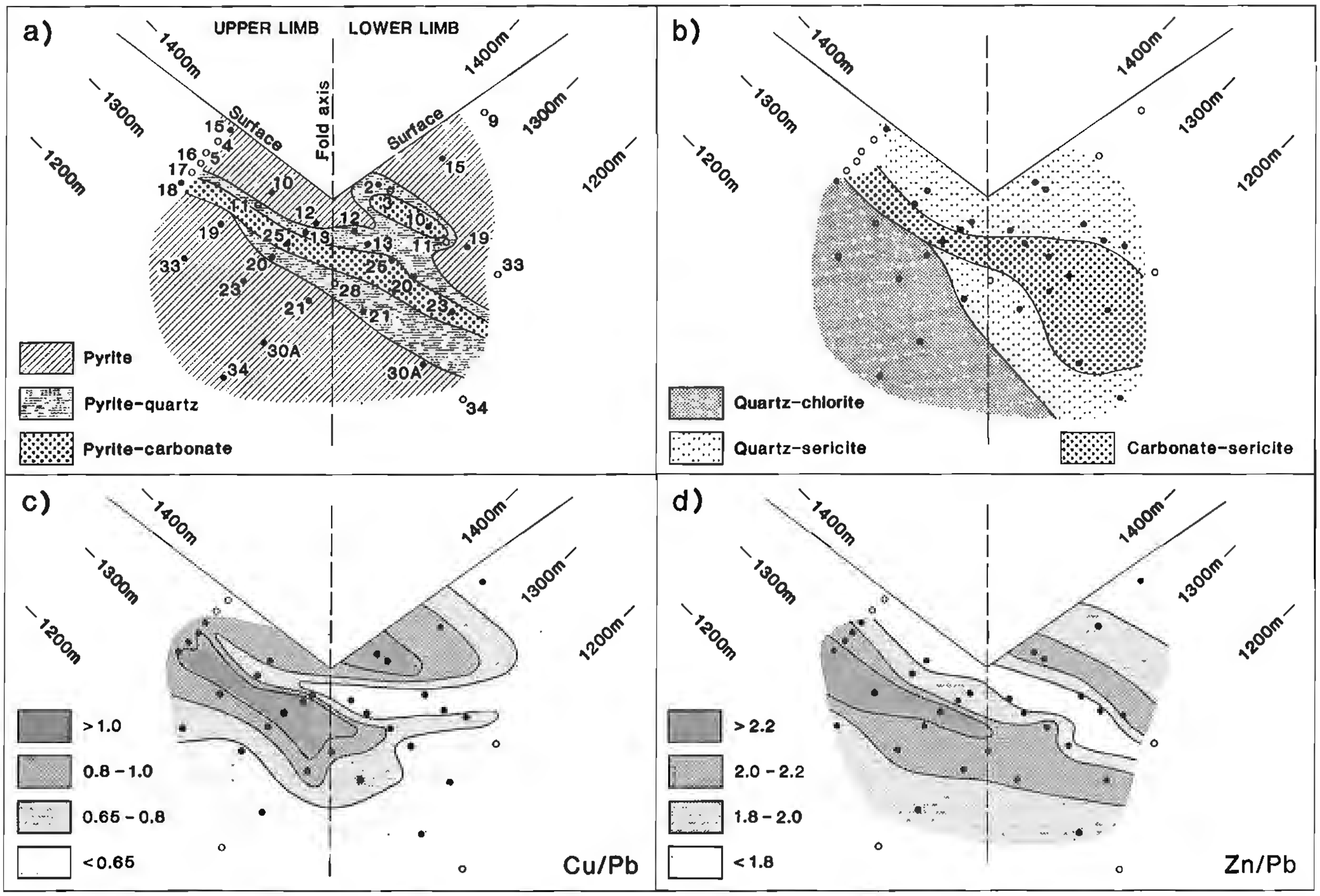

Figure 5. Vertical projection of northeastern part of Marg sulphide body unfolded about assumed northeasterly plunging fold axis. Base map from Cathro (1988). **(a)** Distribution of dominant sulphide facies. The points where the section is pierced by drill holes are indicated by circles with drill hole number. Solid circles indicate drill holes logged in this study; open circles are drill holes not logged. **b)** Distribution of dominant alteration facies in footwall schists. **c)** Drill hole assay data for sulphide horizon: Cu/Pb ratio. Metal ratios from R. Cathro (unpublished data). **d)** Drill hole assay data for sulphide horizon: Zn/Pb ratio. Metal ratios from R. Cathro (unpublished data).

Pyrite facies

Massive sulphide with less than 20 % intergrown quartz and ferroan carbonate is the most extensive facies in the eastern Marg deposit (Fig. 5a). On the upper limb of the fold in section 2510E, pyrite facies overlies pyrite-quartz or pyrite-carbonate facies at the top of the sulphide body . Pyrite facies is often banded and composed of pyrite with lessor sphalerite, chalcopyrite and galena. Bands reflect compositional or grain size variation on a scale of millimetres to a centimetre that define the S1 foliation. Very fine-grained pyrite is commonly interbanded with medium-grained pyrite-sphalerite-galena-chalcopyrite (e.g. DH25-181.8m).

Pyrite-quartz facies and pyrite-carbonate facies

Sulphide-rich rock composed of greater than 20 % quartz and ferroan carbonate, and with quartz more abundant than carbonate is referred to as the pyrite-quartz facies (Fig. 4e). On the upper limb of the fold in section 2510E, the pyrite-quartz facies is transitional laterally to pyrite-carbonate facies, both of which are overlain by a blanket of pyrite facies.

Sulphide-rich rock composed of greater than 20 % quartz and grey ferroan carbonate and carbonate more abundant that quartz is referred to as pyrite-carbonate facies. Pyrite-carbonate facies may be massive, display weakly developed carbonate-quartz-rich and pyrite-rich bands, or be strongly banded with compositional layers of very fine-grained pyrite, and medium-grained carbonate-sulphide and pyrite-carbonate-sulphide.

Minor bands to 50 cm thick of massive brownish-grey weathering grey carbonate occur within the pyrite-carbonate facies (e.g. DH20-184.7m). The carbonate is medium-grained with up to 15 % disseminated pyrite, minor pods of sericite rimmed by buff carbonate up to 2-5 mm thick, and irregular veinlets of buff carbonate, quartz, and chalcopyrite. A similar 10 cm thick massive carbonate band occurs in the core of the fold stratigraphically below the sulphide horizon on the lower limb interbedded with sericite-carbonate-quartz-pyrite schists (e.g. DH20-215.4-215.5m).

ZONATION WITHIN EASTERN MARG DEPOSIT

Facies zonation, sulphide body

The sulphide body is strongly zoned with respect to dominant ore facies. Based on a limited number of drill intersections, pyrite-carbonate appears to be the dominant facies in an elongate region greater than 400 m in length but possibly less than 30 m wide (Fig. 5a); pyrite-carbonate facies on the lower limb in drill hole 10 may represent a second sulphide-carbonate zone. Pyrite-quartz facies is dominant in a narrow envelope around the sulphide-carbonate facies. The pyrite facies is dominant over a broad region flanking the pyrite-quartz facies.

Metal zonation, sulphide body

The sulphide body is zoned with respect to metal ratios. Elevated ratios of Cu/Pb and Zn/Pb occur in an elongate pattern largely coincident with the extent of the pyrite-carbonate and pyrite-quartz facies (Fig. 5c,d). This zonation is best displayed in the upper limb sulphide horizon of the 2510E drill hole section where the highest copper and zinc enrichment with respect to lead coincides with the pyrite-carbonate facies (e.g. DH-25); this enrichment decreases outwards through pyrite-quartz and pyrite facies. Maxima in Ag/Pb and Cu/Cu+Zn+Pb ratios also coincide with the sulphide-carbonate facies. The metal ratio patterns are considered to be primary in origin because they mimic sulphide facies patterns. Some degree of metal mobility may have occurred during metamorphism and deformation but it does not appear to have obscured the original pattern of metal distribution.

Mineralogical zonation, 'footwall' schists

Both schists and carbonaceous pyritic metacherts immediately underlie the sulphide horizon (Fig. 3). Little evidence of hydrothermal alteration exists in metacherts likely due to the more chemically inert nature of these quartz-rich rocks. However, a strong mineralogical zonation exists within schists reflecting a reactive volcanic protolith suceptable to hydrothermal alteration. Sericite-carbonate-quartz-pyrite and carbonate-quartz-sericite-pyrite schists occur in an elongate zone underling much of the pyrite-carbonate facies of the sulphide horizon (Fig. 5b). These carbonate-bearing schists are flanked by pyritic quartz-sericite schists that underlie the pyrite-quartz facies and proximal part of the pyrite facies. Chlorite-quartz schists occur downdip on the upper fold limb underlying pyrite facies.

DEPOSITIONAL SETTING OF MARG MASSIVE SULPHIDE

Schists lack original depositional textures though sharp lithological contacts may represent bedding. A volcanic origin of at least some of the quartz-sericite-chlorite-rich schists in the Marg sequence is supported by (1) an early Mississippian U-Pb age for zircons from a quartz-sericite schist in the stratigraphic footwall of the sulphide horizon, (2) the presence of quartz phenocrysts within the schists, and (3) the copper-rich nature of the sulphide body, a characteristic of volcanogenic but not sediment-hosted sulphide deposits. Harris (unpublish report to Archer Cathro and Associates (1981) Ltd. 1988) has noted that the absence of albite or epidote in the sericite and chlorite schists, characteristic of greenschist grade metavolcanic rocks, argues against a volcanic component. However, feldspar destruction and alteration to quartz, sericite, chlorite and carbonate has been noted within many alteration pipes below massive sulphide deposits (Franklin et al., 1981). Major and trace element analyses of the schists are currently underway and will provide better constraints on the chemical nature of the schist protolith.

The Marg sulphide deposit occurs interbedded with carbonaceous metacherts and schists. In section 2510E, the sulphide body overlies a sequence of schists at least 30 m thick, and is overlain by at least 200 m of dominantly black phyllite and carbonaceous pyritic metachert. This stratigraphy suggests that sulphide deposition coincided with waning of volcanic activity within an anoxic deep marine basin dominated by hemipelagic and pelagic deposition of muds, biogenic silica and organic matter.

In eastern and central Yukon, Mississippian alkaline intermediate to felsic volcanic rocks with associated Cu-Zn-Pb sulphide deposits were only known in the Pelly Mountains west of the Tintina fault (Mortensen and Godwin, 1982; Gordey, 1981) prior to the discovery of the Marg deposit. This quartzite, carbonaceous siliceous shale and felsic volcanic assemblage represents a new exploration target. Equivalent strata have been traced across central Yukon intermittently from Dawson (116B) to Nahanni (105I) map areas (Green, 1972; Abbott, 1983, Gordey, in press). Although the genetic relationship, if any, between the quartzite and sulphides is unclear, the quartzite, being thicker, more extensive, and better exposed than the volcanic rocks should serve as a useful exploration guide.

The Marg deposit is a member of a family of Devonian and Mississippian sediment-hosted and volcanic-hosted exhalative sulphide and barite deposits that occur within the Cordillera from Alaska to Mexico (Turner, 1988). Volcanogenic massive sulphide deposits include the Ambler, Alaska (Schmidt, 1986), Clear Lake and MM, Yukon (Grapes, 1987; Mortensen and Godwin, 1982), and Samatosum, Silver and Homestake deposits, B.C. (Hoy and Goutier, 1986). Studies of the Arctic, MM and Samatosum deposits all indicate association with alkaline volcanic rocks typical of rift environments; host stratigraphies suggest an outer continental margin setting.

FORMATION OF MASSIVE SULPHIDE ORES

The ferroan carbonate-rich portion of the suphide body and underlying schists is interpreted to represent the hydrothermal upflow zone or vent complex during deposition of the massive sulphide body based on: (1) copper enrichment as is often noted in the vent complexes of other volcanogenic massive sulphide deposits (e.g. Franklin et al., 1981); (2) the abundance of carbonate-quartz bands within footwall schists interpreted to represent a deformed vein stockwork; and (3) analogy to less deformed sediment-hosted zinc-lead deposits of similar age that have iron carbonate-rich vent complexes (Turner, 1986; Turner et. al., 1989; Goodfellow et al., 1989).

A model for the formation of the Marg deposit is proposed. The core of the hydrothermal upflow zone is represented by carbonate-quartz stockwork in sericite-carbonate-quartz-pyrite altered volcanic rock and overlying carbonate-rich massive sulphide. Quartz-rich massive sulphide rock and quartz-sericite-pyrite altered volcanic rocks represent a more peripheral alteration associated with the upflow zone. Pyrite facies massive sulphide reflects less altered sedimentary suphides, and chlorite-quartz schists reflect less altered volcanic rocks away from the center of hydrothermal alteration. A core of sericitic alteration transitional outwards into chloritic alteration is commonly noted in Zn-Pb-Cu volcanogenic massive sulphides (Franklin et al., 1981). The presence of ferroan carbonate-enrichment in the vent zone has been described in only a small group of volcanogenic massive sulphide deposits such as Mattabi, Ontario (Franklin et al, 1975) and Madenkoy, Turkey (Cagatay and Boyle, 1980).

The absence of a baritic fringe on the Marg deposit is suprising as it is typical of Devonian and Mississippian sediment-hosted (e.g. Jason, Tom, Cirque) and volcanogenic (e.g MM) deposits elsewhere in the Selwyn Basin.

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Stratigraphy of the middle Proterozoic Gillespie Lake Group in the southern Wernecke Mountains, Yukon

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Abstract

The Gillespie Lake Group (GLG) is the uppermost unit of the Wernecke Supergroup (>1.3 Ga). In the study area (106D/8), the GLG is about 1.2 km thick. However, thrust faults repeat the GLG on many ridges, giving an appearance of greater thickness. Some subdivisions of the GLG proposed for its type area were not recognized. In the study area the most practical GLG division is into lower and upper units. The lower GLG consists of dolosiltites, dololutites and fine grained dolarenites, gradational from underlying Quartet Group siliciclastics. The upper GLG contains distinctive shallowing-upward sequences, commonly capped by columnar and domal stromatolites. Several features of the upper GLG could serve as local marker horizons for detailed mapping and resolution of fault-repeated stratigraphy. These include rare volcanic flows or tuffs, distinctive stromatolites, cyclic sedimentation units, beachrock and beach rosette horizons, and unusually well-preserved oolitic beds.

Résumé

Le groupe de Gillespie Lake est l'unité sommitale du supergroupe de Wernecke (> 1,3 Ga). Dans la région étudiée (106D/8), le groupe mesure environ 1,2 km d'épaisseur. Toutefois, les failles chevauchantes répètent le groupe qu'on retrouve sur de nombreuses crêtes, donnant l'apparence d'une plus grande épaisseur. Certaines subdivisions proposées du groupe pour sa région type n'ont pas été reconnues. Dans la région étudiée, la division la plus pratique du groupe consiste en une unité inférieure et une unité supérieure. L'unité inférieure est constituée de siltites dolomitiques, de pélites dolomitiques et d'arénites dolomitiques à grain fin, gradationnelles à partir des roches détritiques siliceuses du groupe de Quartet sous-jacent. L'unité supérieure renferme des séries distinctes de moins en moins profondes vers le haut, dont le sommet est généralement recouvert par des stromatolithes columnaires et en dôme. Plusieurs structures de l'unité supérieure pourraient servir d'horizons marqueurs locaux pour un levé géologique détaillé et une résolution de la stratigraphie répétée par faille. Ces structures renferment de rares coulées ou tufs volcaniques, des stromatolithes distincts, des unités de sédimentation cyclique, des horizons de grès de plage et des horizons avec des textures en rosette de plage, ainsi que des bancs oolitiques exceptionnellement bien conservés.

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INTRODUCTION

The Gillespie Lake Group (GLG) is the uppermost unit of the tripartite Wernecke Supergroup (Delaney, 1978, 1981, 1985), the oldest sedimentary package exposed in the northern Cordillera (Fig. 1). These rocks are crosscut by heterolithic breccias emplaced at about 1.3 Ga (Parrish and Bell, 1987) and have been correlated with the middle Proterozoic Belt-Purcell Supergroup of B.C. and northern U.S.A. (Gabrielse, 1972; Young et al., 1979; Delaney, 1981, 1985). Characteristically orange-brown weathering, the thick succession of dolostones which make up most of the Gillespie Lake Group forms a distinctive map unit in both Ogilvie and Wernecke mountains.

Delaney (1981, 1985) suggested that the total thickness of the GLG exceeds 4 km in its type area (about 50 km north of our study area). He tentatively subdivided the GLG into seven units of probable formational status, and recognized subdivisions within some of these units. The main divisions include a basal unit (G-TR) transitional from the underlying Quartet Group, overlain in ascending order by units G-2 to G-7. Delaney recognized that some of these units are not laterally persistent and noted that factors such as fault repetition, abrupt facies changes and the cyclicity of some facies associations cast uncertainty on the validity of some detailed subdivisions and the total thickness estimate for the GLG. However, he speculated that at least units G-TR and G-7 would prove to be regionally persistent.

This study compares the stratigraphy of the GLG in an area south of the area examined in detail by Delaney. We here describe distinctive lithological associations and sedimentary structures which should be useful for regional and detailed mapping and should also aid geologists concerned with recognizing or tracing parts of the GLG which may be economically significant, especially in areas of locally complicated structure.

This study complements 1:50 000 mapping in the southern Wernecke Mountains, initiated in 1989 by the Geological Survey of Canada (Roots, 1990) in response to renewed interest in the Pb-Zn-Ag potential of carbonate rocks. The results are pertinent to evaluation of GLG strata on the nearby Blende property (Fig. 2; Roots, 1990) and support correlation with strata in the upper Fifteenmile River area of the Ogilvie Mountains, 250 km to the west.

REGIONAL STRUCTURE AND STRATIGRAPHY

The Wernecke Mountains comprise middle to Late Proterozoic rocks uplifted by northward telescoping of the ancient North American miogeocline beginning in mid-Jurassic time. To the south, east and northwest, these Proterozoic strata are overlain by lower Paleozoic carbonate units of the Mackenzie Platform.

The area mapped in 1989 (Fig. 2; Roots, 1990) is south of the high sawtoothed ridges of the central Wernecke Mountains. Although the topography is more subdued, units younger than the Wernecke Supergroup are preserved, and repetition of the stratigraphic succession across the map area is apparent, consistent with the thin-skinned thrust faulting recognized in the Ogilvie and Wernecke mountains (Thompson and Eisbacher, 1984).

The Proterozoic stratigraphy in the map area comprises the top two groups of the Wernecke Supergroup (Quartet and Gillespie Lake groups) and an overlying, unnamed unit, here referred to as unit 4. The Quartet Group in the map area consists of grey to grey-green siliciclastic siltstone, argillite and minor fine-grained sandstone. The contact between the Quartet Group and the overlying GLG is gradational, reflecting an upward transition from siliciclastic rocks to dolostone. Unit 4 comprises dark-brown-weathering carbonates containing some stromatolitic layers, black siliciclastic siltstone and brown-to-grey mudstone. This unit is tentatively correlated with a Middle to Upper Proterozoic succession to the north and east termed the Pinguicula Group by Eisbacher (1981). Some unusual stromatolites in unit 4 are of the same form as stromatolites in the informally named Fifteenmile group in the Ogilvie Mountains to the west, also the possible equivalent of the

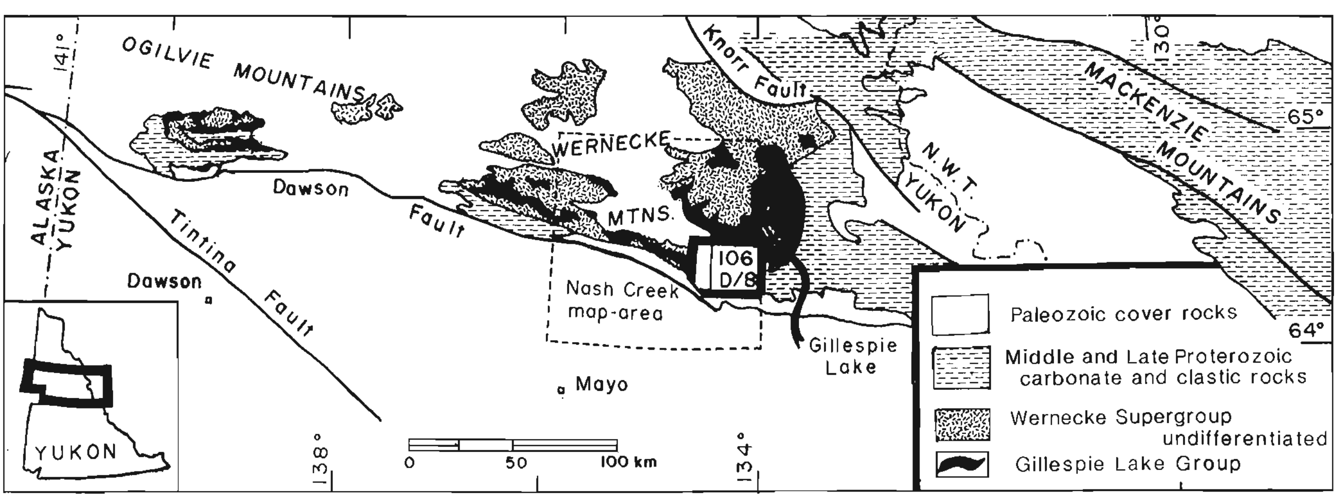


Figure 1. Distribution of Gillespie Lake Group in the northern Cordillera of the central Yukon. The study area is shown in detail in Figure 2.

Pinguicula Group (P.S. Mustard and J.A. Donaldson, unpub. data). More detailed study of all these units is required to confirm these tentative correlations. Disconformably overlying unit 4 is a thick-bedded, light grey dolostone, the equivalent of unit CDb of Norris (1976; unit 8 of Green, 1972). The CDb unit ranges in age from Cambrian to Middle Devonian.

Diorite sills up to 100 m thick intrude unit 4 and GLG rocks, and commonly form the soles of thrust panels. The age of the dykes is unknown, but lithologically similar dykes in CDb 30 km west of the mapped area indicate Paleozoic igneous activity.

In the mapped area, three structural blocks are separated by two broad, northwest-trending valleys that probably obscure transcurrent faults which postdate most or all thrusting. Within each structural block are a series of folds (containing CDb and unit 4) or imbricated thrust panels (all Proterozoic strata) that result in consistently south-dipping

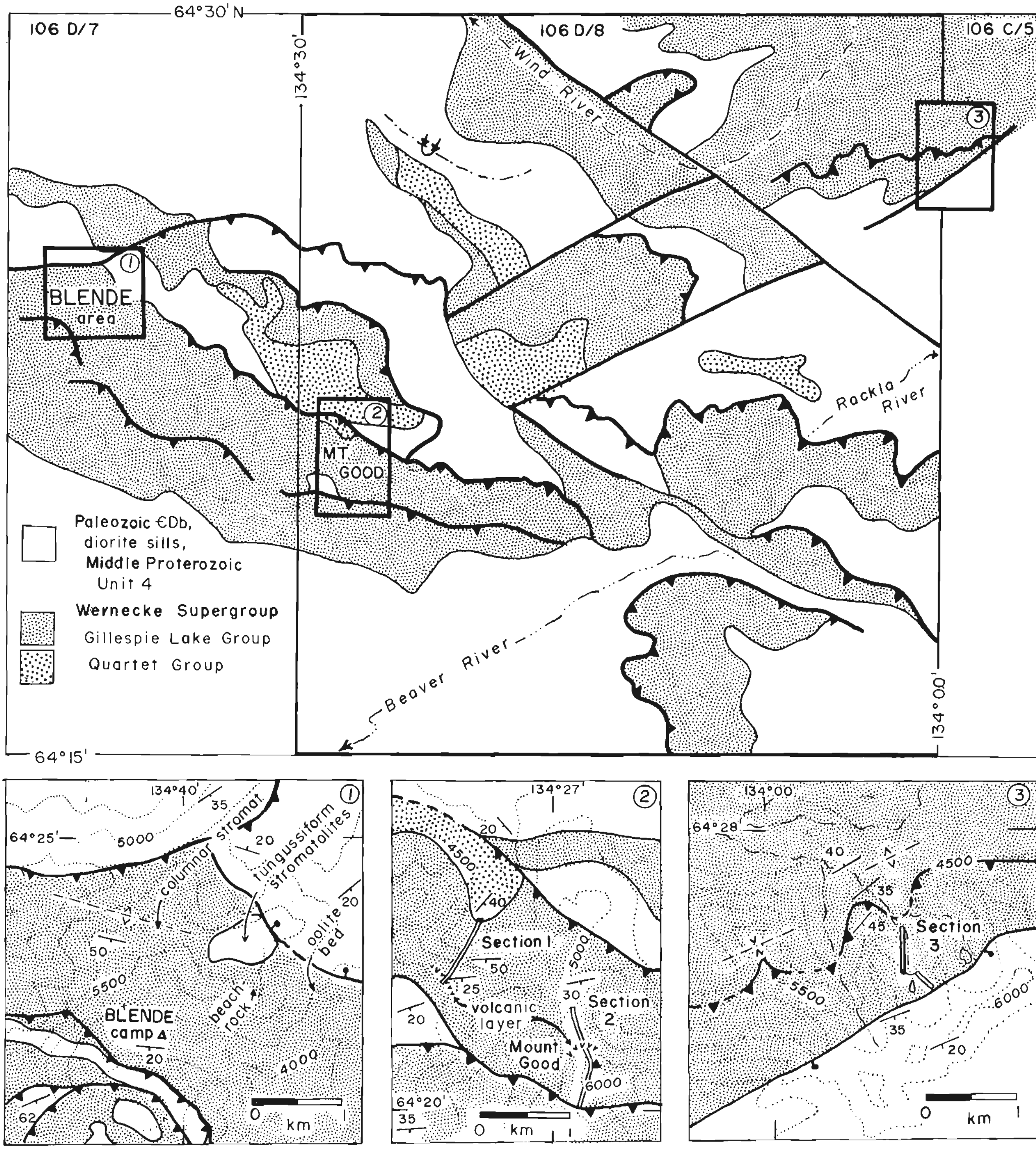


Figure 2. General geology and location of the measured sections.

successions. Actual fault surfaces are obscured by rubble and presumably parallel the bedding. The stratigraphy of the GLG was examined both in the southwest block (here called the Mt. Good panel) and in the northeast corner of the mapped area.

THE GILLESPIE LAKE GROUP IN MAP AREA 106 D/8

The Mt. Good panel displays the GLG in a 10 km long north-facing scarp up to 1500 m high. This scarp preserves the most complete section of GLG in the map area, with the underlying Quartet Group exposed in the canyon north of the scarp, and the overlying unit 4 preserved in a down-faulted outlier to the south. A vertical fault juxtaposes uppermost GLG dolostone and unit 4 chert. These strata are so commonly adjacent along the Mt. Good and other panels that the stratigraphic throw across the fault is unlikely to be more than a few tens of metres. Two sections were measured and are described below (Fig. 4, 6).

A third section was measured through an unusually well-exposed outcrop of the upper part of the GLG in the northeastern corner of the mapped area, 27 km south of Gillespie Lake. To the east, north and west of this locality, all ridges are composed entirely of moderately tilted GLG strata. The great apparent thickness of the GLG in these ridges (>3 km) is deceptive, because close inspection of long ridge spurs trending across strike (eg. Fig. 3) reveal repetition of packages distinguished by slight variations in lithological associations or overall colour, demonstrating that layer-parallel thrust faults are common in the GLG. The true thickness of the GLG, at least in this map area, is thought to be about 1.2 km (see below), although a range of several hundred metres in thickness is predicted regionally because of variations of internal facies.

The following observations result from the detailed measurement of the three stratigraphic sections (Fig. 2), supplemented by the regional work of Roots (1990) and examination of outcrop at the Blende Pb-Zn-Ag prospect (Fig. 2). The GLG is pervasively dolomitized and some original textures are lost. However, it is possible to distinguish or surmise the original clay, silt and sand sizes of the clastic carbonates which make up most of the GLG. For this reconnaissance we have used simple lithologic terms for these dolostones (dololutite, dolosiltite, dolarenite). Delaney (1985) has described the dolomitization textures and petrography in considerable detail.

Section 1, about 2 km northwest of Mt. Good (Fig. 2, 4), is in the lower part of the Gillespie Lake Group, which is in gradational contact with the underlying grey siliciclastic mudstone, siltstone and fine-grained sandstone of the Quartet Group (Fig. 5a). Delaney defined the contact by the first occurrence of orange- to brown-weathering dolosiltite or dololutite. The lowest outcrops shown in Figure 4 contain <10 % of these lithologies, and thus may be within a few tens of metres of the base of the G-TR unit. Orange-weathering dolosiltites and brown to medium-grey dololutites are increasingly common upward, giving a striped appearance to the beds (Fig. 5b). This part of the section is talus-covered and may correspond to Delaney's units G-2

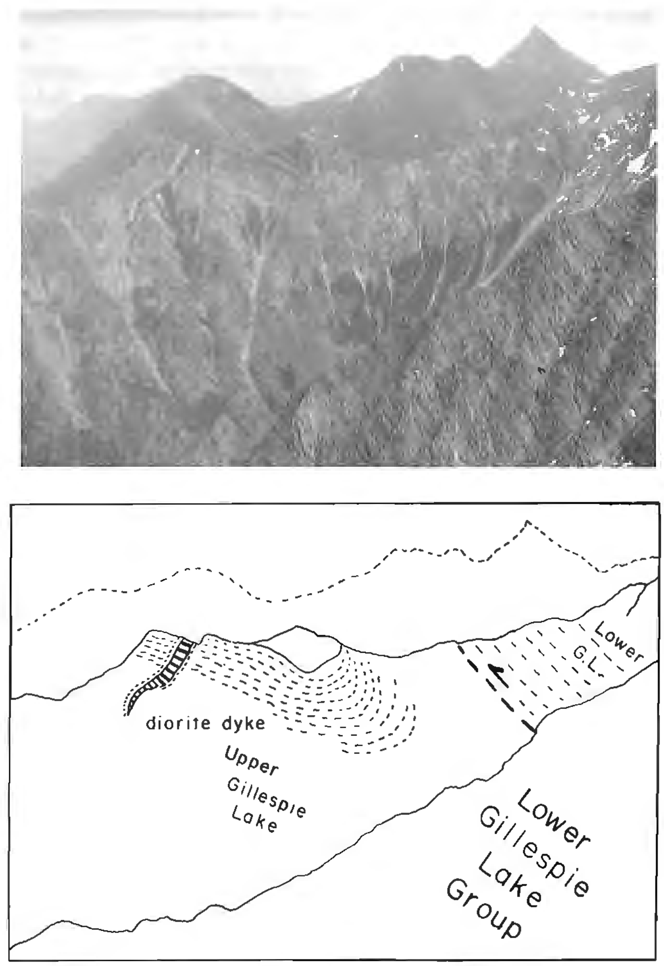


Figure 3. Structural thickening within the Gillespie Lake dolostone is indicated by the faulted fold, and juxtaposition of lower division over upper division strata. This spur is 1 km west of section 3.

and G-3. Coarsening and thickening-upward sequences up to 7 m thick are well-developed over about a 100 m interval in the central part of the section (Fig. 4, expanded sketch and Fig. 5c). These consist of basal grey mudstone (mixed dolomitic-siliciclastic) which grade upward to orange, fine-grained dolarenite. The thickest sequences are capped by up to 3 m of pink-orange weathering dolosiltite-arenite with contorted and folded internal bedding. These soft-sediment slump folds show a preferential rollover of folds towards the south. Except for the rare beds showing evidence of soft-sediment deformation, bedding and laminae are generally continuous, parallel, and straight to slightly wavy. Normal grading is common in the dolosiltite. Although not all of these small-scale cycles are well-developed, they can generally be recognized over several hundred metres vertically in the section. Delaney (1981, 1985) recognized repeated coarsening upward sequences of similar scale and lithology as part of his unit G-4. The upper part of section 1 consists generally of parallel, straight to slightly wavy laminated and thin-bedded dolosiltite, lutite and fine-grained dolarenite, commonly normally graded (Fig. 5d). These may correspond to part of Delaney's unit G-5. Rare cross-laminated ripples and small-scale trough crossbedding all demonstrate

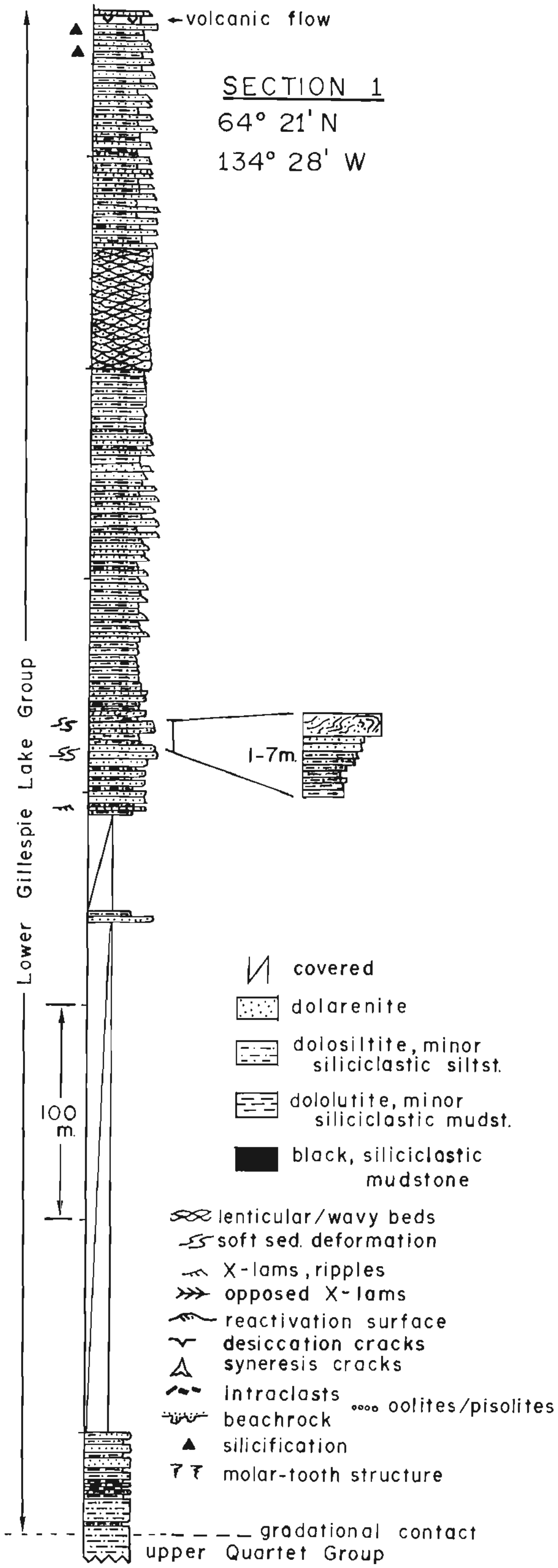


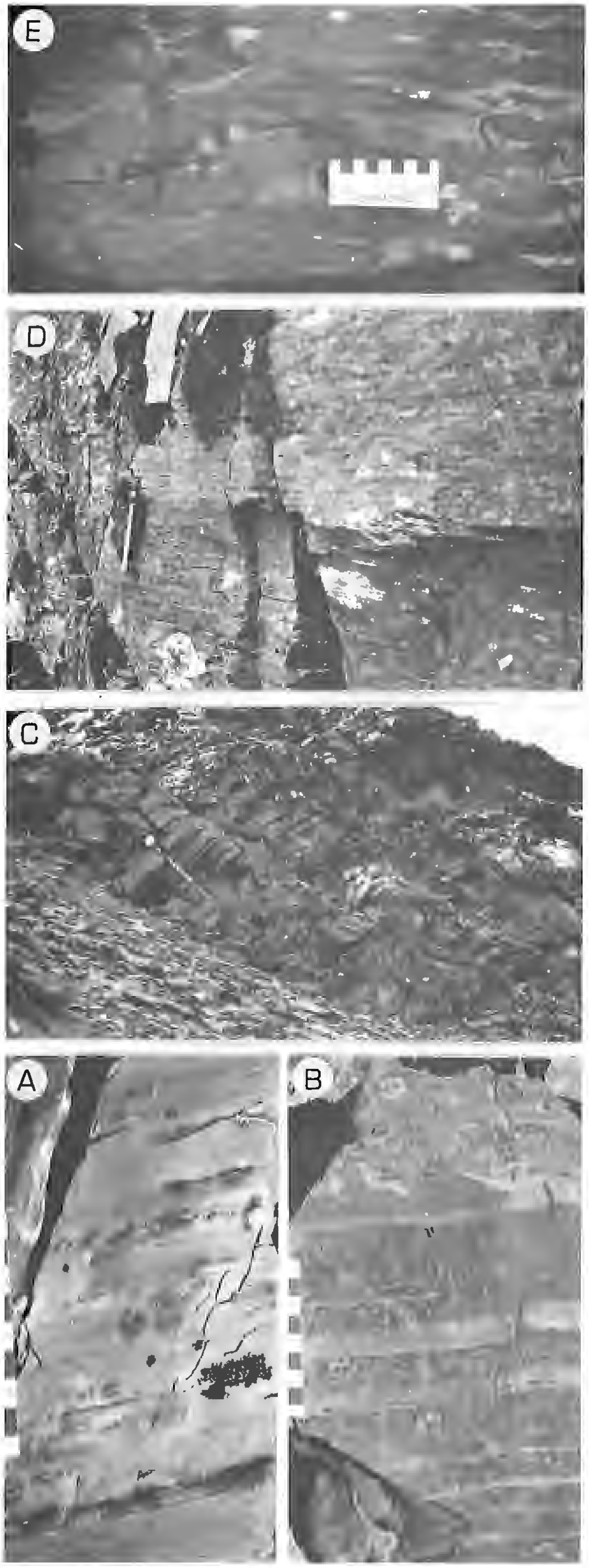
Figure 4. Section 1, about 2 km northwest of Mt. Good, shows the stratigraphy of the lower Gillespie Lake Group in this area. This section includes the approximate equivalents of units G-TR, G-2 to G-6 of Delaney (1981, 1985), although not all of these units were recognized in this section.

southerly paleotransport. An exception is an interval about 50 m thick consisting of lenticular interbedded dolosiltite and lutite and abundant elongate carbonate nodules (Fig. 5e).

A thin, discontinuous pale green, highly altered volcanic bed (unique on this ridge) marks the top of the measured section. Now foliated and chloritic, this granular volcanic rock is increasingly fine-grained upward and is locally intercalated with tan-weathering dolostone. The altered volcanic may have been a mafic tuff with a reworked top. Rubbly outcrop above this volcanic bed contains stromatolites and other structures similar to those observed about 1500 m southeast on the well-exposed northern face of Mt. Good (section 2). No major faults were recognized between these sections. A single volcanic bed in the Mt. Good section is thought to occur at the same stratigraphic position as the volcanic bed in section 1. The volcanic rock in section 2 is also altered, foliated and chloritic, but contains clay-filled amygdules and green phenocrysts, equant and up to 3 mm across, possibly originally augite. Correlation of sections 1 and 2 indicates the GLG is 1.2 km thick. A high-angle fault separates the top of section 2 from the overlying map unit 4, so this thickness is a minimum. However, the thick, stromatolite-rich unit of section 2 can only be the equivalent of the G-7 unit of Delaney, the top GLG unit, so faulted out amount of section is unlikely to be significant.

The upper part of the GLG was examined in two sections (Fig. 6 and 8, located in Fig. 2). Although about 25 km apart, both sections comprise similar stromatolite-rich successions containing most of the features Delaney used to define his unit G-7, the top unit of the GLG. Delaney suggested that unit G-7 ranges from 400-700 m in thickness in the Gillespie Lake type area. We measured about 600 m at section 2 (Fig. 6) and 460 m at the eastern section (section 3, Fig. 8), but these are minimal thicknesses because both sections are cut by faults at the top and the lower contacts are not exposed.

The most prominent feature of this unit is the abundance and variety of stromatolites and cryptalgal laminated dololutites-siltites. The stromatolites occur most commonly as the cap of shallowing-upward sequences up to 20 m thick (most 5-15 m). These are well-developed at section 3 and less obvious at section 2. A complete sequence contains interbedded dololutite/siltite at the base passing up to dolarenite and undulose cryptalgal laminates or fine-to-medium grained dolarenite. These are capped by up to 5 m of stromatolites, in many cases with a basal subdivision of columnar stromatolites which are overlain by laterally linked domal stromatolite biostromes. The stromatolites are covered by dololutite of the next cycle, although in some of the well-developed cycles a distinctive black siliciclastic mudstone a few decimetres thick immediately overlies the domal stromatolites. Desiccation cracks are common in the dololutite and siltite; syneresis cracks are rarer. Variations on this typical shallowing-upward sequence occur. Some sequences do not contain stromatolites and are capped by thick dolarenite beds (originally oolitic as in Fig. 7a, but oolites are generally only preserved in silicified patches). Abundant molar tooth structure also occurs (Fig. 9a). This unusual structure, possibly a primary cryptalgal



organosedimentary feature (O'Connor, 1972) or deformed shrinkage cracks (Yeo et al., 1977), is most commonly associated with cryptalgally laminated dolosiltite-lutite where the stromatolites are not present as the cap of a shallowing-upward sequence. The features of these sequences indicate an upward transition of the depositional environment in the order subtidal-intertidal-supratidal (James, 1984).

Patchy silicification of both stromatolites and dolarenite is typical and, in the dolarenite, has preserved textures and structures destroyed by dolomitization of the non-silicified parts of the beds. Features preserved in the silicified patches suggest that most dolarenite was originally oolitic, especially where the dolarenite is medium-grained, and contain abundant trough and planar crossbedding. The dolosiltite and very fine-grained dolarenite show slightly better preservation of sedimentary structures. Ripple cross-laminations are common and small trough crossbeds occur locally. Most crosslaminations indicate southerly paleoflows, but in a few examples closely associated thin beds show reversals of paleoflow, probably a result of tidal currents. Rarely, reactivation surfaces can be recognized in the sets of ripple cross-laminations.

Other distinctive lithotypes and structures include abundant angular to subrounded tabular intraclasts of dololutite and dolosiltite. Many are rip-ups of cryptalgal dolostones (up to 15 cm long, most <1 cm thick) that occur as intraclast breccia beds a few centimetres thick above or below the stromatolite beds, or as fill between domal stromatolites. Some beds of tabular intraclasts are vertically stacked and form polygonal clusters on bedding surfaces (Fig. 7b). In light of the abundant desiccation structures and other evidence of intermittent supratidal and intertidal deposition, these clusters probably formed as beach rosettes, reflecting shoreline swash-backwash conditions (cf. Ricketts and Donaldson, 1979). In section 2 (Fig. 6, 260 m above the base), a fine-grained dolarenite bed occurs with intraclast beds (some forming beach rosettes). The dolarenite has an eroded, irregular upper surface with slightly dislodged blocks in intraclast-filled depressions (Fig. 7c). These are classical features of beachrock (Donaldson and Ricketts, 1979), an indicator of paleostrand position.

Figure 5. Lithology in section 1. The scales in photos A, B and E are in centimetres. **A.** Laminated to thin bedded siliciclastic siltstones and minor mudstones from the top of the Quartet Group. **B.** Interbedded dolosiltite-lutite (darker and recessive) and siliciclastic siltstones-mudstones of the basal GLG unit (G-TR of Delaney). **C.** Coarsening and thickening upward sequence in the central part of the lower Gillespie Lake Group. Jacob's staff rests on thick beds of the upper part of the cycle. **D.** Parallel, continuous, planar-laminated dolosiltite-lutite of the lower Gillespie Lake Group. **E.** Elongate, bedding-parallel carbonate nodules in a cream-coloured dololutite matrix. This distinctive rock type alternates with lenticular bedded dolosiltites and dololutites in the upper part of section 1.

Stromatolite types vary. Domal stromatolites are generally linked, approximately circular in plan view with a diameter up to 3 m (most <1 m), have low synoptic relief (10-50 cm), and exhibit smooth to slightly crenulate internal laminae (Fig. 7d). Columnar stromatolites tend to be of the same type within individual shallowing-upward sequences, although some variation laterally and vertically in a sequence was rarely observed. Most are close-packed, with non-branching or rarely branching columns that are parallel to slightly divergent. Many of the columns are preferentially inclined (Fig. 9b). Many are club-shaped, increasing in size upwards (Fig. 9c); coalescence and marked divergence of branched columnar types were also noted. Rare conical columnar stromatolites (conophyton, Fig. 9d) occur as close-packed non-radiating forms which are overlain by domal stromatolite biostromes. These distinctive forms were only observed in a 30 m interval of section 3.

The overlying unit 4 also contains stromatolites, but most forms are different from those seen in the GLG. Some of the unit 4 stromatolites we observed are very irregular, with marked changes in diameter, and complex radiation and coalescence (Fig. 9e). This tungussiform stromatolite may be sufficiently distinct from GLG types to serve as a basis for distinguishing this map unit regionally. More study of the stromatolite forms in both units is required to confirm this.

DISCUSSION

This reconnaissance of the stratigraphy of the GLG shows it to be generally similar to that in the type area (Delaney, 1981, 1985). Major differences are the total thickness of the group and the degree to which internal subdivisions can be recognized. The top contact of the GLG is in fault contact with the overlying Unit 4 throughout the map area. However, the upper GLG in the map area is correlated with the top unit (G-7) of the GLG in its type area, and is of about the same thickness as measured by Delaney for that unit. Thus the total measured thickness of the GLG obtained from combining sections 1 and 2 of about 1.2 km is probably close to the true thickness in this area. Although this is considerably thinner than the greater than 4 km or more Delaney cited for the GLG about 50 km to the north, Delaney recognized abundant faulting and speculated that structural complications might be important in the type area. In view of our data, we think Delaney's concerns about structural complications were warranted, and that much of the thickness discrepancy reflects structural duplication in the type area. The pervasive grey-to-orange weathering colour of the GLG and lack of internal markers make recognition of fault-repetition difficult where exposure is not ideal.

Most of the seven major GLG subdivisions of Delaney (1981, 1985) may not be confidently recognized. G-TR, the basal unit transitional from the underlying Quartet Group, is present. The upper stromatolite rich unit, G-7, is also easily recognized. The cyclical coarsening upward sequences in the central part of section 1 are probably the equivalent of parts of Delaney's unit G-4. The rest of the lower GLG lithologies contain features which allow a rough subdivision

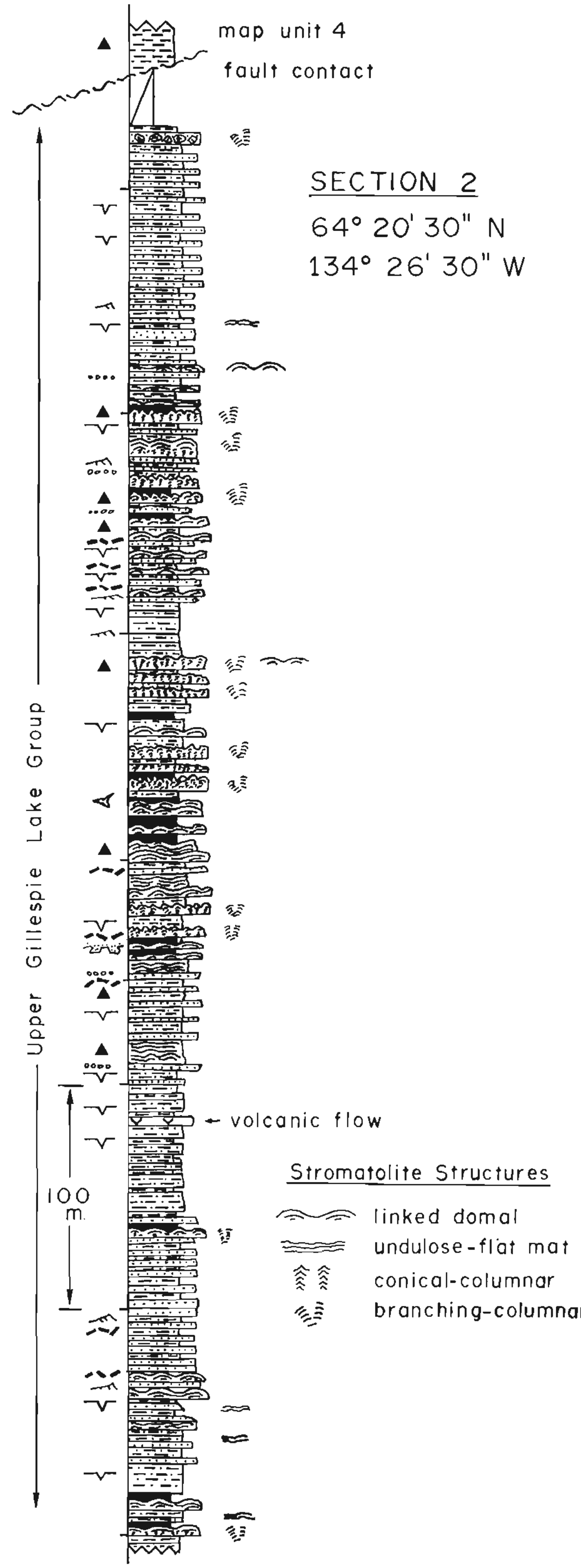


Figure 6. Section 2, exposed in the northwest-trending ridge of Mt. Good, shows the stratigraphy of the upper GLG in this area. This section includes the equivalent of Delaney's unit G-7. The volcanic flow about 180 m above the base of the section is inferred to be equivalent to the flow at the top of section 1.

into some of Delaney's other units, but these units are difficult to discern and much thinner then reported in their type areas.

The lower exposures of the GLG weather recessively, with large sections covered by talus or grassy slopes. For most purposes, a simple division of the GLG into lower and upper units is proposed. The lower unit of such a division comprises the deeper shelf (subtidal, probably most sub-wave base) siliciclastic rocks and dolostone, with the latter increasingly dominant upward. This incorporates Delaney's units G-TR and G-1 to G-6, although his unit G-5 (not recognized in our study area) locally contains columnar stromatolite beds and stromatolitic pebble conglomerate. The upper unit of the proposed two-fold division comprises shallow shelf to supratidal dolostone, rich in stromatolites. This simple two-fold subdivision is readily recognizable and mappable at 1 :50 000 or more detailed scales, with the more resistant upper GLG forming a readily identified regional map unit. Although internal facies may change laterally, the two units appear to persist regionally.

Where greater detail is necessary, for example in property evaluation, the additional subdivisions of Delaney could be useful, although Delaney recognized that some divisions are probably local facies which will not be regionally persistent. The subtle faulting present in the GLG also dictates caution. Several features in the upper unit of the GLG could serve as local markers for detailed mapping by allowing recognition of fault offsets and fault repetition in areas of poor exposure. These features include:

Well-preserved oolite-pisolite beds

Most oolitic beds in the upper GLG are dolomitized to the point where the oolitic texture is preserved only in early silicified lenses or irregular patches. These silicified areas form less than < 10 % of most beds and are absent in many. One remarkably well preserved bed of reverse-graded oolites/pisolites was observed (Fig. 7a). The oolites/pisolites were individually silicified before dolomitization, accounting for their relatively pristine preservation. This distinctive bed is present both at Mt. Good and 7 km to the west at the Blende Pb-Zn-Ag property (Fig. 2; Roots, 1990). The beds at these two sites are of similar thickness and have virtually identical textures, grading and silicification patterns. We conclude that they are the same bed.

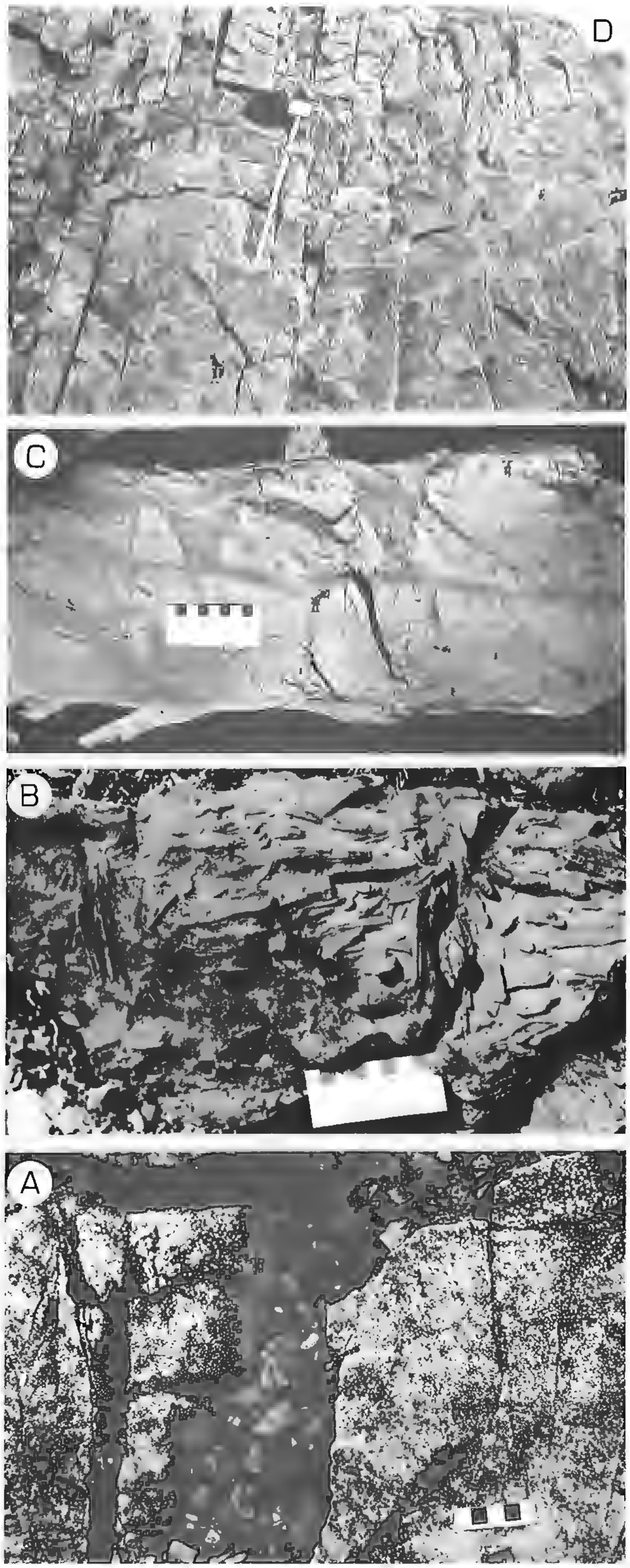


Figure 7. Features of section 2. The scales in photos A-D are in centimetres. **A.** Oolite-pisolite bed that is reverse graded, about 30 cm thick and unusually well-preserved. Dolomitization has generally obscured the original oolitic texture of dolarenite beds, which is displayed only in chert nodules, in irregular silicified patches, or in extensively silicified beds such as in this distinctive marker. **B.** Bedding view of beach rosettes showing closely packed tabular dololutite clasts that are vertically oriented and arranged in a crude polygonal pattern. **C.** Beachrock (view perpendicular to bedding) showing irregular, erosional upper contact and several large clasts of eroded dolosiltite (note correspondence of lithology to flanking source bed) in an intraclast-filled depression. **D.** Laterally linked domal stromatolites at the top of a shallowing-upward sequence. Jacob's staff (1.5 m long) is left of the inflection zone between two domes that have a synoptic relief of about 50 cm, cumulative heights of >4 m, and apparent diameters of >3 m.

Beachrock, beach rosettes

Beachrock and beach rosettes are easily recognizable features which indicate shoreline position. These structures may have been common in the cyclically varying shallow marine-to-supratidal environment of upper GLG deposition. However, the preservation potential of these features is low and their occurrence may reflect an unusual (storm?) event which caused sudden burial. Thus, well-preserved examples may be locally, or even regionally persistent and useful as a marker horizon.

Stromatolites

The variety of stromatolite types in the GLG should be particularly useful for local correlations. The conical stromatolites (conophyton, Fig. 9d) are restricted to a 30 m thick interval in section 3. These easily recognizable forms probably represent a subtidal stromatolitic horizon (cf. Donaldson, 1976; Hoffman, 1976). Overall vertical variations in column types may be locally present and correlatable. However, extensive study of the stromatolites in this unit is required to establish a sound basis for correlation.

Cyclicity of shallowing-upward sequences

The cyclic repetition of shallowing-upward sequences may reflect allocyclic controls (predominantly eustatic sea-level changes and episodic basin subsidence), autocyclic controls (repeated sediment progradation), or some combination (James, 1984). Studies which propose eustatic sea-level changes as the major control demonstrate a regional correlation of the overall pattern of cyclicity (based on changing vertical thicknesses of stacked sequences). Recent examples include Koerschner and Read (1989) and Grotzinger (1986). It may be possible to employ similar correlation techniques with the shallowing-upward sequences in the upper GLG.

Volcanic Flows

Volcanic rocks are known from the upper GLG in several areas (eg. Hart River; Morin, 1979). The thin, discontinuous volcanic bed observed in sections 1 and 2 indicates that volcanism occurred during upper GLG deposition in this area. Single flows or tuffaceous beds, if sufficiently distinctive, could be useful as local markers.

Given the lack of more traditional and reliable marker horizons, the features listed above offer promise for correlation, but should be used with caution. None of the features is necessarily unique to a given part of the stratigraphy and should only be used in conjunction with careful examination of the entire stratigraphic succession.

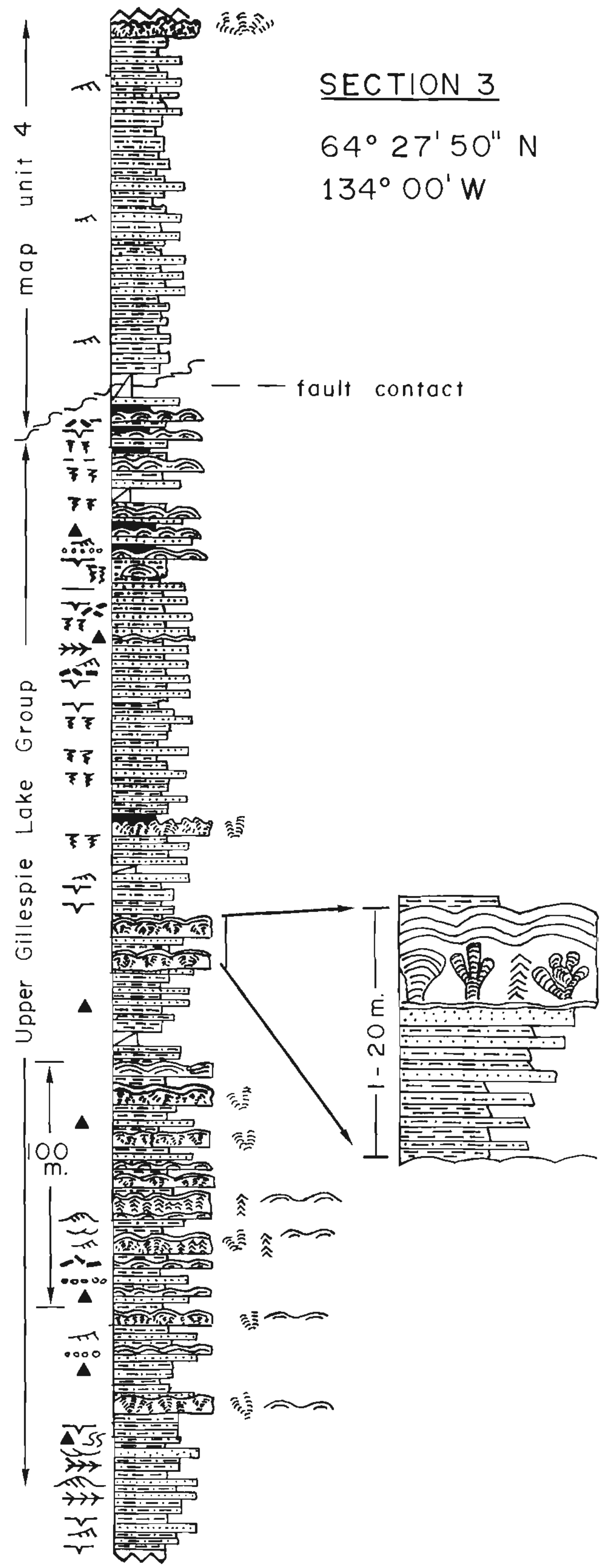


Figure 8. Section 3, near the eastern edge of map area 106D/8, north of Rackla River. This section includes a part of the upper GLG.

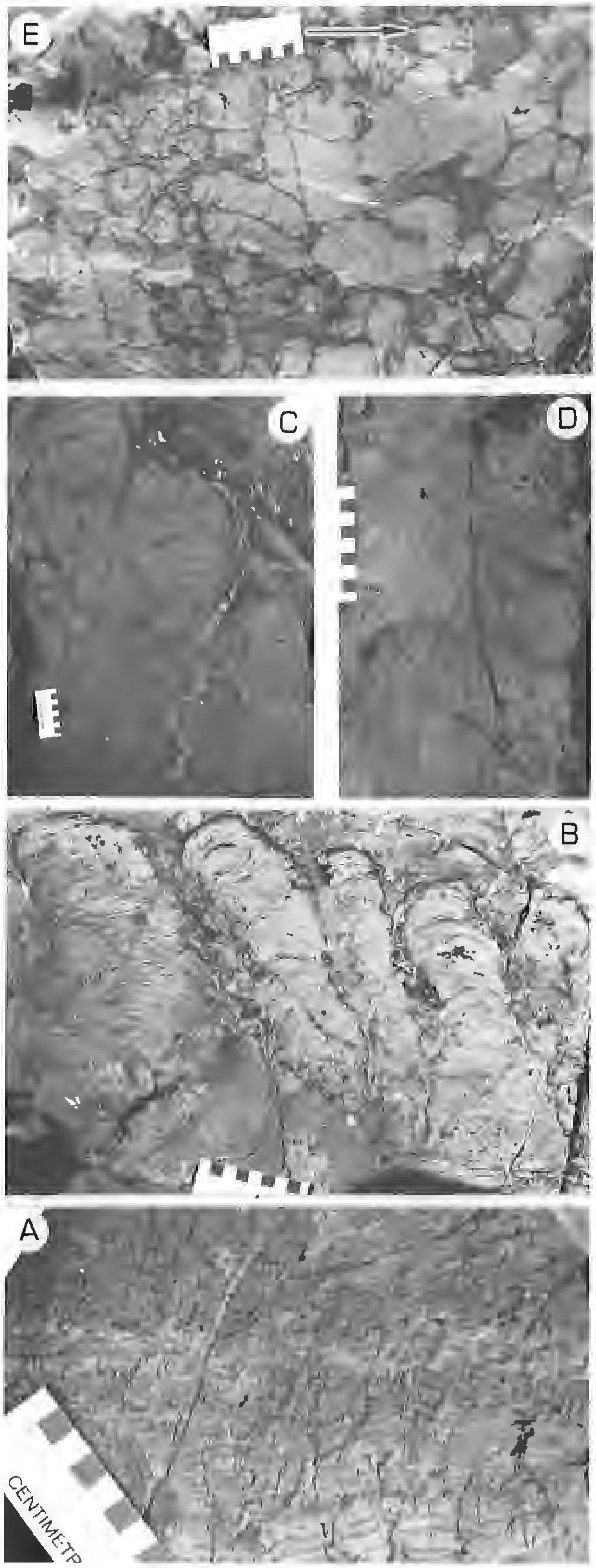


Figure 9. Features of section 3. The scales in all photos are in centimetres. All photos are views perpendicular to bedding with stratigraphic tops towards the top of the photo unless noted. **A.** Molar tooth structure in dolosiltite-lutite, consisting of irregular vertical ribbon structures and less common blob structures (terminology of O'Connor, 1972). **B.** Columnar non-branching stromatolites, grading from vertical at the base to increasingly inclined upward. **C.** Columnar stromatolites displaying club-shaped morphology and a upward increase in diameter. **D.** Conical stromatolite (conophyton). **E.** Complexly radiating, branching and coalescing stromatolites (some are tungussiform) from unit 4, overlying the GLG in section 3. Stratigraphic tops to the right (arrowed).

ACKNOWLEDGMENTS

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Stratigraphic setting of late Paleozoic and Mesozoic fossils, McGregor Plateau, McLeod Lake map area, British Columbia

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Struik, L.C., Stratigraphic setting of late Paleozoic and Mesozoic fossils, McGregor Plateau, McLeod Lake map area, British Columbia; in Current Research, Part E, Geological Survey of Canada, Paper 90-1E, p. 55-58, 1990.

Abstract

Conodonts and pelecypods from limestone and chert in a composite stratigraphic succession (<900 m thick) are dated as Triassic, Permian, Early Carboniferous, Late Devonian, and Middle Devonian. Although the succession has some limestone and basalt it consists mainly of sandstone, cherty tuff, and argillite.

Résumé

L'analyse des conodontes et pélécypodes provenant du calcaire et du chert d'une succession de couches composite, a établi qu'ils datent du Trias, du Permien, du Carbonifère inférieur, du Dévonien supérieur et du Dévonien moyen. La succession a un peu de calcaire et basalte mais se compose principalement de grès, de tuf cherteux, et d'argilite.

INTRODUCTION

Last year, I described three sequences of rock from the east side of the McLeod Lake map area (NTS 93J) (see Fig. 1 and 2 Struik, 1989). They were grouped into the Devonian, Silurian, and Cambrian-Precambrian. Since then, rocks of the "Devonian" sequence have yielded conodonts and pelecypods. The age of those fossils spans the Middle Devonian through Triassic (M.J. Orchard and E.T. Tozer, pers. comm., 1989). This note assigns these new age determinations to the stratigraphic sequence described in Struik (1989, p. 120-121) and elaborates on some of the stratigraphic relationships (Fig. 3).

TRIASSIC ROCKS

East of Fishhook Lake (Fig. 4) several outcrops of limestone and sandy limestone have yielded pelecypods, of which one collection was identified by E.T. Tozer as *Monotis* (pers. comm., 1989). That collection comes from 12 m of section of mainly limestone exposed in a cut 1.5 km southeast along a forest road from Fishhook Lake Recreation Site. At the section, the limestone beds are 5-30 cm thick and are internally laminated. The irregularly spaced but even laminations are defined by variations in orange weathering silt and fine sand content. Rare cross laminations and load casts interrupt the even layering. At the base of the section are thin interbeds of dark grey phyllite. The *Monotis* bearing beds are about 3 m thick and 7 m above the base of the exposure. They are slightly silty and are thinner bedded than the rest of the section.

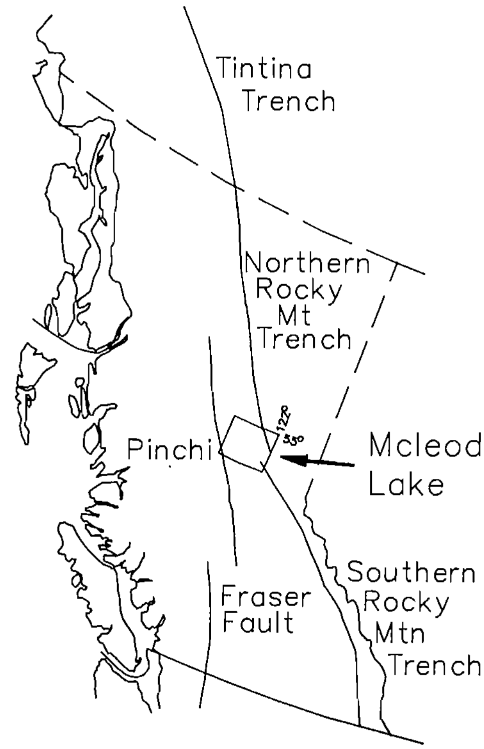

Figure 1. Location of the McLeod Lake map area.

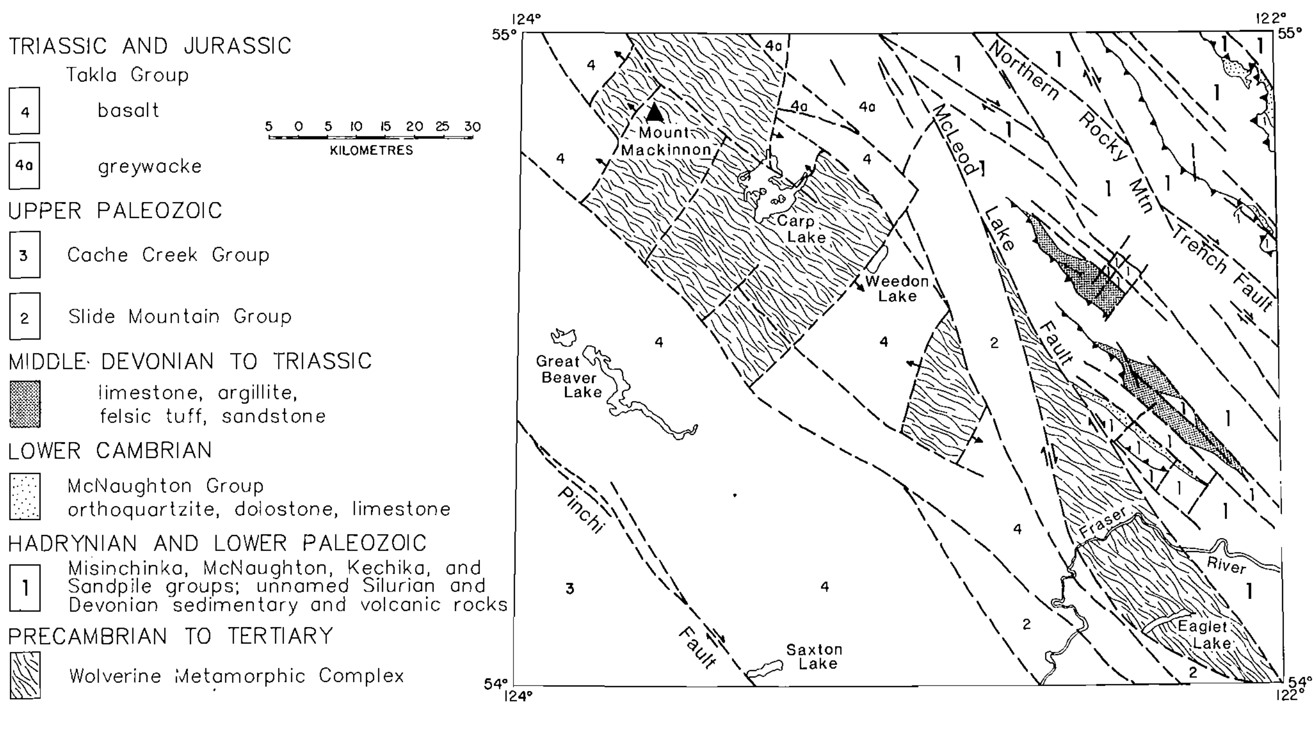

Figure 2. Regional bedrock geology of the McLeod Lake map area.

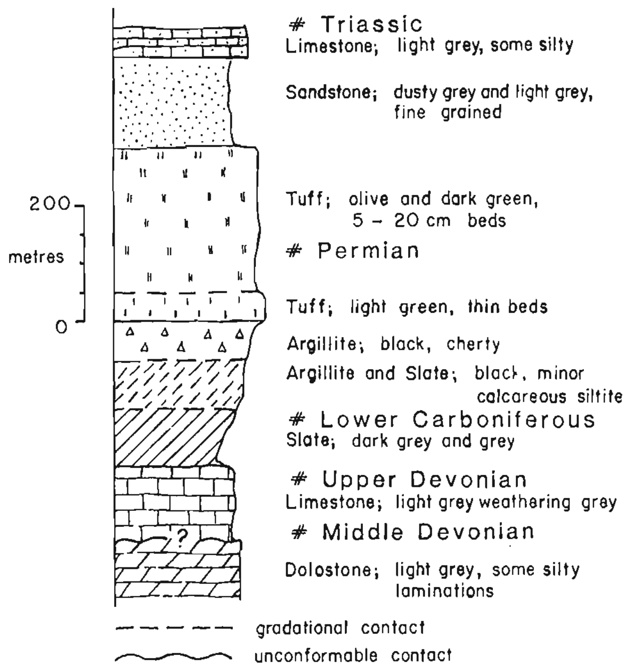

Figure 3. Composite stratigraphic column of upper Paleozoic and Mesozoic rocks from the McGregor Plateau in eastern McLeod Lake map area. The approximate position of the pelecypods and conodonts and their ages as discussed in the text are shown with the hatch marks. The rocks in the sequence and part of the column were described by Struik (1989).

A 3 m wide exposure of similar silty limestone and limestone 600 m east along the forest road also contains pelecypods, but has 10-30 cm beds of brown weathering calcareous fine grained sandstone at the top. Unlike the 12 m exposure, this one has some 20-30 cm thick beds with disrupted laminations. The sandstone of this exposure and the silty limestone to the west both contain irregularly curved, flattened dark tube-shaped forms* (2-20 mm long, 1-3 mm wide) that lie along the bedding and lamination surfaces.

To the southeast and downsection, exposures of ash-grey fine grained sandstone, grey siltstone and minor dark grey

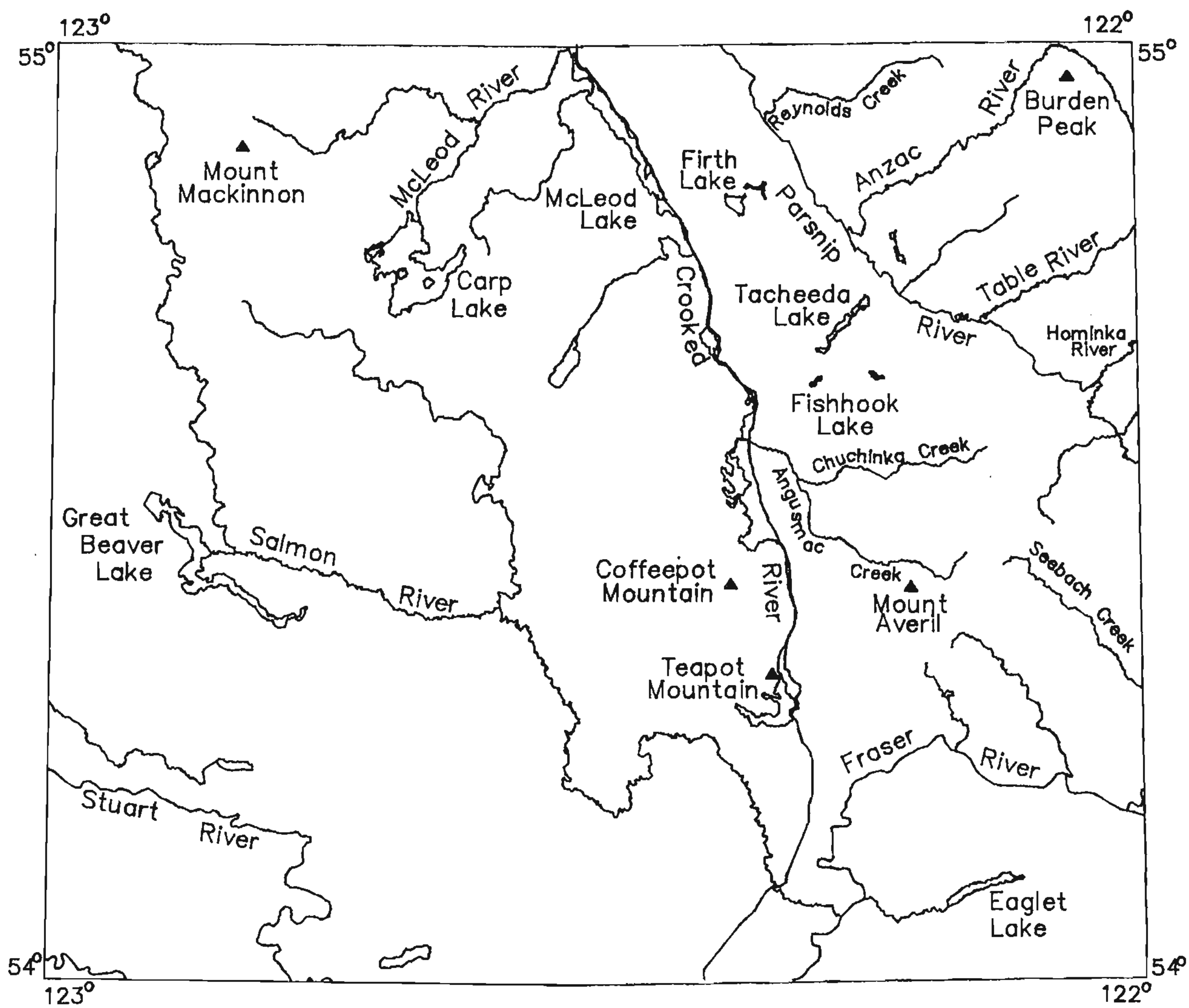

Figure 4. Some geographical features of the McLeod Lake map area that are mentioned in the text.

* *Helminthopsis?* or Cosmorhaphe?

chert resemble the sandstone unit described in Struik (1989, p. 120). Locally those sandstones contain the same flattened tube-like forms found in the limestone to the northwest. Because the Upper Triassic limestone is sandy and silty and locally interbedded with sandstone like that of the underlying sequence, the sandstone unit may also be Triassic.

PERMIAN ROCKS

A single chert sample from an isolated outcrop of chert and cherty tuff south of Chuchinka Creek (Fig. 4) has yielded Permian conodonts (M.J. Orchard, pers. comm., 1989). The chert is on trend with, and resembles, rocks of the tuff unit described in Struik (1989, p. 120).

LOWER CARBONIFEROUS ROCKS

A single limestone sample from a 30 cm bed of dark grey limestone within dark grey and black slate and argillite has yielded Tournaisian conodonts (M.J. Orchard, pers. comm., 1989). The limestone bed is one of three that are spaced some 40-60 cm apart within the slate-argillite unit as described in Struik (1989, p. 120). I do not know the stratigraphic position of the limestone within the unit, but presume it is in the middle of sequence because the rocks interbedded with the limestone are mostly argillite. The bottom part of the slate-argillite unit is mainly grey to dark grey slate; the top part contains dark grey cherty argillite.

DEVONIAN ROCKS

Scattered isolated limestone outcrops throughout the McGregor Plateau have yielded conodonts of the Upper and Middle Devonian (M.J. Orchard, pers. comm., 1989; Pohler et al., 1989). Struik (1989) described these rocks as part of the limestone unit. The limestone is grey and dark grey, finely crystalline and occurs in well defined beds of constant thickness (5-60 cm).

BASALT UNIT

Scattered, discontinuous exposures of basalt are found in the area of the limestone, slate-argillite, and tuff units. The basalt varies widely in texture and amount of calcite amygdules and matrix. It is commonly fragmental but is also pillowed and massive. Muller and Tipper (1969) included the basalts in a Triassic unit and Tipper et al. (1979) mapped them as Silurian.

They have been found directly under the tuff unit (Struik, 1989) and therefore could be Permian or older. In several places outcrops of basalt can be projected along the regional structural trends to lie between the Devonian limestone unit and the Carboniferous slate-argillite unit. The basalt may therefore be Devonian or Lower Carboniferous. However, near Firth Lake (Fig. 4), basalt lies between sandstone of the possible Lower Devonian "tapioca" sandstone unit (Struik, 1989) and Ordovician limestone.

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Metasediments, granitoids, and shear zones, southern Babine Lake, British Columbia

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Struik, L.C. and Erdmer, P., Metasediments, granitoids, and shear zones, southern Babine Lake, British Columbia; in Current Research, Part E, Geological Survey of Canada, Paper 90-1E, p. 59-63, 1990.

Abstract

On the shore of southern Babine Lake, amphibolite, marble, and metachert which may be metamorphosed Cache Creek Group rocks are intruded by diorite, granodiorite, and rhyolite. The diorite is cut by hornblende-biotite granodiorite, which in turn is intruded by quartz-feldspar porphyry rhyolite. On Silver Island, granodiorite, calc-silicate, and amphibolite are extensionally sheared along a gently to moderately south-dipping zone. The shear zone mineralogy is overprinted by saussurite, silica and carbonate alteration, and iron and copper sulphides.

Résumé

Sur le rivage sud du lac Babine, de l'amphibolite, du marbre, et du chert métamorphisé sont recoupés par de la diorite, de la granodiorite, et de la rhyolite. La diorite est recoupée par de la granodiorite à hornblende et biotite, elle-même pénétrée par un microgranite. Sur l'île Silver, affleure une zone cisaillée d'étirement à faible pendage au sud qui recoupe la granodiorite, les calco-silicates et l'amphibolite. La saussurite, la silice, et le carbonate ont altéré les minéraux dans la zone cisaillée, et des sulfures ferriques et cupoiques ont été déposés dans les roches de la zone.

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INTRODUCTION

This reconnaissance work in the Fort Fraser map area (Fig. 1) is a prelude to update mapping (L.C.S.) and possible detailed student thesis work (P.E.) in the 1990s. Around Silver Island on Babine Lake in west-central British Columbia, J.W.H. Monger (pers. comm., 1988) found what appeared to be high grade metamorphic rocks, atypical of areas underlain by the Cache Creek Group (Fig. 1). A possible analogy with metamorphic core-complexes and associated late hydrothermal activity along bounding fault zones prompted us to explore the Silver Island rocks. The area was mapped regionally by Armstrong (1949) who assigned most of the rocks to the Cache Creek Group and Topley intrusions.

REGIONAL GEOLOGY

Southern Babine Lake lies in the belt of Upper Carboniferous and Permian Cache Creek Group rocks west of the Pinchi Fault (Armstrong, 1949; Tipper et al., 1979). The oceanic rocks of the Cache Creek Group extend along much of the North American Cordillera. They were accreted to North America during Triassic and Early Jurassic subduction (Monger, 1984). In the Fort Fraser map area, Cache Creek Group rocks are intruded by the Topley intrusions, and are overlain by Tertiary basalt flows (Armstrong, 1949).

ROCKS OF SOUTHERN BABINE LAKE

Armstrong's (1949) distribution of the Cache Creek Group at southern Babine Lake is like the one we found. His reconnaissance map, however, does not reflect the greenschist to amphibolite grade metamorphism that these rocks record, nor does it show the types of metasediments preserved. Besides the diorite intruded into Cache Creek Group rocks shown by Armstrong (1949) we found quartz-feldspar porphyry and granodiorite. In this report we describe the complexity of rocks and structures along the lake near Silver and Gullwing islands, speculate on the protolith of the metamorphic rocks, and suggest a sequence of structural and metamorphic events. Descriptions are keyed by a symbol to a geological map of southern Babine Lake (Fig. 2).

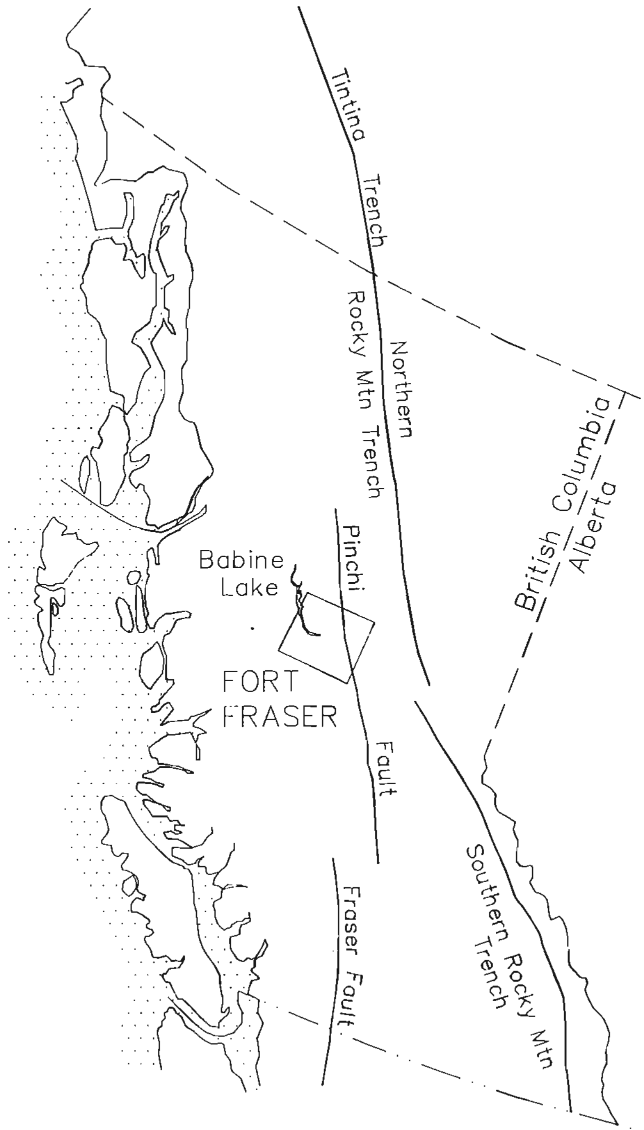

Figure 1. Location of Fort Fraser map area and Babine Lake in British Columbia and relative to some Tertiary dextral strike-slip faults.

Metamorphic rocks

Amphibolite, marble, various types of metachert, and an altered pyroxenite occur along southern Babine Lake. We suspect that they are metamorphosed Cache Creek Group rocks.

Amphibolite is common (localities A1 to A6), and consists of interlaminated plagioclase and foliated hornblende, which are overprinted by hornblende porphyroblasts (1-10 mm). The foliated hornblende is generally finely crystalline (less than 1 mm long). Some plagioclase veinlets contain hornblende porphyroblasts (A2). Calcite veins parallel to the foliation locally have epidote along their margins and, in some places, garnet and diopside (A3). Garnet porphyroblasts also grew in the foliated hornblende (A6).

The amphibolite includes calc-silicate layers (up to 2 m thick), and may be derived from a sedimentary protolith. The sampling area and number of distinguishing features are too small to discard the alternative possibility that part of the protolith was basalt.

Two outcrops of greenstone at the contact of the metasedimentary sequence with a granodiorite may be retrograded amphibolite (A1). They consist of nearly aphanitic greenstone with lit-par-lit injected granodiorite. Quartz and calcite veins parallel layering in the greenstone. In the map area, this is the only example of possible contact metamorphism around the granodiorite.

Dark grey, thinly layered and laminated metachert is strongly deformed (S1-3, S5). On the north (S1) and south (S5) shores of Babine Lake the chert has been intensely sheared, and the shear surfaces define the laminations. Isoclinal folds of the laminations have an axial surface foliation defined by metamorphic biotite (S1, S5). In some places the dark grey metachert displays discontinuous interlayers of black graphitic chert, and small lenses (boudins?) of greenstone (S2).

Marble, found in several localities, is generally light grey to cream coloured, medium to coarsely crystalline, sandy, and banded (2-6 cm thick). The largest exposure is at Boling Point (S4), where the banding is vertical and locally tightly folded. In two places marble is interbedded with metachert (mylonitic, S3, S5). At Boling Point, grey and light khaki chert in contact with, and to the north of, the marble appears to be silicified limestone (S4).

Granitoid rocks

The metamorphic rocks are intruded by diorite, granodiorite, and aplite and quartz-feldspar porphyry dykes. Diorite appears to be the earliest. It consists mainly of biotite, hornblende, and plagioclase, and is melanocratic and equigranular (2-4 mm crystals) (D1). It is intruded by hornblende-biotite granodiorite (D1).

Granodiorite is the most abundant intrusive rock in the area (G1, G2, G3, G4) and is best exposed on Gullwing Island (G1). The granodiorite is equigranular with 1-3 mm

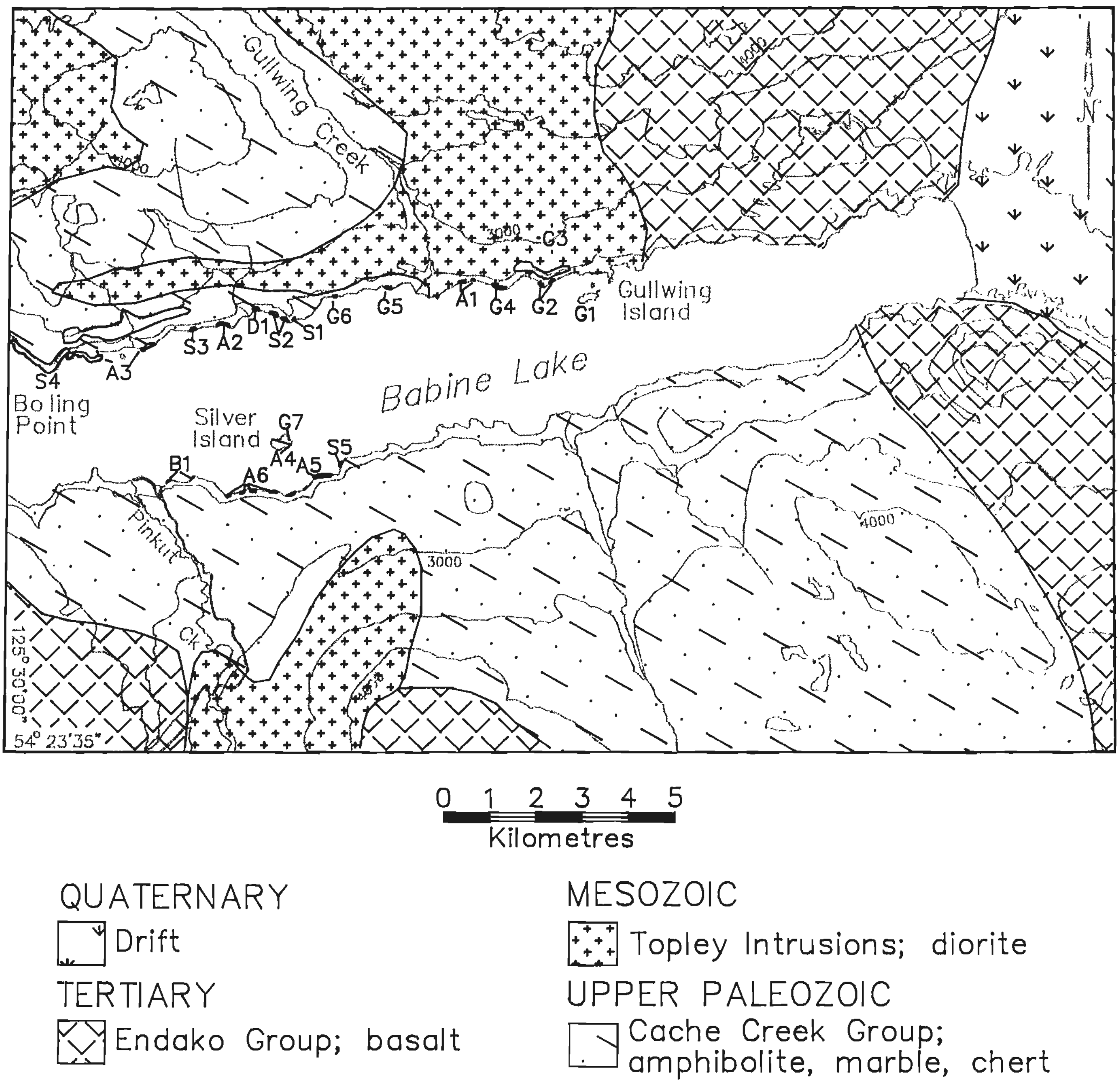

Figure 2. Geological map of the southern Babine Lake area, showing the outcrop areas (e.g. G5) discussed in the text and the general geology as mapped by Armstrong (1949).

crystals, and its main constituents are blocky hornblende (8%), euhedral to subhedral orthoclase (10%), biotite (12%), quartz (25%), and plagioclase (45%). Saussurite alteration has affected about 15-20% of each orthoclase crystal and 5-10% of each plagioclase crystal. Some small hornblende-biotite-rich granodiorite xenoliths are surrounded by finer crystalline granodiorite (G1). Locally the granodiorite is intruded by pink aplite and hornblende-rich mafic dykes.

At G6, the biotite and hornblende are altered to chlorite, and the granodiorite is intruded by calcite vein systems up to 10 cm thick. On Silver Island (G7), a mylonite with micro-augen of plagioclase, orthoclase, and chlorite may be the sheared equivalent of hornblende-biotite granodiorite of Gullwing Island.

Microgranite in one outcrop consists of biotite, quartz, and feldspar microphenocrysts (0.5-2 mm) in an aphanitic pale green matrix (G5). It may be the coarser crystalline equivalent of aplite dykes that intrude the granodiorite and some of the amphibolite on the south shore (A5). The dykes in the amphibolite have microphenocrysts of quartz and feldspar.

Basalt

Basalt agglomerate and breccia near the mouth of Pinkut Creek (B1) is green and purple, and amygdaloidal (zeolites and calcite). The basalt is essentially undeformed, and resembles that of the Hazelton Group (Armstrong, 1949). Amphibolite occurs 0.5 km to the east of the undeformed basalt, with no visible structural or metamorphic gradient between them. Hazelton Group rocks have not previously been mapped in the area, therefore we are reluctant to place much weight on this correlation.

DISCUSSION

We discuss the geological features from oldest to youngest, beginning with the protolith to the metamorphic rocks, and proceeding through the structural and metamorphic history.

Protoliths of the metamorphic rocks

We interpret the marble, sandy marble, and calc-silicate as metamorphosed limestone and impure limestone, because in places they are compositionally layered, and lack the exotic mineralogy associated with carbonatites. The amphibolite may similarly be derived from calcareous rocks because locally it includes calc-silicate lenses. As there are few calc-silicate occurrences, the amphibolite could be derived from basalt flows with interlayered muddy carbonates. Armstrong (1949) does not describe sediments interlayered with the basalt of the Cache Creek Group.

The banded siliceous rocks consist mainly of cryptocrystalline quartz and less finely crystalline white mica, biotite, and plagioclase. In places the banding may be remnant bedding, but thin sections show shear lamellae throughout. These rocks could have been muddy chert, or cherty tuff. As ribbon chert is common in the Cache Creek Group this is the likeliest protolith.

Altered pyroxenite (hornblendite) occurs at the adit on Silver Island. Ultramafic rock is characteristic of the Cache Creek Group.

Shear

Marble and chert at two sites, one on each side of the lake, display a mylonitic foliation (S1, S5). In the marble, wispy laminations typical of mylonite fabric are preserved but the calcite is coarsely crystalline. Chert in contact with the marble has retained its macroscopic and microscopic shear fabric.

Metamorphism and isoclinal folding

Metamorphism outlasted the shear because calcite in marble is recrystallized, and metamorphic biotite is parallel to axial surfaces of isoclinally folded shear lamellae in the chert. Hornblende porphyroblasts in the amphibolite may have formed during the final stages of metamorphism, and the hornblende foliation that they overprint developed during the shear and folding events. Metamorphic mineral assemblages include garnet and hornblende in the amphibolite; diopside, grossularite, epidote, and calcite in the calc-silicate; and biotite in the metachert.

Diorite, granodiorite, and quartz-feldspar porphyry intrusions

Intrusions in the metamorphic rocks truncate the metamorphic foliation. Finely crystalline greenstone at the margin of the Gullwing Island granodiorite, may be the hornfelsed equivalent of amphibolite (A1). Dykelets of granodiorite intruded the diorite, and both include aplite dykes.

Silver Island shear zone

Mylonite on the north side of Silver Island dips 20-30° south, and consists of micro-augen of plagioclase, orthoclase, and chlorite in a matrix of muscovite and polycrystalline quartz ribbons. The protolith of the mylonite could be granodiorite like that on Gullwing Island. At the adit on Silver Island, aplite dykes intrude hornblendite (metapyroxenite), and the foliation and dykes dip north. Between the aplite and hornblendite and the granitoid mylonite, is a zone of intensely folded and sheared granitoid rock, amphibolite, and calc-silicate. At G7, a fold of granitoid rocks in a steeply dipping part of the zone was distorted along a flow axis that plunges 6° east-southeast.

Within the granitoid mylonite 100 m north of the distorted fold, an S-C fabric, pressure shadow tails, and lineations as seen in oriented thin sections show top down to the south displacement. This sense of displacement and the orientation of the lineations are not parallel to the flow axis of the refolded fold.

Hydrothermal precipitation and brittle faults

On Silver Island (Fig. 2), chalcedonic quartz and calcite veins cut metamorphic and ductile shear fabrics. The country rock adjacent to parts of the veins has been bleached. Pyrrhotite, pyrite, and chalcopyrite occur in the veins, and as disseminations in country rock. The alteration, and quartz, calcite, and sulphide precipitation are found elsewhere in the area, particularly in the quartz-feldspar porphyry dykes and amphibolite along the south shore of the lake. Many of these altered areas have been tested with adits in the past.

Faults with broken rock, gouge, and other brittle cataclastic characteristics occur throughout the area. They offset the ductile shear fabric, the metamorphic mineral foliation, the folds, and the granitoid intrusions. Calcite and chalcedonic quartz veins occupy some of the faults. The rocks may therefore have faulted before or during the hydrothermal event, and the faults may have served as conduits for the fluids.

Most displacement is extensional, and on Silver Island and the south shore of Babine Lake many faults dip moderately to gently south. On Silver Island, south-dipping brittle extension faults offset the granitoid mylonite.

CONCLUSIONS

Metamorphic rocks along the shore of southern Babine Lake may have been derived from rocks similar to the Cache Creek Group in the Fort Fraser map area. They were sheared, folded, and metamorphosed prior to the sequential intrusion of diorite, granodiorite, and quartz-feldspar porphyry. Granitoid mylonite on Silver Island may have undergone two phases of shearing; one is subhorizontal, the other down to the south. Hydrothermal precipitation of calcite, quartz, and sulphides may have been concurrent with brittle faulting, some of which is extensional, with top down to the south sense of displacement.

The "Cache Creek Group" rocks were metamorphosed while they were sheared and isoclinally folded. This metamorphism and deformation predates intrusion of granitoid plutons, that do not appear to have imposed large metamorphic aureoles. Although we infer from the degree of metamorphism and from the ductile shearing that the rocks were involved in crustal thickening, it is not clear whether we can infer from the ductile and brittle extensional faults on Silver Island that the region was uplifted by crustal extension along those faults.

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Polyphase tectonic, metamorphic, and magmatic events in the Wolverine Complex, Mount Mackinnon, central British Columbia

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Deville, E. and Struik, L.C., Polyphase Tectonic, metamorphic, and magmatic events in the Wolverine Complex, Mount Mackinnon, Central British Columbia; in Current Research, Part E, Geological Survey of Canada, Paper 90-1E, p. 65-69, 1990.

Abstract

At Mount Mackinnon the Wolverine Complex consists of three main suites of rock: 1) a basement of strongly metamorphosed (amphibolite facies) sedimentary rocks, 2) deformed intrusions of leucogranite, and 3) a dyke complex of microgranite, rhyolite, dacite and basalt. The metamorphism and leucogranitic intrusions formed by deep burial of the sedimentary rocks, probably during Cretaceous and older crustal thickening. The dyke complex formed by uplift and extension of the overthickened crust during the Tertiary. The crust probably extended as pull-aparts between large-scale dextral strike-slip faults.

Résumé

Dans le secteur du mont Mackinnon, le complexe métamorphique de Wolverine comprend trois principaux types de roches: 1) un socle métasédimentaire fortement métamorphisé dans des conditions de faciès des amphibolites, 2) des intrusions déformées de leucogranites et, 3) un complexe filonien comprenant des intrusions de microgranites, de rhyolites, de dacites et de basaltes. Cet édifice résulte d'une histoire compliquée débutant par l'enfouissement profond des roches sédimentaires dans la zone à sillimanite. L'enfouissement est probablement associé à un fort épaississement de la croûte terrestre survenu avant ou durant le Crétacé. Ultérieurement, des tectoniques de distension progressives ont été à l'origine de la remontée des roches métamorphiques lors du Tertiaire. Les distensions ont probablement eu lieu au sein de domaines en "pull-aparts" dans un contexte de décrochements dextres à grande échelle.

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INTRODUCTION

The Wolverine Metamorphic Complex (Armstrong, 1949; Roots, 1954; Parrish, 1976; Mansy, 1986; Struik, 1988; Struik and Fuller, 1988; Struik, 1989) underlies a large area of central and north-central British Columbia between the Pinchi and Northern Rocky Mountain Trench-McLeod Lake faults (Fig. 1). The complex consists of metamorphosed plutonic and sedimentary rocks and unmetamorphosed igneous intrusions. Within the McLeod Lake map area, the Wolverine Complex is separated from the overlying low grade rocks of the Takla and Slide Mountain groups by gently to steeply dipping extension faults and steeply dipping strike-slip faults (Fig. 2; Struik and Fuller, 1988; Struik, 1989).

The southern part of the Wolverine Complex is best exposed at Mount Mackinnon in McLeod Lake map area (Fig. 2). It is one place where detailed mapping has given us a better understanding of the chronology of the tectonic, metamorphic, and magmatic events within the complex. The area surrounding the Mackinnon ridge system is drowned in Quaternary sediments that mask the relationships between the Wolverine Complex and its low grade cover.

GEOLOGICAL FRAMEWORK AT MOUNT MACKINNON

The Wolverine Complex at Mount Mackinnon comprises three suites of rock. From oldest to youngest they are: highly metamorphosed sedimentary and igneous rocks; deformed granite; and a dyke complex of microgranite, rhyolite, and basalt (Fig. 3).

Metamorphosed rocks

The metamorphic rocks were deformed under amphibolite to greenschist conditions. They include coarsely crystalline paragneiss, amphibole-garnet phyllitic quartzite, garnet-bearing micaschist, and minor amphibolite. These lithologies resemble other metasedimentary sequences described elsewhere from the Wolverine Complex (Parrish, 1976; Mansy, 1986; Ferri and Melville, 1988; Struik, 1989). Carbonate rocks typical of the Wolverine assemblage at other places are missing from the sequence at Mount Mackinnon. Metasedimentary rocks of the complex are probably part of the Precambrian to lower Paleozoic sequence deposited on the western edge of the North American continental margin (Parrish, 1976).

The metasediments are intruded by granitic aplite and coarsely crystalline muscovite or biotite pegmatite. These intrusive rocks have clearly been deformed together with the metasedimentary sequence during amphibolite facies metamorphism, but it has not been determined whether they are older or the same age as the metamorphism. The aplite and pegmatite are similar to large bodies of garnet-bearing pegmatitic orthogneiss found in many areas of the Wolverine Complex (Parrish, 1976; Mansy, 1986; Ferri and Melville, 1988).

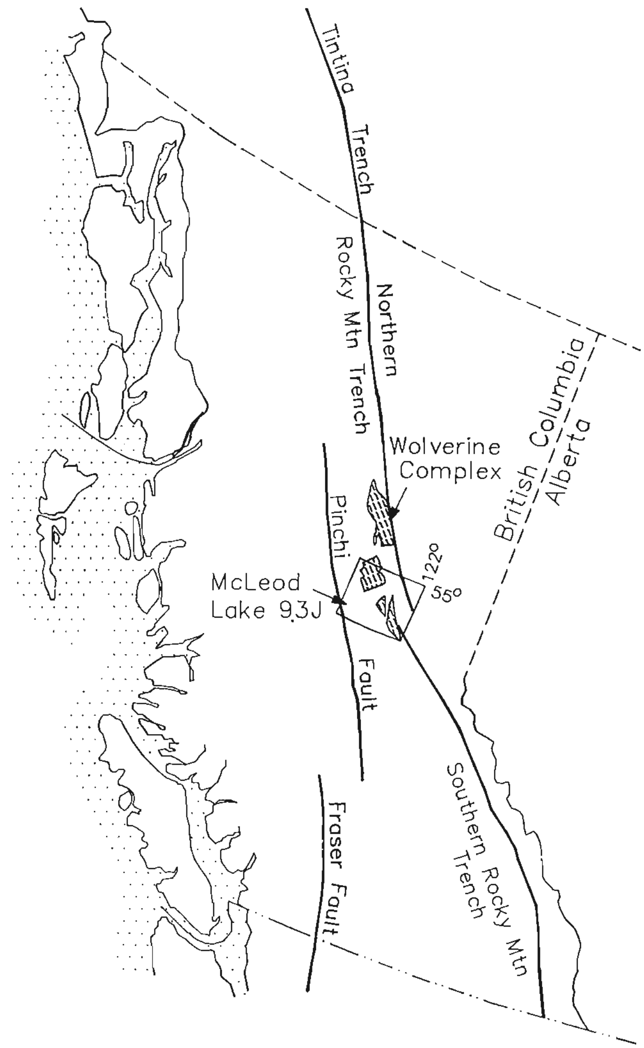


Figure 1. Distribution of the Wolverine Metamorphic Complex and some dextral strike-slip faults in British Columbia.

Deformed granitic intrusions

These finely crystalline leucogranites intruded many areas of the high grade metamorphic rocks (Fig. 3). Mafic minerals in the granite include biotite, muscovite, and amphibole. The leucogranite was deformed by flattening and stretching (clear plano-linear fabric). It includes rafts of older metamorphosed rocks (metasediments with aplito-pegmatitic intrusions) which were folded and foliated under amphibolite facies conditions before they were engulfed by the granite.

Dyke complex

Dykes of microgranite, rhyolite, dacite, and basalt intruded both the metamorphic sequence and the deformed granite (Fig. 3). The thickness of the dykes ranges from a few centimetres for the basaltic dykes to several decametres for the microgranitic dykes. The dykes have chilled margins and are concentrated along the margins of older dykes or crosscut them obliquely. They are dominantly subvertical and strike at 20-60°.

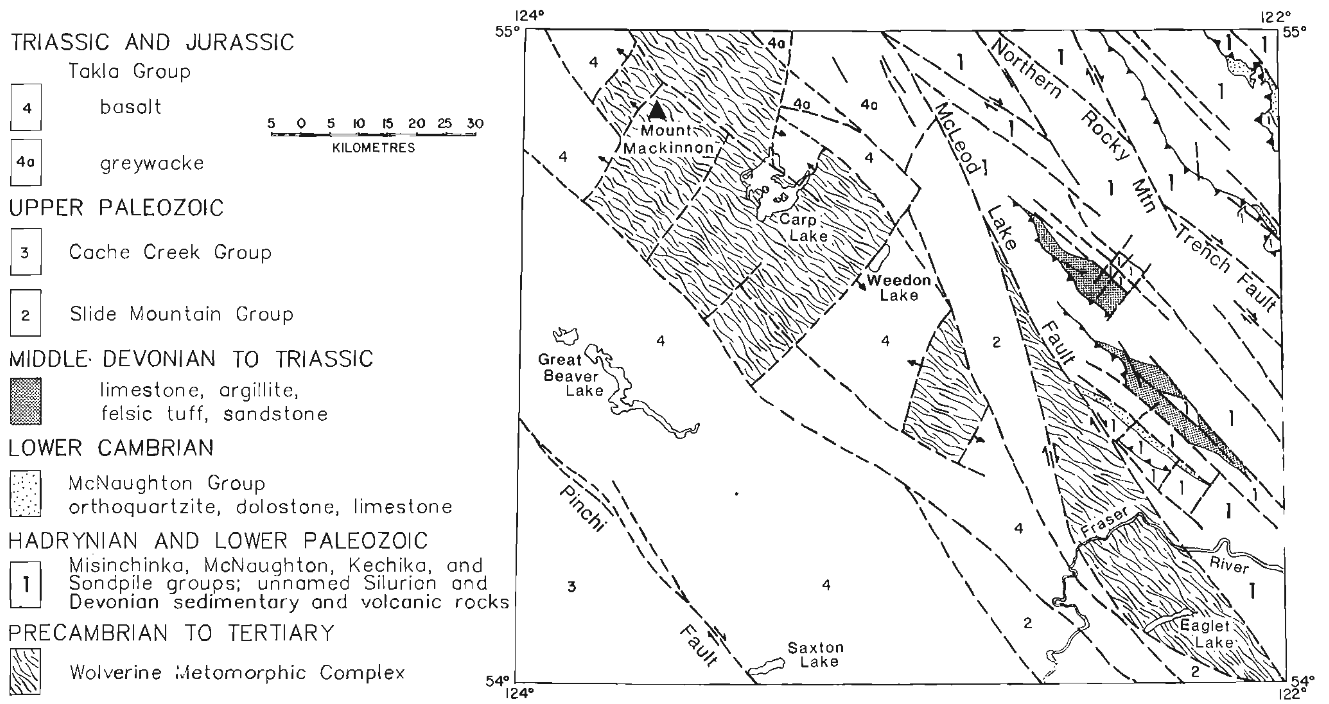


Figure 2. Map of some of the geographical and geological features of McLeod Lake map area that are mentioned in the text.

Micrograniticdykes have hornblende and megacrysts of feldspar up to several centimetres long. Locally these dykes show a subvertical foliation of probable magmatic origin, which can be related to the final emplacement of these rocks. These dykes are generally thick (up to hectametres) and laterally continuous (Fig. 3).

Ubiquitous rhyolite and dacite dykes are quartz-feldspar porphyries that have phenocrysts several centimetres long. They strike 40-60°, are from a metre to a decametre wide and intrude the metamorphic rocks, the deformed granite, and the microgranite.

Basalt dykes with phenocrysts of olivine and zoned plagioclase are the youngest intrusives. They cut the other rocks but appear more commonly with the rhyolite and dacite dykes. The basalt dykes strike between 20-40° degrees and are generally from a decimetre to a metre thick. The wider ones on the southwestern slopes of Mount Mackinnon are up to several decametres thick. These rocks resemble those of the Miocene Chilcotin Group.

TECTONIC, METAMORPHIC, AND MAGMATIC EVENTS AT MOUNT MACKINNON: NATURE AND CHRONOLOGY

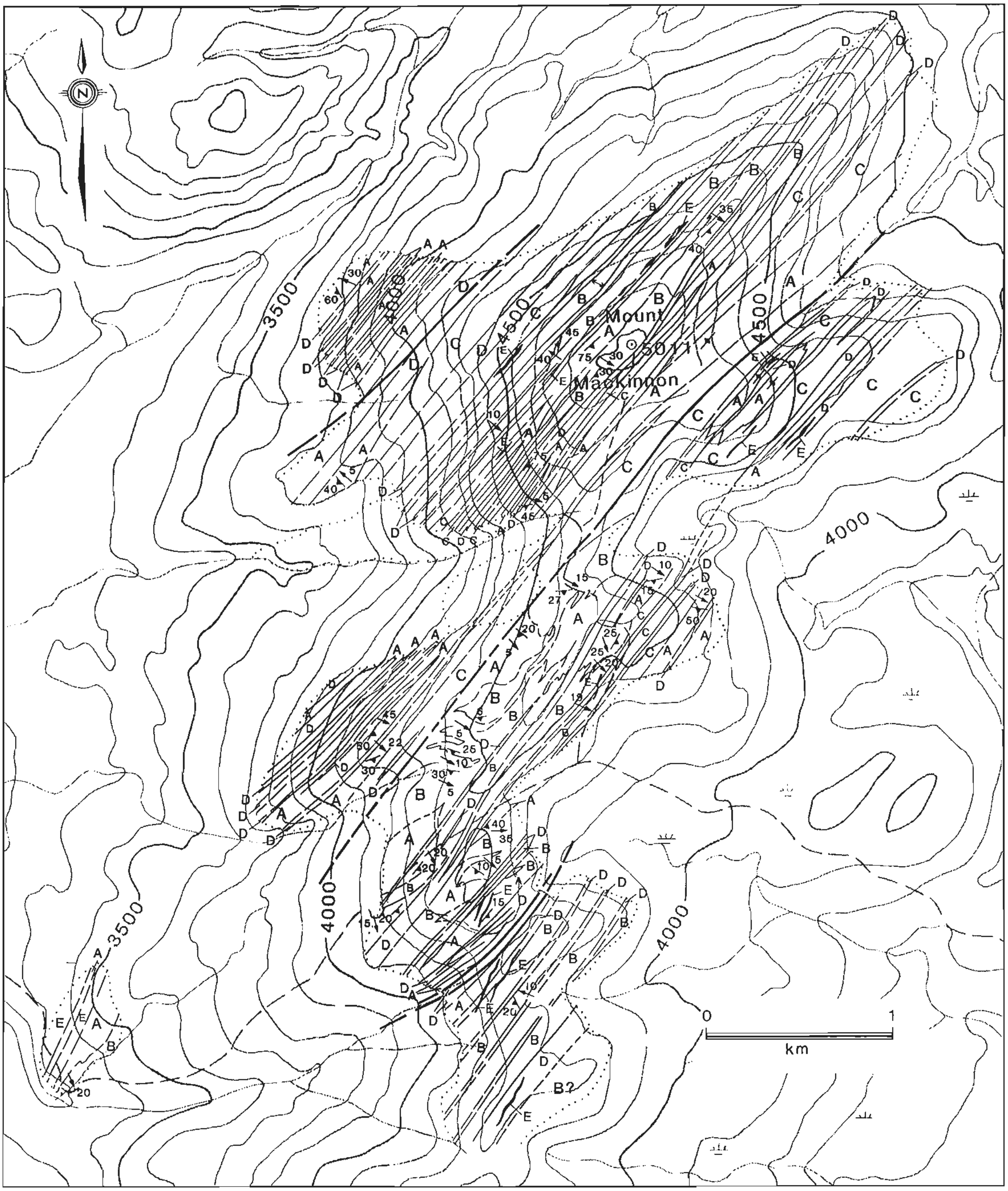
The Mount Mackinnon rocks contain evidence of a prolonged and multistaged tectonic, metamorphic, and magmatic history. From older to younger, five events are distinguished: 1) compression, 2) granitic intrusion, 3) ductile extension, 4) dyke intrusion, and 5) brittle extensional faulting.

An early compressional event is recorded in the metasedimentary rocks by tight folds, amphibole-mica-quartz-feldspar schistosity, relics of a northeast-trending amphibole lineation, and an associated amphibolite facies metamorphism (sillimanite zone). The schistosity, defined in part by the preferred orientation of phyllosilicates and quartzofeldspathic ribbons, has been considerably modified by later events. Isotopic ages from metamorphic rocks of the Wolverine Complex are mid-Cretaceous or older (K-Ar and Rb-Sr ages on metamorphic minerals between 89 and 106 Ma; Wanless et al., 1971; Parrish, 1976).

A granitic intrusion event of primarily leucogranite postdates the amphibolite facies metamorphism. The leucogranite crosscuts the amphibolite facies foliations and lineations, and are themselves deformed. The leucogranite is similar to the Blackpine Lake granitic rocks, known farther north in the Wolverine Complex. The Blackpine Lake granite was dated as early Tertiary (Rb-Sr whole rock isochron of 62 Ma and Rb-Sr mineral isochron of 44.7 Ma by Parrish, 1976).

A ductile extensional tectonic event is characterized by heterogeneous northwest-southeast stretching and local development of mylonitic shear zones. Stretching and mineral lineations that trend 100 to 150° lie in a composite foliation that has been folded in places. The lineations are defined by the stretched quartz and feldspar porphyroclasts, by mineral elongation, boudinage of micas and quartz grains, and by elongation of mineral-filled pressure shadows. The foliation and lineation are the result of flattening and shearing. Textures within the these structures are consistent with extensional ductile deformation due to detachment fault processes.

The dyke intrusion event consists of a dyke complex that crosscuts the metasedimentary and granitic rocks. Some of



E	Basalt dykes
D	Rhyolite, porphyry dykes
C	Microgranite
B	Granite; stretched
A	Paragneiss, micaschist, minor amphibolite and aptite-pegmatite

- - - - Geologic contact (defined, approx.)

. . . . Outcrop limit

——— Fault (defined, approx.)

Foliation

Lineation

Figure 3. Geological map of the Mount Mackinnon area. Contour interval is 50 feet.

the intrusions have deformed and locally tilted to vertical the previous foliations and lineations. Intruded in sequence are microgranite, rhyolite, dacite, and finally olivine basalt. Elsewhere in the Wolverine Complex, most radiometric dates on felsic dykes fall between 37 and 43 Ma (Parrish, 1976, p. 78). The younger basaltic dykes are probably part of the Miocene Chilcotin Group basalts.

A late brittle extensional event is marked by steeply dipping faults that strike 20-40°. They are approximately parallel to the trend of the dykes of the dyke intrusion event. Striae on most of the fault surfaces indicate normal movement consistent with a northwest-southeast brittle extension.

INTERPRETATION AND CONCLUSIONS

The geological history of the Mount Mackinnon rocks can be divided into tectonic events. The first corresponds to the regional high grade metamorphism and compressional deformation, probably the result of collisional processes that thickened the crust with recumbent folds and thrust sheets until Tertiary time. The second corresponds primarily to granitic intrusion, flattening and stretching, and intrusion of a dyke complex. It may be the result of rifting during transform faulting where the crust was thinned in one or more pull-aparts between Tertiary strike-slip faults such as the Northern Rocky Mountain Trench, the McLeod, and the Pinchi (Eisbacher, 1985; Gabrielse, 1985; Struik, 1985a,b; Mansy, 1986; Price and Carmichael, 1986; see Fig. 1). Uplift of the Wolverine Metamorphic Complex within McLeod Lake map area could then be ascribed to thinning of the crust and denudation of the complex as the Takla and Slide Mountain groups (and older rocks?) were stripped off along low-angle normal faults (Struik, 1988, 1989). Such processes have been proposed to explain the crustal extension and uplift of metamorphic complexes elsewhere in the North American Cordillera (Coney, 1980; Armstrong, 1982; Wernicke, 1985; Tempelman-Kluit and Parkinson, 1986; Wilkins et al., 1986; Parrish et al., 1988).

ACKNOWLEDGMENT

Christine Davis drafted Figure 3.

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Stratigraphy and structure, southern Rocky Mountain Trench to the headwaters of the North Thompson River, Cariboo Mountains, British Columbia

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Murphy, D.C., Stratigraphy and structure, southern Rocky Mountain Trench to the headwaters of the North Thompson River, Cariboo Mountains, British Columbia in Current Research, Part E, Geological Survey of Canada, Paper 90-1E, p. 71-80, 1990.

Abstract

The area between the southern Rocky Mountain Trench and the headwaters of the North Thompson River is underlain by early Proterozoic basement orthogneiss and mantling quartzite and marble and an unconformably overlying, complexly folded but unfaulted sequence of lower Windermere Supergroup. Basement orthogneiss cores a regional-scale Middle Jurassic southwest - vergent anticline. The whole sequence overlies in fault contact (Purcell Fault) a marble of presumed Cambrian age. The trace of the Purcell Fault is offset 3-5 km by displacement along the North Thompson-Albreda normal fault and approximately 55 km by combined strike- and dip-slip along a newly discovered strike-slip shear zone and down-to-the-west faults, both of which lie along the eastern boundary of the southern Rocky Mountain Trench.

Résumé

Le sous-sol de la région située entre la partie sud du sillon des Rocheuses et les eaux d'amont de la rivière Thompson-Nord est constitué d'orthogneiss du socle du Protérozoïque inférieur, recouverts de quartzite et de marbre, au dessus desquels repose en discordance une série à plissement complexe mais non faillé du supergroupe de Windermere inférieur. L'orthogneiss du socle forme le coeur d'un anticlinal de vergence sud-ouest, à l'échelle régionale, du Jurassique moyen. Toute la série recouvre par contact par faille (faille de Purcell) un marbre qu'on suppose être d'âge cambrien. La trace de la faille de Purcell est décalée de 3 à 5 km par un déplacement le long de la faille normale de North Thompson-Albreda et est d'environ 55 km par une combinaison d'un rejet horizontal d'un déplacement normal le long d'une zone de cisaillement à rejet horizontal et de failles plongeant vers l'ouest, nouvellement découvertes; cette zone et ces failles longent la limite orientale de la partie sud du sillon des Rocheuses.

INTRODUCTION

Since 1968, when Campbell published the first preliminary map of the Canoe River (1:250 000) map area (Campbell, 1968), about 80 percent of the map area has been mapped in considerable detail by members of various university groups and officers of the Geological Survey. As such, an update of Canoe River is timely and is the focus of my postdoctoral project at the GSC. This summer's fieldwork was directed toward mapping the remaining unexamined regions, re-mapping problematical regions, and tying together areas where continuity was lacking.

The region in the southern Cariboo Mountains between Valemount and the headwaters of the North Thompson River, including the Premier Range and the headwaters and upper reaches of Canoe River (Fig. 1), holds many of the clues to some of the fundamental questions about this portion of the Cordillera. These questions include: (1) What is the role of the Malton and other bodies of basement gneiss in the Mesozoic and Cenozoic kinematic evolution of the region? (2) What happens to displacement on the North Thompson-Albreda normal fault (NTAF) as it approaches the southern Rocky Mountain Trench (SRMT)? (3) What is the geometry and nature of faulting in the SRMT near Valemount? (4) What is the Premier Anticlinorium (Campbell, 1968, 1973) and what happens to it at the deeper structural levels of the southern Cariboo Mountains? (5) What is the nature of the contact between the thick marble of possible Cambrian age in the SRMT around Valemount and the Windermere metasediments to the southwest?

STRATIGRAPHY OF THE PREMIER RANGE AND VICINITY

A surprising and exciting finding of the summer was an unfaulted sequence of the lower Windermere Supergroup from its unconformity with basement orthogneiss and mantling metasediments up to and including the lower Kaza Group (Windermere 'grit' unit). The basal unconformity of the Windermere on crystalline basement is exposed in only one other place in the Canadian Cordillera, in the Deserters Range on the northeast side of the northern Rocky Mountain Trench some 750 km to the northwest (Evenchick et al., 1984).

Orthogneissic basement and mantling metasediments

Walker and Simony (1989) mapped felsic and mafic gneiss south of Canoe River and interpreted it as basement similar to, and possibly correlative with, the Malton gneiss. This interpretation is supported by an unpublished Early Proterozoic U-Pb zircon date.

The gneiss was traced this summer into the Canoe River valley and north into the region west of Valemount (unit 1, Fig. 3). In the Canoe River valley, the gneiss consists of well foliated granitic to quartz dioritic orthogneiss locally with augen of porphyroblastic potassium feldspar. North of Canoe River, the gneiss consists of leucocratic quartz diorite orthogneiss with screens and dykes of amphibolite and garnet amphibolite.

West of Valemount, basement orthogneiss is overlain by micaceous quartzite, quartz-pebble conglomerate, clean, white, locally crossbedded (strained planar tabular sets indicate transport to northeast) quartzite (unit 2, Fig. 3), marble, and calcareous schist (unit 3). Quartzite immediately above the contact is somewhat flaggy and muscovitic but shows little field evidence of larger than background strain; thus, the contact is inferred to be a nonconformity. Orthogneiss, quartzite, and marble are all crosscut by undated garnet amphibolite dykes which do not penetrate the overlying Windermere Supergroup.

Windermere Supergroup

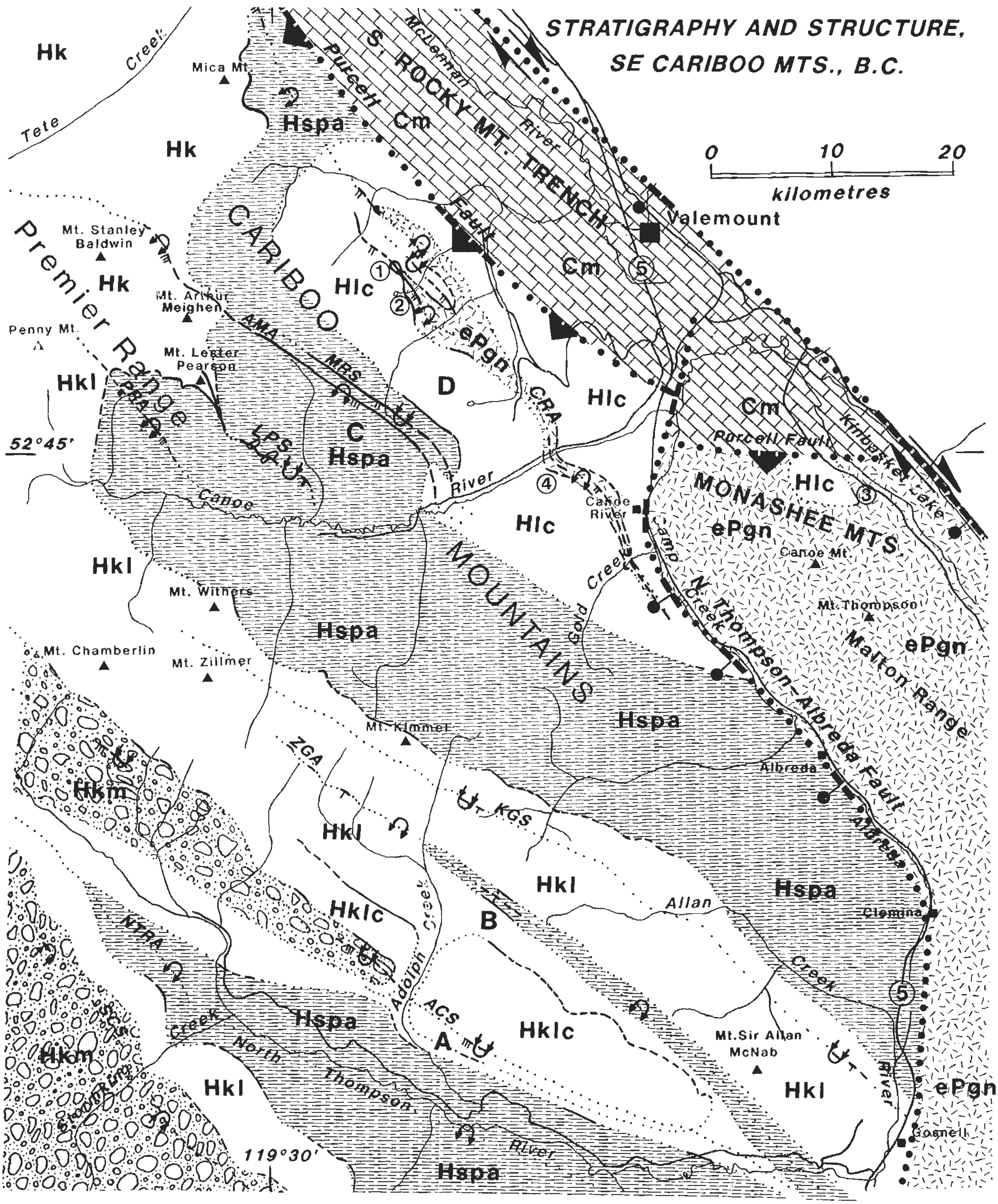
Lower clastic unit (unit 4, Fig. 3)

West of Valemount (locality 1, Fig. 1), quartzite and marble mantling basement orthogneiss are overlain by a pebble to boulder conglomerate. The conglomerate contains rounded to subrounded (elongate parallel to northwest-trending fold hinges) clasts of quartzite, granitic rocks, amphibolite, calcsilicate rocks, and marble — all the lithologies immediately underlying the conglomerate — in a quartz-rich matrix which was probably sand-sized before metamorphism. Clast density varies from very low, where the conglomerate is matrix-supported, resembling a dropstone diamictite, to very high, where the rock is clast-supported. Normal and inverse graded bedding are the only sedimentary structures observed. This unit is inferred to correlate with conglomerate found above basement gneiss south of Canoe River by Walker and Simony (1989).

The basal contact of the conglomerate with underlying rocks is inferred to be an unconformity rather than a fault or shear zone as has been inferred for basement-cover contacts in the Monashee and Rocky mountains (Campbell, 1968; Morrison, 1982; Giovanella, 1967; Oke and Simony, 1981; McDonough and Simony, 1988). This interpretation in supported by the following observations: (1) the conglomerate contains clasts of all underlying lithologies; (2) amphibolite (metabasalt) dykes are truncated at the base of the conglomerate; (3) there is no evidence of brittle sliding at the contact; and (4) crossbedding is preserved in quartzite less than 10 m below the contact and dykes below the contact do not change orientation near the contact, precluding a strain gradient near it.

The unconformity either possesses a certain amount of relief or is angular. Conglomerate is deposited on marble

Figure 1. Stratigraphy and structure of the southeastern Cariboo Mountains. 1. Cambrian marble, 2. middle Kaza Group, 3. carbonate division of lower Kaza Group, 4. lower Kaza Group, 5. semi-pelite and amphibolite unit of Windermere Supergroup, 6. lower clastic unit of Windermere Supergroup, 7. Early Proterozoic orthogneiss and mantling paragneiss. 1-4 are locations discussed in text. A-D are locations of stratigraphic sections shown in Figure 3. Geology southwest of North Thompson River from Pell (1984) and in the vicinity of Mt. Sir Allan McNab from Currie (1988) and Walker and Simony (1989).



Cm 1. Hkm 2. Hklc 3. Hkl 4.

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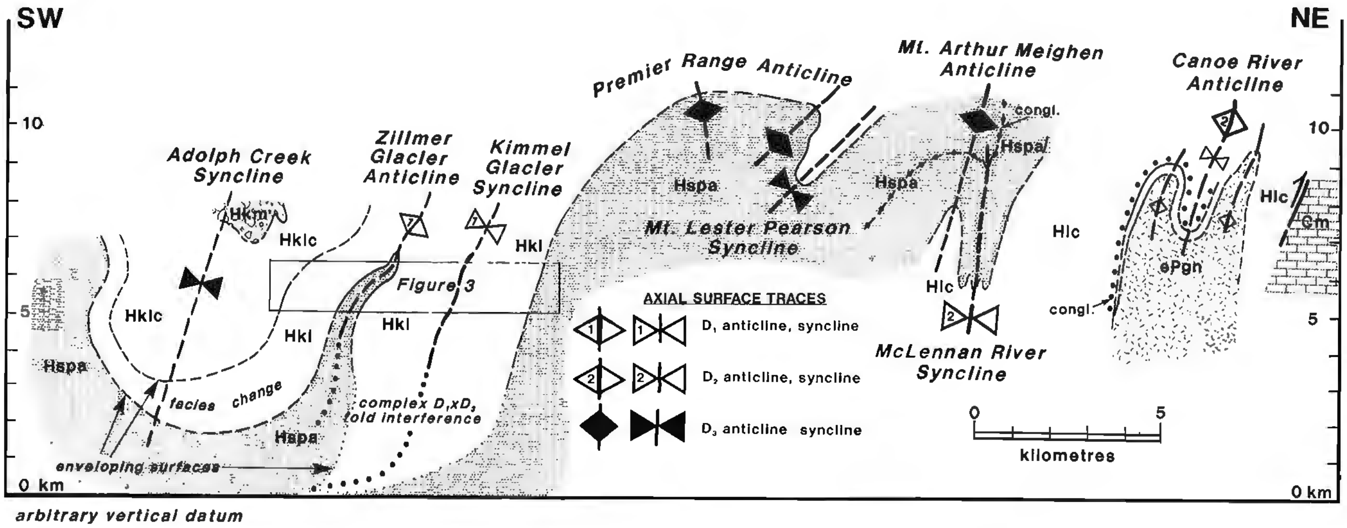

Figure 2. Composite vertical cross-section of southeastern Cariboo Mountains. Section compiled from 8 vertical cross-sections derived from 1:25 000 and 1:50 000 scale mapping north of the Canoe River valley, the area between Mt. Withers and Mt. Kimmel, and the area between the headwaters of Camp Creek and the upper North Thompson River valley. No vertical exaggeration. Legend as for Figure 1.

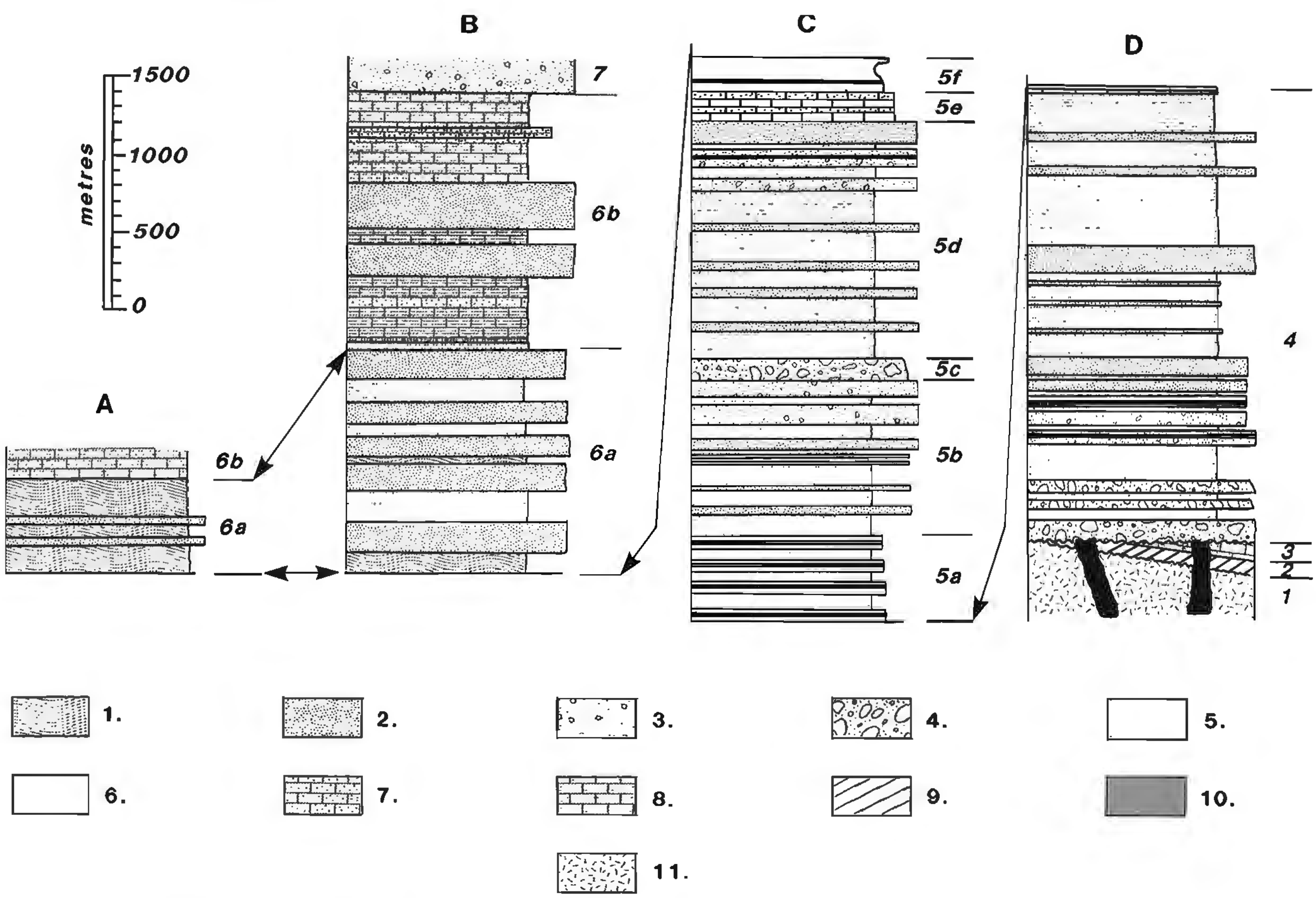

Figure 3. Stratigraphy of Windermere Supergroup, southern Cariboo Mountains. All thicknesses of major units are structural thicknesses measured from cross-sections. Thicknesses of individual beds are schematic, not measured, although relative proportions of various rock types is faithful to observations. 1. graphitic phyllite, 2. predominantly micaceous quartzofeldspathic psammite, 3. predominantly pebbly coarse grained micaceous quartzofeldspathic psammite ('grit'), 4. pebble to boulder conglomerate, 5. predominantly pelitic schist, 6. predominantly flaggy quartzose psammite, 7. sandy, locally pebbly marble, 8. micaceous marble, 9. quartzite, quartzite pebble conglomerate, micaceous quartzite, 10. amphibolite, 11. quartz diorite orthogneiss with pods and screens of amphibolite. Locations of sections shown on Figure 1.

west of Valemount (locality 1, Fig. 1), on quartzite southwest (locality 2) and south (locality 3) of Valemount, and directly on basement orthogneiss south of Canoe River (locality 4) where basement was discovered in the Cariboo Mountains by Walker and Simony (1989).

The lower clastic unit consists primarily of rusty weathering, recessive kyanite-staurolite-garnet-muscovite-biotite-quartz schist with interbedded conglomerate (at the base) and varying amounts of subfeldspathic psammite, subfeldspathic grit, and amphibolite (Fig. 3). Indicators of stratigraphic facing direction such as graded beds and channel truncations show that, except for local minor folds, the sequence is upright everywhere southwest of the contact with the underlying gneiss.

The total structural thickness of this unit exceeds 2.5 km.

Semipelite-amphibolite unit (SPA; unit 5, Fig. 3)

Conformably overlying the lower clastic unit are flaggy subfeldspathic psammite, amphibolite, pelite, conglomerate, marble, and grit of the semipelite-amphibolite unit (SPA) of Simony et al. (1980). The contact is placed at the base of the first beds of white, flaggy, quartzose, subfeldspathic psammite and amphibolite above the coarse grained, grey, thick-bedded psammite of the lower clastic unit. Thin (<2 m) marble and calc-silicate is found locally near this contact, possibly correlative with the 'brown zone' of Simony et al. (1980).

The SPA has been subdivided into 6 regionally mappable units:

Unit 5a. The basal unit of the SPA consists of thin- to medium-bedded, flaggy quartz-biotite-plagioclase psammite, thin (0.5-50 cm), stratiform amphibolite schist, massive, conformable garnet amphibolite, and kyanite-staurolite-garnet-biotite-muscovite-quartz-plagioclase schist (locally with quartzofeldspathic knots and laminae). Subunits of flaggy psammite and amphibolite up to several tens of metres thick alternate with pelitic schist of approximately the same thickness.

Unit 5b. This unit consists of pelitic schist with minor amounts of psammite laced with quartzofeldspathic stringers lending the appearance of migmatite. At the top of this unit psammite coarsens and thickens to become about 70 percent of the sequence.

Unit 5c. Capping the coarsening- and thickening-upward sequence at the top of unit 5b is matrix-supported pebble to boulder conglomerate; thin-bedded, fissile, flaggy, friable, maroon calcareous psammite and marble; and locally micaceous quartzite. Conglomerate clasts are rounded to subrounded and elongate parallel to local northwest-trending fold hinges; clasts include quartzite, marble, calc-silicate rock, and granitoid rocks. This unit was traced for several kilometres across and along strike and is inferred to correlate with an SPA conglomerate mapped by Walker and Simony (1989) south of Canoe River.

Unit 5d. This unit consists of a sequence of rusty psammite, pelitic schist, grit, and rare amphibolite. Amphibolite within this unit is inferred to be intrusive based on local crosscutting relationships and porphyroblast-rich margins possibly indicative of chilled margins.

Unit 5e. This unit consists of a sequence of flaggy, quartzose and calcareous psammite, thin marble, and minor pelitic schist and amphibolite. Although marble is not everywhere present, this calcareous unit can be traced for several tens of kilometres within the Premier Range.

Unit 5f. Capping the SPA is a thin (up to 400 m thick) sequence of thin- to medium-bedded, rusty-weathering, flaggy micaceous psammite, pelitic schist, and amphibolite. In the headwaters of the North Thompson River, the upper contact of the SPA with the lower Kaza Group is marked by a thin whitish-grey marble; this marble is absent to the east.

The thickness of SPA east of Mt. Sir Lester Pearson is approximately 3.5 km.

Kaza Group

Conformably overlying the SPA is pelite, psammite, grit, graphitic phyllite, marble, and calc-silicate rock considered to be correlative with the lower and middle Kaza Group of Pell and Simony (1987). In the headwaters of the North Thompson River, the base of the Kaza Group is mapped at the base of a distinctive 600 m thick, dark grey graphitic phyllite and brown psammite unit which overlies thin marble capping a sequence of flaggy quartzose psammite, pelite, and amphibolite of the upper SPA. In the Premier Range and the region of the headwaters of the Canoe River, the base of the lower Kaza Group is mapped at the base of the first medium grey pelitic schist overlying quartzose subfeldspathic psammite, rusty pelitic schist, and amphibolite of the upper SPA and underlying a thick monotonous sequence of grey, locally graphitic pelitic schist or phyllite, psammite, and quartz and feldspar pebble grit.

The southwest-dipping, stratigraphically upright SPA-Kaza Group contact has been mapped from Mt. Sir Lester Pearson around the head of the Canoe River valley, through Mt. Carpe (about a kilometre northeast of Mt. Withers), and into a contact interpreted by Currie (1988) and Walker and Simony (1989) as the inverted basal contact of the SPA with the stratigraphically older lower grit unit. Upright graded beds and channel truncations observed along this contact near Mt. Sir Lester Pearson, south of Penny Mountain, near Mt. Withers, and in a branch valley north of the west branch of Kimmel Creek support the correlation of these gritty rocks overlying the SPA with the Kaza Group.

Lower Kaza Group (unit 6a, Fig. 3)

From Mt. Sir Lester Pearson around the head of the Canoe River to Mt. Zillmer, the lower Kaza Group is composed of 1.2 km of grey, locally graphitic, muscovite-biotite-garnet-quartz-plagioclase(-kyanite-staurolite) pelitic schist, subfeldspathic psammite, and graded quartz- and feldspar-pebble grit. In the headwaters of the North Thompson River, the lower Kaza Group consists of 600 m of graphitic phyllite, brown psammite, and minor lenticular bodies of quartz-and feldspar-pebble grit. These observations imply that the lower Kaza Group changes facies across the region (Fig. 2, 3).

Carbonate subdivision of lower Kaza Group (unit 6b, Fig. 3)

A unit consisting of fissile, brown, micaceous marble, quartzite-clast marble conglomerate, calcareous schist or phyllite, psammite, grit, and near the top, massive, sandy grey marble, is found above graphitic phyllite and brown psammite on the southwest limb of the Adolph Creek Syncline (ACS, Fig. 1, 2) and above grit, psammite, and, locally, graphite schist or phyllite on the northeast limb of the syncline. This unit is correlated with the carbonate subdivision of the lower Kaza Group mapped by Pell (1984) south of the North Thompson River (Fig. 1). Northwest of Adolph Creek, the total thickness of calcareous rocks at the top of the lower Kaza Group is about 1 km.

Middle Kaza Group (unit 7, Fig. 3)

Less than 200 m of middle Kaza Group at its base was examined. It consists of fining and thinning upward beds of quartz- and feldspar-pebble grit, psammite, and grey-green phyllite.

Correlations with other sections of the Windermere Supergroup

With one main exception, Windermere units described in this report correlate well with subdivisions of the Supergroup in the northern Monashee Mountains (Simony et al., 1980) and the southern Cariboo Mountains (Pell, 1984; Pell and Simony, 1987). The exception is the absence of the prominent middle marble marker unit between the SPA and the Kaza Group. This prominent marker has been mapped south of and into the upper North Thompson River valley by Pell (1984) but does not appear on the eastern limb of the Adolph Creek Syncline. One possible explanation is that the marble changes facies to the northeast, possibly into unit 5e of the SPA. Supporting this possibility is the presence of amphibolite-bearing meta-clastic rocks above and below the calcareous units in this localities, similar to what is observed in the North Thompson River valley. If this correlation is correct, then unit 5f, the amphibolite-bearing meta-clastic rocks above the calcareous unit, correlates with the lower Kaza Group of Pell (1984) and Pell and Simony (1987), rather than SPA. Further work is required to establish more accurate links between the regions north and south of the upper North Thompson River.

Trench marble (unit 8, Fig. 3)

Underlying the SRMT northwestwards from Kinbasket Lake to near where Kiwa Creek enters the SRMT is massive to foliated, grey to brown, micaceous marble; dark grey, mylonitic, graphitic marble; and tremolite- and talc-bearing calc-silicate schist (Campbell, 1968). Campbell assigned this unit to the Cambrian, correlating it with thick massive carbonate of this age in the Rocky Mountains. This correlation implies that the contact of the marble with Windermere Supergroup to the west is a thrust fault.

METAMORPHISM

From the SRMT southwestward to just north of the upper North Thompson River valley, kyanite and staurolite coexist in rocks of the Windermere Supergroup. Large areas lack staurolite but this may reflect a lack of appropriate bulk composition rather than a difference in metamorphic conditions. Garnet zone rocks lie to the northwest along the trace of the Adolph Creek Syncline and to the southwest of the northwest-trending part of the North Thompson River.

Knots of quartz and plagioclase occur throughout much of the higher grade parts of the region, locally in sufficient concentration to lend the appearance of migmatite. The conditions of metamorphism in the southern part of this area have been discussed by Currie (1988) and Walker and Simony (1989).

STRUCTURAL GEOLOGY

Regional deformation

An exciting finding of the summer was basement-involvement in structures inferred to be part of Middle Jurassic, southwest-vergent (away from the craton) D_2 deformation. In the northeastern part of the region, from the Purcell Fault to Canoe River, a large northwest-trending, northeastwardly overturned anticline cored by early Proterozoic basement (Canoe River Anticline, CRA, Fig. 1, 2) and a syncline to the southwest (McLennan River Syncline, MRS) are folded over a later regional-scale antiform (Mt. Arthur Meighen Anticline, AMA). The structural geometry requires that the early anticline-syncline pair verge to the southwest, away from the craton. The Canoe River Anticline may be the continuation of the southwest vergent Raush Anticline of Murphy (1987) which lies down-plunge to the northwest in the hinge region of the Premier Anticlinorium of Campbell (1968, 1973).

East of the North Thompson-Albreda Fault, Malton gneiss is in a similar structural position to the body of gneiss coring the Canoe River Anticline and is therefore indirectly implicated in Middle Jurassic deformation as initially suggested by Morrison (1982). Because displacement on the North Thompson-Albreda Fault is small near the southern Rocky Mountain Trench, as evidenced by the presence of a basal Windermere unconformity on both sides of it (localities 1-4, Fig. 1) and only a few kilometres of offset of the trace of the Purcell Fault, it is likely that Malton gneiss projected beneath the rocks west of the North Thompson-Albreda Fault before its displacement and was connected with the body of basement gneiss in the Cariboo Mountains.

The prominent phase of deformation with overall vergence away from the craton is the second phase of regional-scale deformation (D_2 of Murphy, 1986, 1987). In southern McBride and northwestern Canoe River map areas, southwest-vergent structures fold an earlier foliation and northeast-vergent folds which, in this area, are rarely larger than outcrop-scale (Murphy, 1987). As the Canoe River Anticline and the McLennan River Syncline are D_2 folds, the Mt. Arthur Meighen Anticline must belong to at least a third phase of folding (D_3).

Mt. Arthur Meighen Anticline is the northeasternmost of a train of regional-scale, northwest-trending, northeast-vergent D_3 folds. From the axial surface trace of the McLennan River Syncline southwestward to the Premier Range drainage divide (Fig. 1, 2), regional-scale D_3 folds predominate. These include the Mt. Arthur Meighen Anticline, the Mt. Sir Lester Pearson Syncline (LPS) and the Premier Range Anticline (PRA). The Premier Anticlinorium as defined by Campbell (1968, 1973) is a composite of these three structures.

From the crest of the Premier Range Anticline westward to the upper North Thompson River valley, the structural geometry results from the interference of early cryptic regional-scale folds correlated with D_1 folds found in northwestern Canoe River and southern McBride map areas, and D_3 folds. D_1 folds (Zillmer Glacier Anticline and Kimmel Glacier Syncline) are northeast-vergent, pre-metamorphic, and their axial planar foliation is a schistosity defined by the parallel orientation of micaceous minerals. Later northeast-vergent folds (Premier Range Anticline, Adolph Creek Syncline, and North Thompson River Anticline) are late- to post-metamorphic with axial planar crenulations similar to those of the third phase structures described above.

Although shown as a simple coaxial interference pattern in Figure 2, lower Kaza Group rocks west of the Premier Range drainage divide are highly deformed (Fig. 4). On the northeast limb of the Adolph Creek Syncline, southwest-dipping, stratigraphically upright grit, psammite, and dark, locally graphitic phyllite of the lower Kaza Group folded by northeast-vergent D_1 folds pass down-stratigraphic-section to the northeast into a sequence of amphibolite, grit, quartzose psammite, calc-silicate rock, and rusty pelitic schist inferred to be SPA. This sequence in turn passes down-structural-section to the northeast into isoclinally folded, highly foliated psammite and locally graphitic pelitic schist inferred to be lower Kaza Group. Southwest-vergent D_1 folds associated with inverted graded beds facing to the northeast permit the inference of a D_1 anticline (Zillmer Glacier Anticline). These rocks pass farther down-structural-section to the northeast across a 5 km-wide zone of highly foliated psammite and schist folded by symmetrical isoclines into less foliated psammite and schist folded by tight, northeast-vergent D_3 folds. In this region, stratigraphic tops are to the southwest and the vergence of D_1 folds is to the northeast, permitting the inference of a D_1 syncline in the strongly foliated and symmetrically folded rocks to the southwest (Kimmel Glacier Syncline). This first phase northeast-vergent anticline-syncline pair is the only regionally mappable first phase structure yet defined in the southern Cordillera; however, more work is necessary to confirm this structural geometry.

Fabrics associated with the first three phases of regional deformation are ubiquitously deformed by transverse (northeast-trending) outcrop-scale and smaller kinks and crenulations of a fourth (D_4) phase of deformation. Occurring from the western Main Ranges of the Rocky Mountains to at least as far west as Bowron Lakes Provincial Park in the western Cariboo Mountains with a similar northeast trend, D_4 structures are somewhat enigmatic and difficult to interpret kinematically.

Mesoscopic fabrics and the sequence of deformation and metamorphism

First phase folds are characterized by an axial planar foliation defined by the parallel orientation of micas. This foliation is everywhere overgrown by prograde porphyroblasts and is therefore pre-metamorphic peak.

Second phase deformation is inferred to be syn-metamorphic but pre-metamorphic peak. The axial planar fabric of second phase folds is a foliation defined in pelitic rocks by tight to isoclinal crenulations of fine to coarse grained micas aligned in the D_1 foliation, by coarser micas aligned parallel to the axial surfaces of D_2 crenulations, and in coarser clastic rocks, by the plane of flattening of flattened and elongate clasts. The long axis of clasts elongate within the D_2 foliation is uniformly to the northwest, defining an important regional lineation. The D_2 fabric is gener-

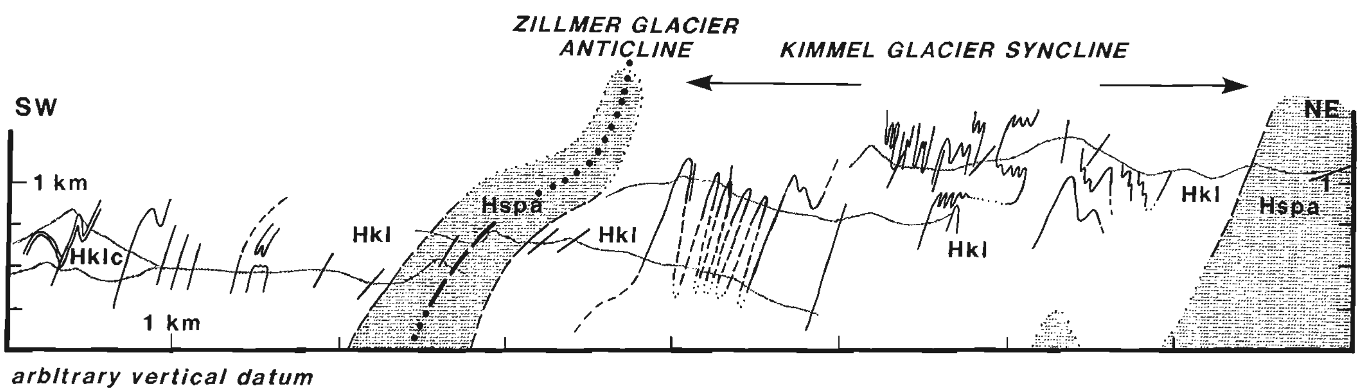

Figure 4. Cross-section from headwaters of Camp Creek to the upper North Thompson River valley illustrating complex structure in region where Zillmer Glacier Anticline and Kimmel Glacier Syncline are inferred. Exact location of axial surface trace of KGS is not known but in inferred to occur in zone of symmetrical isoclinal folding beneath the word 'Glacier' in Kimmel Glacier Syncline label. Section compiled from 4 cross-sections (topographic profiles shown) based on 1:50 000 scale mapping.

ally overgrown by randomly oriented prograde minerals, indicating that D_2 deformation is pre-metamorphic peak. The amount of syn-metamorphic deformation, however, is not clear. Prograde minerals aligned parallel to the axial planes of D_2 crenulations and to the D_2 lineation have been used in the past as evidence of synmetamorphic deformation (Murphy, 1986); it is also possible to interpret this observation as a result of mimetic recrystallization during a purely post-kinematic metamorphism.

Third phase folds are characterized by axial planar crenulations. D_3 crenulations generally accommodate porphyroblasts, as though porphyroblasts predate D_3, but porphyroblasts locally partly truncate crenulations, implying that some of the strain occurred before complete growth of the porphyroblast. Third phase deformation, therefore, is late- to post-metamorphic.

It is very difficult, in isolated outcrops, to distinguish third phase crenulations from crenulations associated with the second phase of deformation which occurred during metamorphism, but before its peak. Second phase crenulations are generally completely overgrown by porphyroblasts but locally, the second phase foliation is flattened around porphyroblasts, presumably due to third phase strain, creating a fabric-porphyroblast relationship similar to that of third phase structures. Only in areas where the two phases of northwest-trending crenulations occur together is the distinction between second and third phase fabrics unambiguous.

Major faults

The stratigraphy and structures described above lie in a structural block bounded on the east and northeast by major faults. To the east, the block is bounded by the North Thompson-Albreda Fault (NTAF), a down-to-the-west normal fault (Campbell, 1968; Pell, 1984). The trace of the NTAF is marked by a zone of intense fracturing and clay gouge. Shear fractures affirm earlier interpretations of downdip normal displacement. The trace of the fault has been extended to the north into the SRMT where it offsets by 3-5 km the trace of the Purcell Fault which bounds the structural block on the northeast. Ambiguity in the amount of strike-separation reflects inaccuracy in locating the offset trace of the Purcell Fault east of the NTAF.

To the northeast, the structural block is bounded by a fault juxtaposing rocks of the lower Windermere Supergroup just described against a foliated marble of unknown but presumed Cambrian age. The fault is post-metamorphic because progressively lower grade rocks are juxtaposed against the marble as one proceeds to the northwest along it. Thrust displacement is inferred from stratigraphic juxtaposition; the fault is not exposed, so kinematic data are unavailable. Following Campbell (1968, 1973) and Mountjoy (1980), this fault has been correlated with the Purcell Fault farther southeast because it is post-metamorphic, relatively steep, likely to be a thrust fault, and it juxtaposes multiply deformed rocks of the lower Windermere Supergroup against less deformed, presumably Cambrian rocks.

The Purcell Fault is truncated by the system of faults which make up the northeastern boundary fault of the SRMT. As is discussed by Murphy (1990), this fault has an early history of dextral strike-slip and a later history of down-to-the-west normal faulting. The combined effect of these displacements is to offset the Purcell Fault approximately 55 km to the southeast.

Neither the Camp Creek nor Allan Creek faults inferred by Currie (1988) and Walker and Simony (1989) are required by the present interpretation.

AGES OF DEFORMATIONS, METAMORPHISM(S?), AND FAULT DISPLACEMENTS

The age of the first phase of folding is unknown; second phase (southwest-vergent) deformation is most likely Middle Jurassic (Pigage, 1977; Gerasimoff, 1988; Murphy, 1989). Third phase deformation (D_2 of Currie, 1988) is constrained to be between 154 $\pm$ 6 Ma and 125 $\pm$ 7 Ma, the age of pre- and post-D_3 pegmatites (Currie, 1988). If porphyroblasts indeed grew early in D_3, then D_3 deformation can be even more closely constrained to lie between 135 $\pm$ 4 Ma, the age of metamorphic monazite and 125 $\pm$ 7 Ma (Currie, 1988). The age of the final phase is unknown.

The main metamorphism in this area is 135 $\pm$ 4 Ma (Currie, 1988). Pigage (1977) has presented evidence for a Middle Jurassic period of metamorphism in similar rocks at the northern end of Wells Gray Park. Middle Jurassic mineral growth in the area of this report may be indicated by the apparent synkinematic growth of minerals parallel to the D_2 foliation and lineation. However, it is also possible that these minerals may have grown mimetically during the Cretaceous metamorphism.

On the basis of timing of rapid uplift of the Monashee Mountains east of the NTAF, motion on the NTAF has been constrained to be between 45 and 51 Ma (Sevigny et al., in press). Motion on the Purcell Fault follows Cretaceous metamorphism and, as it is cut by the NTAF, is pre-Eocene.

Down-to-the-west normal faulting in the SRMT is post-metamorphic and may be the same age as motion on the NTAF although there are no pertinent geochronological data. Strike-slip faulting in the SRMT postdates motion on the Purcell Fault and probably predates Eocene down-to-the-west faulting.

SUMMARY OF MAIN FINDINGS

1. The southern Cariboo Mountains are underlain by an unfaulted sequence of the lower Windermere Supergroup from a basal unconformity on basement orthogneiss and mantling metasediments up to and including the middle Kaza Group.

2. Early Proterozoic basement orthogneiss and younger mantling metasedimentary rocks core a major southwest-vergent fold nappe, demonstrating the existence and nature of basement involvement in probable Middle Jurassic deformation.
3. The North Thompson-Albreda normal fault has been extended into the SRMT where it offsets the trace of the Purcell Fault by up to 5 km. This implies a linkage between late down-to-the-west faulting in the SRMT and down-to-the-west faulting on the North Thompson-Albreda Fault.
4. The Purcell Fault is truncated and offset 55 km by a combination of dextral strike-slip and normal dip-slip along the fault forming the northeastern structural boundary of the SRMT.

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Geology of the Ptarmigan Creek map area (east half) and adjacent regions, Main Ranges, Rocky Mountains, British Columbia

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Abstract

Middle Miette Group strata overlie a lower panel of Upper Proterozoic Miette Group strata and several tens of metres of possible Lower Cambrian Gog Group. The complex boundary between these two panels is suggested to be an important decollement, thought to be one of three important folded low-angle thrust faults exposed in the Selwyn Range. The Rockingham Creek and Ptarmigan Creek thrust zones imbricate strata in the hanging wall and footwall of the Selwyn Range decollement, respectively. Recumbent northeast-verging folds and minor thrust faults are associated with these faults. Later upright folding produced the Fraser River antiform. The Moose Lake/Chatter Creek fault is a late fault on the eastern margin of the Selwyn Range and is thought to cut the early, folded faults. The large amount of shortening has implications for the geometry of subsurface structures elsewhere in the Main Ranges of the Canadian Rockies.

Résumé

Les couches du groupe de Middle Miette recouvrent un panneau inférieur des couches du groupe de Miette du Protérozoïque supérieur et plusieurs dizaines de mètres du groupe de gog du Cambrien inférieur possible. On suppose que la limite complexe qui sépare ces deux panneaux est un décollement important qu'on pense être une des trois failles inverses à faible angle de plissement, qui affleurent à la chaîne Selwyn. Les zones de chevauchement de Rockingham Creek et de Ptarmigan Creek donnent respectivement des couches imbriquées au toit et au mur du décollement de la chaîne Selwyn. Des plis couchés de vergence nord-est et des failles inverses mineures sont associés à ces failles. Un plissement droit postérieur a produit l'antiforme Fraser River. La faille de Moose Lake/Chatter Creek est une faille tardive située sur la marge orientale de la chaîne Selwyn et, d'après l'auteur, elle traverse les failles plissées antérieures. Le raccourcissement important a des répercussions sur la géométrie des structures souterraines ailleurs dans les chaînes Main des Rocheuses canadiennes.

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INTRODUCTION

The Ptarmigan Creek area (Fig. 1, 2) is underlain by upper Proterozoic metasediments of the Miette Group. These rocks have been deformed and subjected to greenschist or low amphibolite facies metamorphism.

The region was mapped on a reconnaissance basis by Price and Mountjoy (1970) as part of Project Bow-Athabasca. Recent, more detailed studies include Leonard (1984, 1985), Forest (1985), Mountjoy et al. (1985), Mountjoy and Forest (1986) and McDonough and Simony (1986, 1988). These studies showed that a major structural culmination was centred on the Ptarmigan Creek region. A large structure in this culmination, the Fraser River antiform, brings up the deepest known structural levels in the Main Ranges. A stack of thrust faults, which includes the Ptarmigan thrust of Mountjoy and Forest (1986) is folded by the Fraser River antiform (Fig. 3, section A-A'). The core of the antiform is exposed in a window beneath these faults.

These papers recognized a four-fold division within the Miette Group. An upper granule conglomerate-dominated unit overlain by a pelite-dominated unit were correlated to the middle and upper Miette groups, respectively, based on lateral continuity with these units in other regions. The lower pair of granule conglomerate-dominated unit and pelite-dominated unit were assigned to the lower Miette Group based on their perceived stratigraphic position.

The present paper describes an important fault zone, the Selwyn Range decollement, that divides the stratigraphy described in previous work into two panels (Fig. 1) and is interpreted to duplicate Miette Group stratigraphy in the Selwyn Range. The previously interpreted lower Miette Group strata are hereby reassigned to middle and upper Miette Group.

STRUCTURE

The present work clarifies the structural style of the previously described "Ptarmigan thrust" (Mountjoy et al., 1985; Mountjoy and Forest 1986) and also presents the first information about the important Selwyn Range decollement, which is responsible for large scale structural and stratigraphic duplication in the Main Ranges.

Deformation in the Selwyn Range can be thought of as occurring in several stages. One set of structures includes the Rockingham Creek thrust, Selwyn Range decollement, Ptarmigan Creek thrust zone and associated tight northeast verging folds. Because the earliest fabric associated with these structures is locally a crenulation, these major faults must be at least the second set of structures in the Selwyn Range. These second phase folds and faults are overprinted by minor third stage southwest-verging folds and associated small-scale faults. The fourth set of structures includes the Fraser River antiform and the Moose Lake/Chatter Creek fault. Still later features are minor kink bands and thrust faults with oblique right-lateral offset.

These five sets of structures are treated as distinct phases as they appear to be unrelated on the scale of the Selwyn Range. Only the two main sets of structures, the second and fourth, are described in this paper.

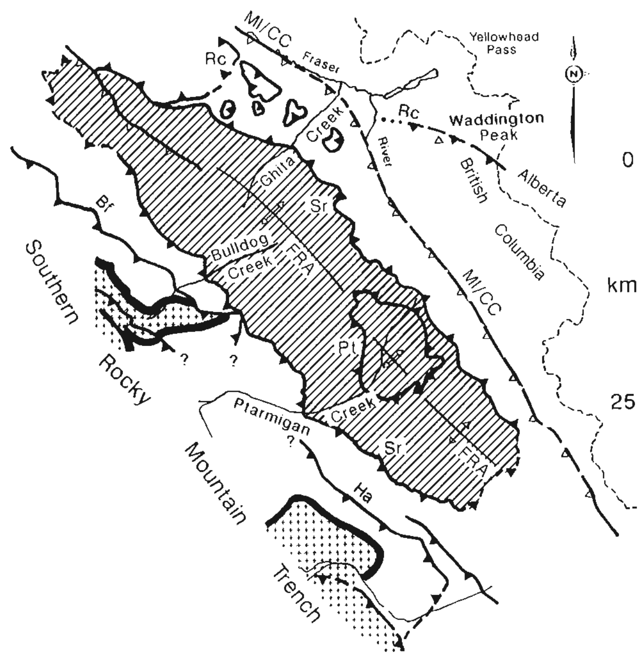

Figure 1. Tectonic elements of the Selwyn Range, in part after McDonough and Simony (1986, 1988). Lower structural panel shown by hachured pattern, upper panel is unpatterned, basement gneisses are demarked by a cross pattern. Early thrust faults have solid teeth, late ones have hollow teeth. The Moose Lake/Chatter Creek fault (Ml/CC) represents the boundary between the eastern and western Main Ranges. FRA - Fraser River antiform, Pt - Ptarmigan Creek thrust zone, Sr - Selwyn Range decollement, Rc - Rockingham Creek thrust, Bf -Bearfoot thrust, Ha - Hugh Allan thrust.

Second Phase Structures

Selwyn Range Decollement

The Selwyn Range décollement is a 50-100 m thick, typically pelitic, shear zone that can be traced around most of the Selwyn Range, where exposure and access permit. F_2 folding is most common near the shear zone and folds decrease in both size and abundance upwards and downwards from the shear zone (Fig. 3, section B-B'). It is thus suggested that the shear zone was produced during D_2.

Pelites outside of the Selwyn Range décollement generally contain S_2 and S_4 but still preserve submillimetre-scale laminations. Within the fault zone, evidence for bedding is typically absent. Pelitic material is dominated by a fabric element not present outside of fault zones. This fabric, S_2', is a northeast-directed crenulation cleavage interpreted as shear bands (Fig. 4, cf. White et al., 1980). S_2' typically cuts up to the northeast more steeply than does S_2. Initially, this geometry leads to shortening and rotation of S_2; this rotation causes the development of small sigmoidal lenses of quartz between S_2 folia (Fig. 4b). The geometry is similar to small scale **en echelon** extension veins typical of

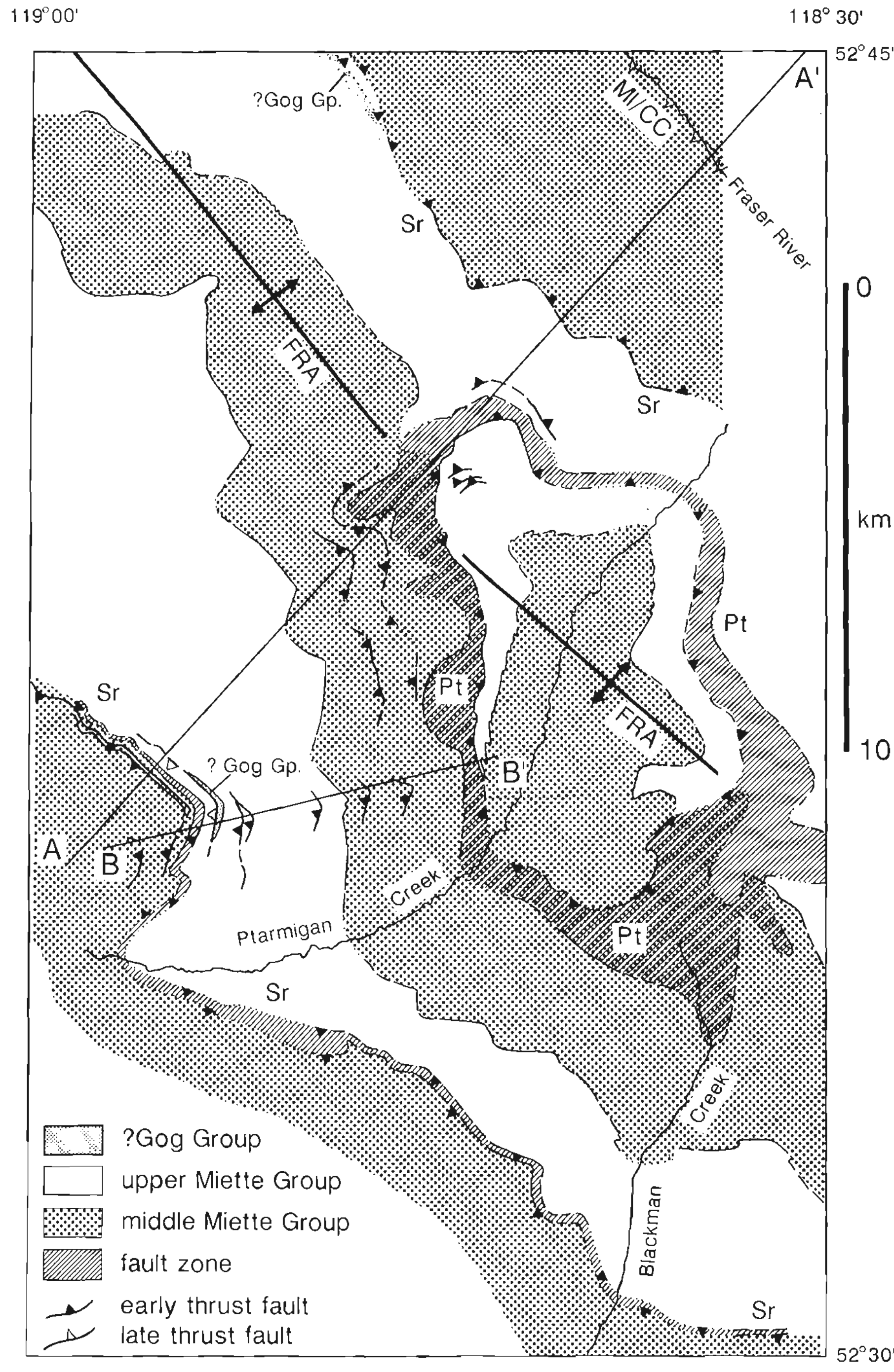

Figure 2. Geological map of the Ptarmigan Creek E 1/2 map area. Abbreviations as in Figure 1.

brittle-ductile shear zones (*see* summary in Ramsay, 1980). However, at the scale at which this feature is developed in the Selwyn Range, pressure solution, diffusion and reprecipitation of silica may have occurred without brittle failure. With enough rotation, extension of S_2 occurs (cf. Carreras et al., 1980, their Fig. 3c) and a new S surface nucleates (Fig. 4c). Offset of bedding on individual S_2' shear bands outside of the main fault zone is typically on the order of 5-30 cm.

Locally, foliation fish (Hanmer, 1984), indicating top-to-the-northeast motion, are also present within the fault zone. Conglomeratic sediments, where present in the main and subsidiary fault zones, display well developed northeast-southwest stretching lineations.

What is the structural evidence that the Selwyn Range décollement is a thrust fault? Outside of the major shear zones, it is readily observable that S_2' (which has top-to-the-northeast motion on it) cuts up through bedding towards the northeast. Within major shear zones such as the Selwyn Range décollement, bedding is absent, although S_2 is present. Regionally, S_2 cuts up to the northeast relative to bedding. Since shear bands (S_2') are observed to cut up to the northeast across S_2, they are also constrained to cut up to the northeast relative to bedding. It is thus suggested that the Selwyn Range décollement, like its internal components, also cuts up to the northeast in this region.

Ilmenite and epidote crystals lie within S_2. Garnet and biotite porphyroblasts overgrow these early minerals and early stages of S_2' development. Final tightening of S_2' postdates garnet growth but micas outlining S_2' subsequently recrystallized.

The Selwyn Range décollement forms the boundary between two panels of broadly similar mechanical behaviour but contrasting lithologies. The sandstone/conglomerate and silty chloritic pelitic sequences of the middle Miette Group in the upper structural panel are distinct from sandstone-quartzite-conglomerate and aluminous pelite sequences of the lower structural panel. Along the eastern margin of the Selwyn Range, the Selwyn Range décollement separates two dominantly pelitic sequences.

The Selwyn Range décollement presently crops out as an oval, 60 by 15 km in dimensions (Fig. 1), because it and its fabrics have been folded by the doubly plunging Fraser River antiform. Because of this later folding, the fault zone shows reverse motion on the west-dipping southwest limb of the Fraser River antiform and apparent normal motion on the northeast limb of the Fraser River antiform (Fig. 3, section A-A').

Ptarmigan Creek Thrust Zone

Mountjoy et al. (1985) and Mountjoy and Forest (1986) defined the "Ptarmigan thrust" as the lowest fault in a stack of faults that they mapped near Ptarmigan Creek. The present mapping indicates such a definition is not feasible as there is no clearly defined lower boundary to the faulting. Faults outside of the fault zone have small offset, while those within the zone are an order of magnitude larger. The Ptarmigan Creek thrust zone generally occurs structurally higher than the faults previously mapped as the Ptarmigan thrust (Fig. 2).

The Ptarmigan Creek thrust zone is here defined as the belt in which phase two folds show an "S" geometry when viewed looking towards the northwest along their fold axes (Fig. 3, section B-B'). Both the footwall and the hanging wall of the fault zone contain "Z" folds. Overturned stratigraphy is thus most common within the fault zone, whereas upright stratigraphy dominates both the footwall and hanging wall.

Associated with the Ptarmigan Creek thrust zone and smaller northeast-verging fault zones are two sets of tight folds (F_2 and F_2'), two sets of northeast-verging crenulations (S_2 and S_2') and local top-to-the-northeast extensional crenulations (cf. Platt and Vissers, 1980). F_2 and S_2 are related to a large recumbent fold pair that is obvious in Figure 3, section A-A'. These features are most intensely developed near and in the Ptarmigan Creek thrust zone, and change asymmetry across the fault zone boundaries. Large F_2 developed in conglomerate within the Ptarmigan Creek thrust zone have highly attenuated overturned limbs. Later

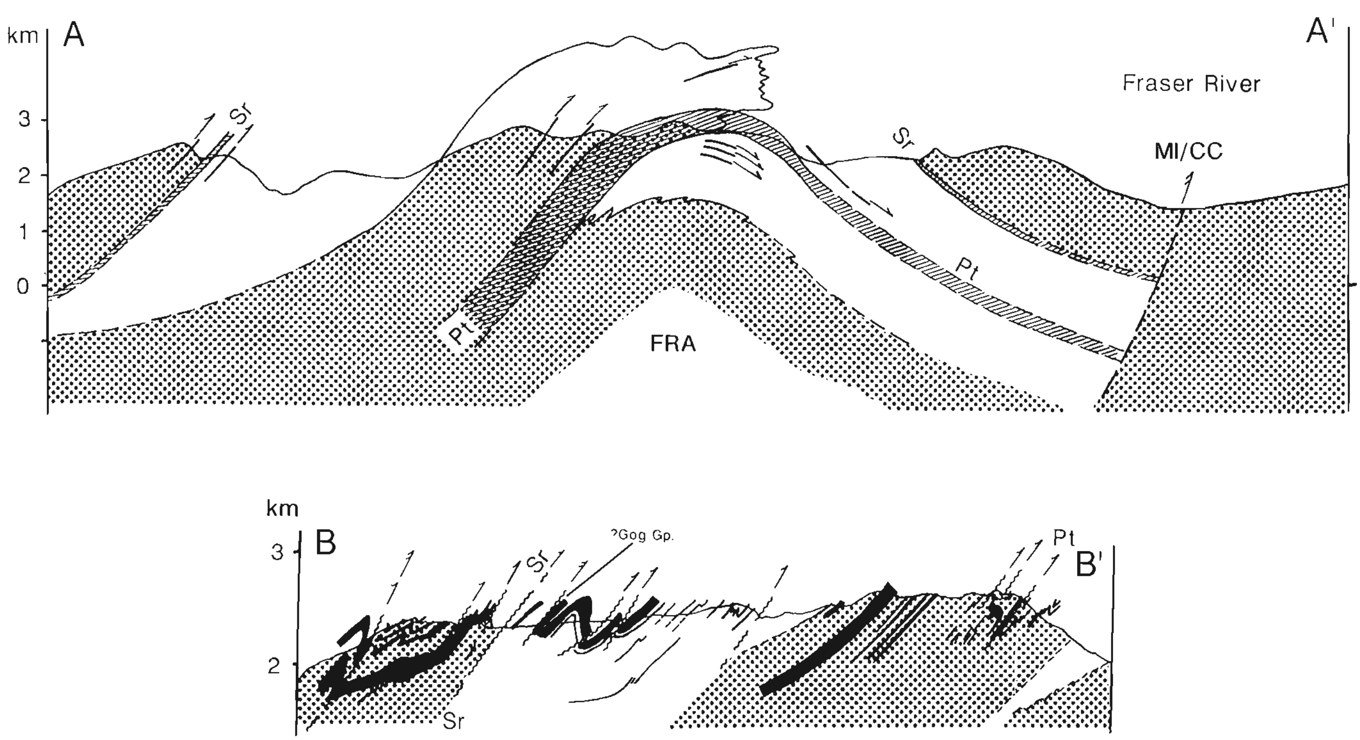

Figure 3. Structure sections of the Ptarmigan Creek area. Lines of section shown in Figure 2. Section B-B' contains quartzite and granule conglomerate sequences in black. Single arrowheads indicate early thrust faults and double arrowheads represent late faults. Note different scales on the two cross-sections.

structural elements, S_2' (interpreted as shear bands), F_2' and the extensional crenulations, indicate that top-to-the-northeast shear was superimposed on the overturned limb of the major F_2 structure. These late elements are restricted to fault zones and are most common within the Ptarmigan Creek thrust zone itself. Because bedding is typically present (although transposed) within the Ptarmigan Creek thrust zone, northeast-directed folds (F_2') superimposed on previously overturned strata are common.

Fault zone rocks preserve a history of increasing metamorphic grade during deformation. Ilmenite grains that lie in S_2 are broken and rotated by S_2'; epidote porphyroblasts overgrow early stages of S_2' but are subsequently rotated during later stages of shearing. Garnet porphyroblasts overgrow S_2' but predate S_4; their relationship with S_3 is unknown.

Like the Selwyn Range décollement, the Ptarmigan Creek thrust zone is arched by the Fraser River antiform and thus crops out as a 10 by 14 km oval (Fig. 1, 2). Shortening, calculated by bed length, of the Ptarmigan Creek thrust zone and related structures is about 18 km. The large bulk strain has been ignored in this calculation.

Rockingham Creek Thrust

The Rockingham Creek thrust (Fig. 1) cores a large anticline east of the Moose Lake/Chatter Creek fault (Dechesne and Mountjoy, 1990). West of the Moose Lake/Chatter Creek fault, a stack of thrust faults in the hanging wall of the Selwyn Range décollement in the northeastern portion of the Selwyn Range are interpreted to be the western equivalents of the Rockingham Creek thrust. Southern equivalents of these faults have not been identified due to poorer exposures in the southern Selwyn Range. However, it is possible that the Cube Ridge thrust (Mountjoy et al., 1985), suggested by Mountjoy and Forest (1986) to be an out- of-sequence thrust, may instead be part of this set of structures (Dechesne and Mountjoy, in press; Mountjoy and Grasby, 1990).

The Rockingham Creek thrust splays off from the Selwyn Range décollement along the northern margin of the window and is responsible for considerable structural thickening in low portions of the middle Miette Group in the hanging wall of the Selwyn Range décollement on the east side of the Fraser River antiform.

Fourth Phase Structures

Fourth phase folds, faults and crenulations were mapped throughout most of the study area.

The Fraser River antiform is the largest structure produced during this event. All earlier structures were arched by this structure. The Fraser River antiform is traceable from shallow structural levels of the northern and southern Selwyn Range to the Ptarmigan Creek valley in the central part of the Selwyn Range, where the deepest levels of the region are exposed (Fig. 1, 2).

Fourth phase faults commonly exploit southwest-dipping and northeast-dipping portions of second phase faults. A minor fourth phase thrust fault (the Packsaddle fault of McDonough and Simony, 1986) may reuse a portion of the Selwyn Range décollement along the western side of the study area. These fourth phase faults transfer their motion to open folds with near vertical axial surfaces at higher structural levels (McDonough and Simony, 1986).

The Chatter Creek fault, a steep fault lying in the upper Fraser River valley (Fig. 1), is correlated with the Moose Lake fault to the north (Dechesne and Mountjoy, 1990). This fault was produced during D_4 (samples collected by Bryan Hannis of Valemount, British Columbia, show post-metamorphic fabrics).

The Ptarmigan thrust was initially interpreted to merge eastwards with the Chatter Creek fault (Mountjoy and Forest, 1986) but the present mapping shows that there is insufficient displacement on the Moose Lake/Chatter Creek fault to accommodate the shortening of the Ptarmigan Creek

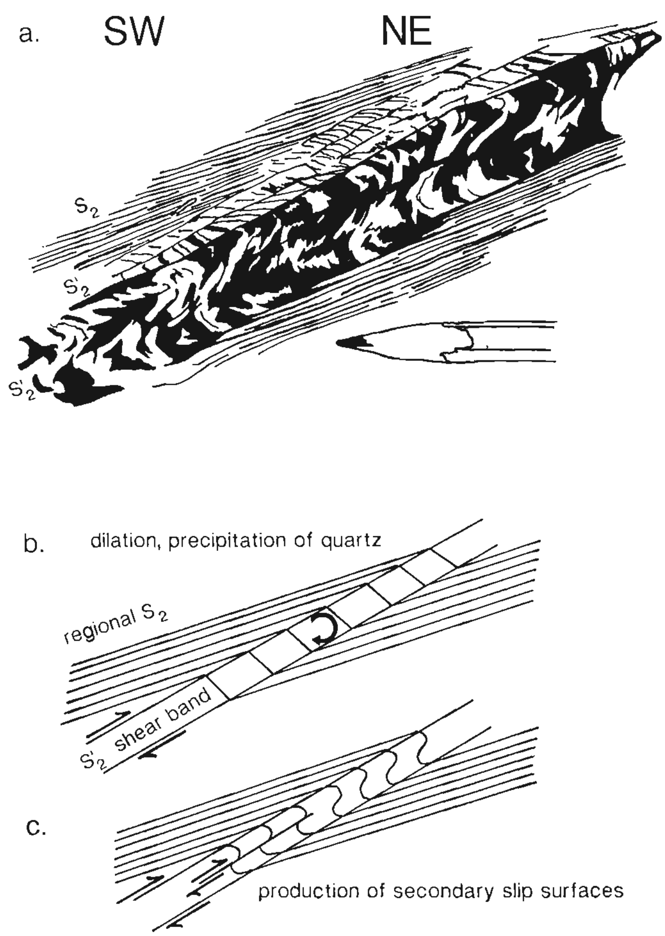

Figure 4. a). Tracing of photograph of S_2' overprinting S_2 within Selwyn Range decollement on west side of window (locality is on section B-B' of Fig. 3). Black denotes phyllosilicates, white, quartz. Note incipient slip surface within the large S_2' shear band. b). Early stages of S_2' development, such as the smaller examples in Figure 4a, rotate and cause dilation between S_2 folia. c). With further rotation, new slip surfaces within the shear band accommodate extension of S_2 folia.

thrust zone, let alone the Selwyn Range décollement and Rockingham Creek thrust. Based on the trajectories of the early faults determined by mapping out the orientation of S_2, the Moose Lake/Chatter Creek fault must cut these early, low-angle faults (Dechesne and Mountjoy, 1988). The Moose Lake/Chatter Creek fault also offsets part of the western limb of the Mount Robson syncline and the Rockingham Creek thrust (Dechesne and Mountjoy, 1990).

Implications for Subsurface Geometry East of the Moose Lake/Chatter Creek Fault

Eastern equivalents of the Selwyn Range décollement and the Ptarmigan Creek thrust zone have not been identified, but probably include a combination of early thrust faults from the eastern Main Ranges (eg. Snaring, Monarch, Simpson Pass, Moose Pass thrusts; Dechesne, 1989; Dechesne and Mountjoy, 1988, 1990, in press). Further stratigraphic work in the Miette Group is necessary to more fully constrain the possibilities.

STRATIGRAPHY

Stratigraphy of the Ptarmigan Creek region can be divided into two panels (Fig. 1), with the Selwyn Range décollement taken as the boundary. Middle and upper Miette Group are present in the upper panel. Strata of the lower panel were previously assigned to the lower Miette Group (eg. Mountjoy and Forest, 1986; McDonough and Simony, 1988), but new structural evidence (Selwyn Range décollement) and stratigraphic information suggest that the lower panel is more appropriately correlated with middle and upper Miette Group.

Based on a variety of sedimentary structures, all of the strata exposed within the study area are interpreted as subaqueous mass flow and turbidite deposits.

Upper Panel

Upper Miette Group

Upper Miette Group of the upper structural panel do not out crop within the Ptarmigan E 1/2 map area, but are found in adjacent regions (McDonough and Simony, 1988; Dechesne and Mountjoy, 1990; Mountjoy and Grasby, 1990).

Middle Miette Group

The middle Miette Group is composed of alternating thick cliff-forming sandstone/granule conglomerate sequences and more recessive units of chloritic silty and sandy pelitic material. Its thickness in the region is approximately 2400 m.

Five stratigraphic subdivisions are recognized within the middle Miette Group in the Ptarmigan Creek region (Fig. 5). The lower four are present along the western side of the Selwyn Range; the upper four out crop in the Lucerne E 1/2 map area (Dechesne and Mountjoy, 1990).

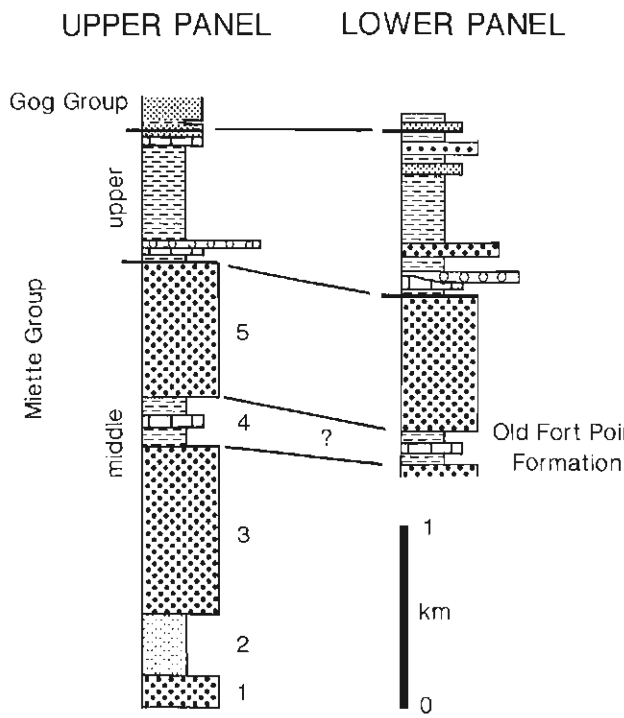

Figure 5. Generalized stratigraphic columns for the Miette Group within the study area. Upper panel from mapping in the Ptarmigan Creek E 1/2 and Lucerne E 1/2 map areas; upper Miette and Gog Group from the Miette Pass region of the Lucerne E 1/2 map area (Dechesne and Mountjoy, 1990). Lower panel from Ptarmigan Creek E 1/2 map area.

Along the western side of the Selwyn Range, between Ptarmigan and Bulldog creeks, in the immediate hanging wall of the Selwyn Range décollement, the lowest subdivision consists of sandstone and conglomerate sequences and silty chloritic pelites. These strata total about 150 m, but are not present along the eastern side of the range.

A second subdivision, composed of greenish grey silty to sandy pelite at least 350 m thick overlies the lowest subdivision. This unit lies upon the Selwyn Range décollement along the eastern side of the Selwyn Range. This subdivision contains local lensoid bodies of sandstone and conglomerate.

The third subdivision is composed of sandstone and conglomerate sequences with recessive sequences of silty chloritic pelite. About 900 m of this subdivision are recognized. In the upper structural panel along the east side of the Selwyn Range and east of the Fraser River, the uppermost package contains considerable amounts of pebble and cobble conglomerate.

The Old Fort Point Formation represents the fourth subdivision of the middle Miette Group. At Waddington Peak, the Old Fort Point Formation is 305 m thick. Forest (1985), McDonough and Simony (1986) and Mountjoy and Grasby (1990) also recognized this marker which consists, from bottom to top, of green pelite and semipelite, orange weathering carbonate, and black pelite; local matrix supported conglomerate (pelite and carbonate clasts), interpreted as debris flows, were noted.

The uppermost subdivision is approximately 700 m thick. Like the lowest and the third subdivision, the uppermost subdivision of the middle Miette Group is dominated by sandstone/conglomerate and sandy and silty pelite sequences.

Lower Panel

Two main stratigraphic units were mapped (Fig. 5), a granule conglomerate - pelite sequence (a little over 800 m thick) and an overlying pelite - quartzite sequence (1100 m thick), that are here interpreted as middle and upper Miette respectively.

Fault bounded slices of pink and purplish quartzite at the top of the lower structural panel (Fig. 2, 3), may represent uppermost Miette Group or possibly, basal Gog Group.

Middle Miette Group

In the lower structural panel, three subdivisions are present. The lowest subdivision is about 100 m thick and consists of sandstone, granule conglomerate and silty pelite. The base of this unit is carried on the Ptarmigan Creek thrust zone.

The overlying subdivision is about 70 m thick and consists of local greenish-grey pelite and siltstone, overlain by black, reddish-orange and orange-brown weathering carbonate, and then by dark, graphitic carbonate and pelite. Matrix-supported conglomerate 0.7 m thick locally caps this subdivision. The lithologies and tripartite nature of this subdivision are similar to those of the Old Fort Point Formation of the upper panel, to which it is correlated.

The uppermost subdivision is approximately 700 m thick and comprises alternating units of granule conglomerate and silty greenish grey pelite.

Upper Miette Group

The upper Miette Group in the lower panel is 1100 m thick and is dominated by pelites and quartzites. At its base is a marker sequence that consists of dark, graphitic pelites and limestone, orange weathering psammites and quartzites which grade laterally to orange weathering matrix-supported boulder conglomerates.

Above the marker sequence is a thick succession of brown weathering pelites and minor grey weathering quartzitic psammites. A boulder conglomerate (G_1 of Mountjoy and Forest, 1986) forms a mappable horizon low within this succession. Because of deformation, it is not clear that the unit mapped as G_2 by Mountjoy and Forest (1986) actually represents a single horizon, and therefore it is not possible to define a correlative unit within the present study area.

Quartzite and aluminous pelite of the upper Miette Group pass upwards into carbonate-cemented sandstone, well-sorted granule conglomerate and subordinate aluminous pelite.

?Gog Group

Locally, a 45-50 m thick pink or purplish-brown quartzite, and an overlying pelite cap the upper Miette Group (Dechesne and Mountjoy, 1990). These strata are inferred to represent either the upward transition from Miette Group to Gog Group or actual Gog Group.

DISCUSSION OF STRATIGRAPHIC CORRELATIONS

Lower Miette Group strata were interpreted to be widespread in the Selwyn Range by Mountjoy and Forest (1986) and this interpretation was also used by McDonough and Simony (1986, 1988). The present work suggests that an alternative interpretation is possible. Agter additional fieldwork in the southern Selwyn Range, Mountjoy and Grasby (1990) concluded that most of the strata previously assigned to lower Miette Group in that region are also more properly correlated with middle and upper Miette Group.

Three main lines of evidence suggest that the strata of the lower panel are structural repeats of strata of the upper panel.

Firstly, a thick shear zone (Selwyn Range décollement) separates the two panels. This structure has similar microstructures to those of the Ptarmigan Creek thrust zone, which is readily shown to be a thrust fault (Mountjoy et al., 1985; Mountjoy and Forest, 1986; this work) because of the easily recognized stratigraphic offset.

Secondly, graphitic pelite, carbonate and granule conglomerate that immediately underlie middle Miette Group at the type lower Miette Group region of the Cushing Creek 140 km to the north (Carey and Simony, 1985; Campbell et al., 1973) are not present immediately below the middle Miette Group in the Selwyn Range, suggesting that the strata in the Selwyn Range are not lower Miette Group.

Thirdly, stratigraphic similarity of the lower panel to the upper panel allows correlation of not only broad stratigraphic divisions, but also several distinctive units.

a) The lower panel contains a possible equivalent to the Old Fort Point Formation of the middle Miette Group.
b) The basal portion of the upper Miette Group in the upper panel commonly contains graphitic pelite and black limestone overlain by matrix supported conglomerates containing carbonate clasts (Dechesne and Mountjoy, 1990; Charlesworth et al., 1967). Similar lithologies are present at the base of the interpreted upper Miette Group in the lower panel (this study).
c) Pink and purplish brown quartzites occur in the immediate footwall of the Selwyn Range décollement. These are tentatively correlated with purple and pink quartzites of uppermost Miette Group as at Mount Fitzwilliam (Teitz, 1985) or Mount Edith Cavell. Similar lithologies are present in lowest Gog Group in a number of other regions.

The greenish grey silty and sandy pelites of the second subdivision of the middle Miette Group of the upper panel (Fig. 5) were initially assigned to lower Miette Group by McDonough and Simony (1986), but they later revised their correlation to place most of this subdivision within the base of the middle Miette Group (McDonough and Simony, 1988). Only the lower, dark greyish green portion of this subdivision is placed within the lower Miette Group by McDonough and Simony (1988). Mountjoy and Grasby (1990) referred to this subdivision as lower Miette Group. However, the dissimilarity of these strata to the type lower Miette Group (see above), and the presence of underlying sediments characteristic of the middle Miette Group in other parts of the Selwyn Range suggest that these strata are part of the middle Miette Group.

CONCLUSIONS

Two main deformational events control the geometry of the strata in the Ptarmigan Creek area. The earlier main event (D_2) produced recumbent, northeast-verging, tight folds and associated thrust faults. The second major deformation (D_4) was responsible for upright folds and steep faults. Several important fault zones, the Ptarmigan Creek thrust zone, Selwyn Range décollement and Rockingham Creek thrust were active during the first main deformation. The footwalls of the Ptarmigan Creek thrust zone and the Selwyn Range décollement are exposed in large windows in the core of the Fraser River antiform. The Fraser River antiform and the Moose Lake/Chatter Creek fault are the major structures produced during the second major deformation.

Strata previously assigned to lower Miette Group are suggested to be structural repeats of middle and upper Miette Group across the Selwyn Range décollement.

The Selwyn Range contains two windows of regional scale, implying that the amount of shortening across the Selwyn Range is much higher than previously anticipated. The geometries mapped within the Selwyn Range have implications for deep structures elsewhere in the Main Ranges of the Canadian Rockies. Specifically, higher shortening, multiple low-angle faults and a detachment between the Miette and Gog groups are expected.

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Direct evidence for dextral strike-slip displacement from mylonites in the southern Rocky Mountain Trench near Valemount, British Columbia

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Murphy, D.C., Direct evidence for dextral strike-slip displacement from mylonites in the southern Rocky Mountain Trench near Valemount, British Columbia; in Current Research, Part E, Geological Survey of Canada, Paper 90-1E, p. 91-95, 1990.

Abstract

Horizontally lineated greenschist facies mylonitic orthogneiss was discovered along the northeastern shore of Kinbasket Lake in the southern Rocky Mountain Trench southeast of Valemount. Subsequent thin - section examination revealed structures indicative of dextral displacement. Although this discovery is sure to fuel the recent controversy over the nature and magnitude of faulting in the southern Rocky Mountain Trench, the net strike-separation due to strike-slip and previously recognized dip-slip is inferred to be at most 55 km.

Résumé

Un orthogneiss mylonitiques à faciès schiste vert avec des linéaments horizontaux a été découvert le long de la rive nord-est du lac Kinbasket, dans la partie sud du sillon des Rocheuses, au sud-est de Valemount. L'étude de lames minces a révélé la présence de structures indicatives d'un déplacement dextre. Même si cette découverte alimentera sûrement la controverse récente quant à la nature et l'amplitude des failles dans la partie sud du sillon des Rocheuses, le rejet longitudinal net, dû au rejet horizontal et au déplacement normal reconnus antérieurement, est supposé être de 55 km au maximum.

INTRODUCTION

In the course of regional mapping farther to the compilation of the Canoe River (1:250 000) map area (Murphy, 1990), I examined an outcrop of basement gneiss (Yellowjacket gneiss of McDonough and Simony, 1988a) on the northeastern shore of Kinbasket Lake between Dave Henry and Yellowjacket creeks, along the northeastern physiographic boundary of the southern Rocky Mountain Trench (SRMT, Fig. 1). Down-to-the-west shear fractures had been reported from this stretch of outcrop (McDonough and Simony, 1988a) and used in support of such displacement on the fault along the northeastern side of the SRMT. Although down-to-the-west shear fractures are indeed common along this stretch of outcrop, what is not obvious is that the rock upon which they occur is hard, horizontally lineated, siliceous mylonitic orthogneiss. This surprising and exciting discovery opens anew the controversy over the nature and magnitude of faulting in the southern Rocky Mountain Trench.

EXTENT AND ORIENTATION OF MYLONITIC ROCKS

Mylonitic rocks were found along the main logging road along the northeastern side of Kinbasket Lake from Dave Henry Creek southeastward until the last outcrop northwest of Yellowjacket Creek (NTS 83D/11, 14, and 15). Gneiss southeast of Yellowjacket Creek is not obviously mylonitic. The shear zone boundary is not exposed and its orientation is not known. Mylonitic rocks were not seen on the logging road along the southwestern side of Kinbasket Lake, although their presence cannot be ruled out.

The mylonites have a prominent, steeply dipping foliation and millimetre-scale compositional lamination striking subparallel to the trend of the SRMT at this latitude. The foliation contains a lineation defined by the long axes of micas and elongate domains of quartz and plagioclase feldspar. The lineation is subhorizontal; maximum plunges are less than 5° southeast.

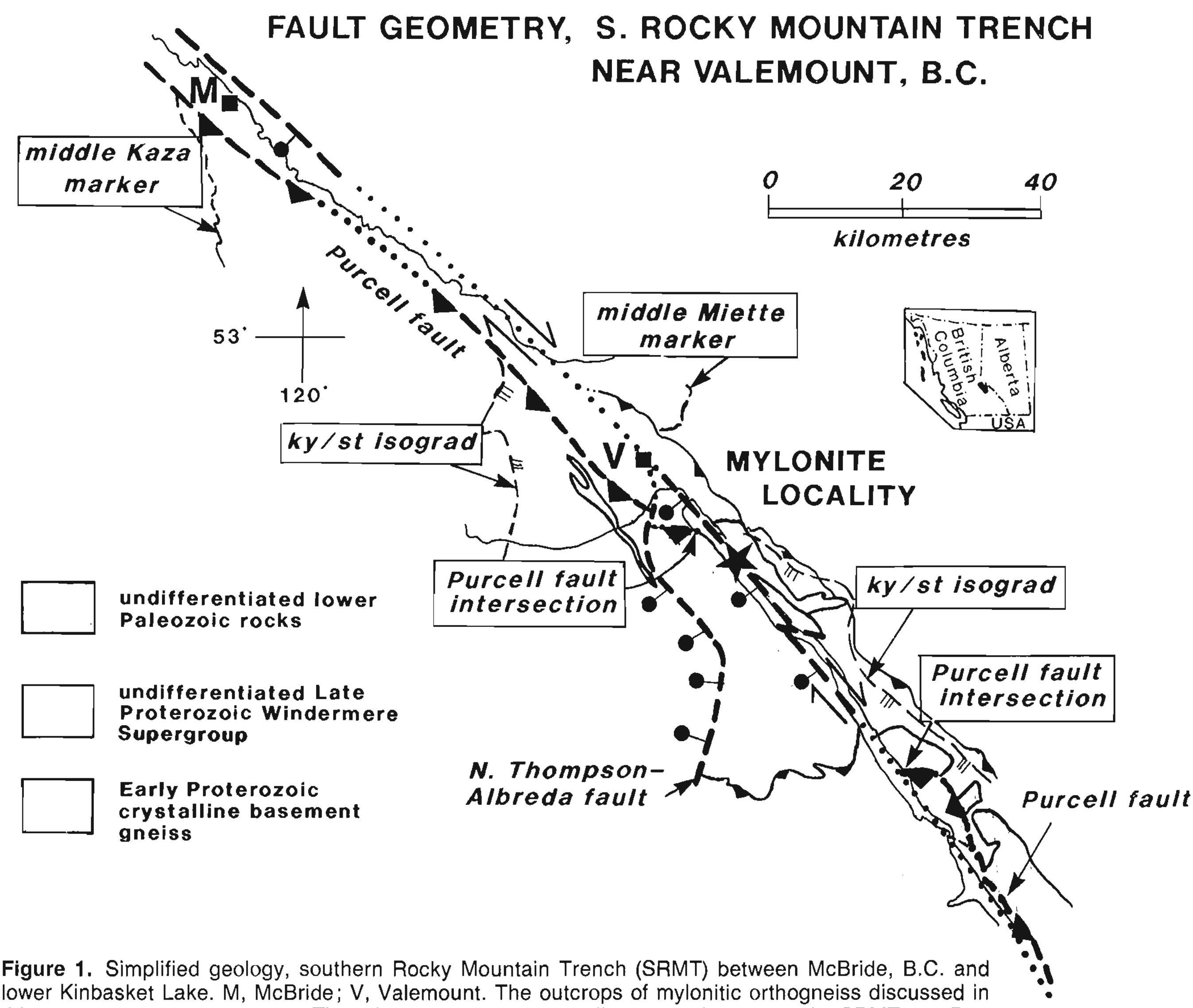

Figure 1. Simplified geology, southern Rocky Mountain Trench (SRMT) between McBride, B.C. and lower Kinbasket Lake. M, McBride; V, Valemount. The outcrops of mylonitic orthogneiss discussed in this report are shown with a star. Three features with dextral strike-separation across the SRMT are: Purcell Fault; trace of the isograd marking the first appearance of coexisting kyanite and staurolite; and the trace of a prominent marker unit within the Windermere "grit" unit (middle Kaza Group southwest of the SRMT, and middle Miette Group, northeast).

MICROTEXTURES

Figure 2 consists of photomicrographs of an oriented thin section of a sample from this locality that illustrates the textures present. The section is cut perpendicular to the prominent subvertical foliation and parallel to the subhorizontal lineation (XZ kinematic plane). The mylonitic nature of the sample is obvious. The sample shows a compositional layering defined by quartz- and feldspar-rich, mica-poor and feldspar-poor, mica-rich layers. Subparallel to the compositional layering is an undulating foliation, defined primarily by the parallel orientation of micas and the long axes of ribbons of quartz, quartz grains and subgrains, and feldspar porphyroclasts. Quartz-rich domains consist of mosaics of fine irregular grains with curviplanar and serrate grain boundaries. Undulose extinction is present in most grains. Quartz c-axes have a strong preferred orientation. Plagioclase feldspar is commonly fractured by dilational cracks or traversed by irregular bands of sericite and quartz inferred to be stress-corrosion cracks. Fragments of plagioclase commonly display undulose extinction and are variably sericitized. Two textural varieties of muscovite are present: isolated coarse grains aligned within the prominent foliation, and fine grains formed from the sericitization of plagioclase or from the deformation of coarse grains of muscovite in shear zones or zones of extension. Biotite is generally chloritized and occurs as rare coarse grains. Coarse grains of biotite and muscovite are locally sigmoidal in shape, resembling "mica fish" (Lister and Snoke, 1984). Coarse micas are locally folded by open to tight kinks and similar folds, or cut by extension fractures oriented at high angles to the prominent foliation. Apatite is a common accessory phase; it is unstrained but commonly fractured and extended. The direction of extension of micas and apatite grains generally lies within the foliation although it is locally inclined in a counterclockwise direction to the prominent foliation.

Quartz and plagioclase textures indicate reduction in grain size during deformation. The small grain size, serrate and irregular grain boundaries, and unrecrystallized lattice strain shown by quartz indicate relatively high-stress dynamic recrystallization. The presence of both brittle and ductile deformation features in plagioclase suggest conditions of deformation at or near the brittle-ductile transition for this mineral. Synkinematic minerals include sericite, quartz, calcite, and chlorite, suggesting relatively low grade conditions of deformation.

KINEMATICS

The reconnaissance nature of the mapping in this area precluded a detailed analysis of the shear zone; its extent and internal structure are both unknown. Various microstructures, however, suggest dextral displacement (Fig. 2). These features include inclined internal foliations (S-planes of Berthe et al., 1979), shear bands, asymmetrical folds of coarse micas, and pulled apart and extended porphyroclasts. As all these data come from only one thin section, the interpretation is clearly preliminary.

CONSTRAINTS ON MAGNITUDE OF DISPLACEMENT

The data suggest that the northeastern structural boundary of the SRMT near Valemount has a history of ductile, dextral strike-slip followed by down-to-the-west dip-slip faulting. Because of the similarity in lithology and metamorphic grade of rocks on both sides of the SRMT, it is unlikely that down-to-the-west displacement is large (McDonough and Simony, 1988a). As such, most of the strike-separation across the SRMT resulted from the early episode of strike-slip.

The amount of strike-slip can be estimated from the offset of the postmetamorphic Purcell Fault. At the northwestern end of Kinbasket Lake, the Purcell Fault southwest of the inferred strike-slip shear zone juxtaposes a hanging wall of Malton gneiss and unconformably overlying basal Windermere Supergroup against a footwall composed of more than 200 m of presumably Cambrian marble (Campbell, 1968; Murphy, 1984, 1990). Along the SRMT to the southeast, near Hugh Allan Creek, basement rocks (Hugh Allan gneiss) and overlying rocks of the Windermere Supergroup are juxtaposed against structurally underlying marble, clastic metasedimentary rocks, and Mount.Blackman gneiss (Oke and Simony, 1981; Mountjoy, 1988). Marble in the footwall is more than 200 m thick and, from published descriptions (Mountjoy, 1988), resembles marble in the SRMT near Valemount. Mountjoy (1988) correlated the Hugh Allan Creek marble with the middle marble of the Horsethief Creek Group, but noted that a correlation with nearby Cambrian marble is also possible. The fault that separates marble from Hugh Allan gneiss is continuous with the Purcell Fault as mapped to the southeast in the southern Park Ranges by Craw (1978); it is truncated to the northwest by the fault along the northeastern margin of the SRMT. Matching the offset segments of the Purcell Fault indicates some 55 km of strike-separation, most of which is likely due to strike-slip.

Two other features (Fig. 1) with strike-separation across the SRMT are the trace of the isograd indicating the first appearance of coexisting kyanite and staurolite, and the trace of a regionally important marker unit within the Windermere grit unit (McDonough and Simony, 1988b; Ross and Murphy, 1988). By restoring the offset Purcell Fault 55 km, the regions underlain by rocks with coexisting kyanite and staurolite are positioned across from one another even though they are on opposite sides of the Purcell Fault. The Windermere marker unit, however, is still displaced; this residual strike-separation may be due to the fact that the marker unit in the Rocky Mountains lies to the east of the Valemount strain zone, a recently described zone across which some dextral displacement is inferred (McDonough and Simony, 1989).

Restoration of 55 km of strike-slip along the SRMT brings into alignment the Purcell Fault and rocks which were metamorphosed to conditions in which kyanite and staurolite coexist. This restoration positions presumably Cambrian marble between blocks containing crystalline basement and Windermere Supergroup. The southwestern contact is the Purcell Fault; the nature of the northeastern contact of the marble is not known.

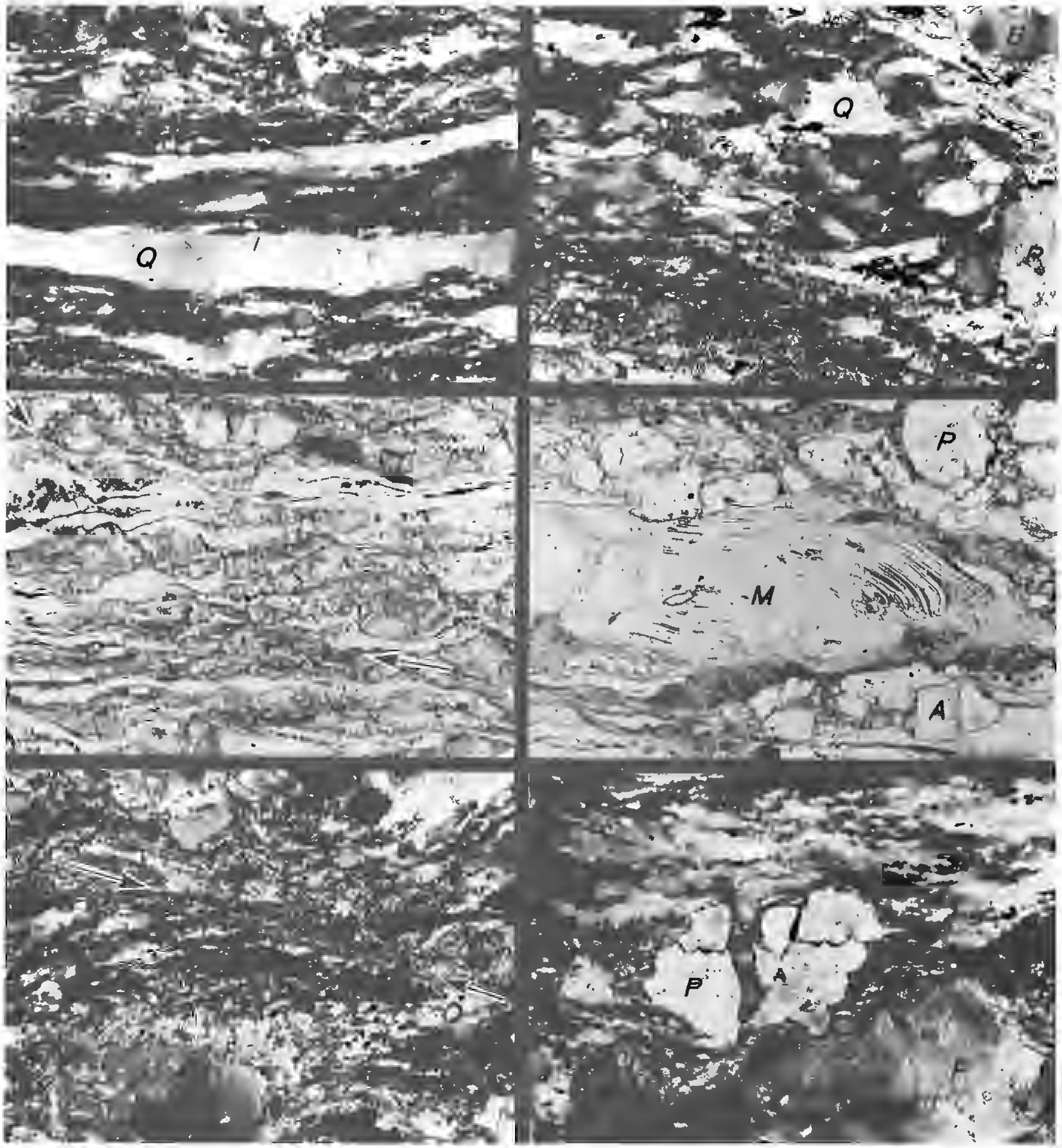

Figure 2. Microtextures of mylonitic orthogneiss. All photographs oriented with northwest to left, southeast to right. Q, quartz; P, plagioclase; B, biotite; M, muscovite; A, apatite. Upper left: quartz ribbon grains. Note bimodal grain size population, patchy extinction, and subgrain fabric locally inclined to direction of elongation of ribbons. view is about 1.5 mm across. Upper right: Serrate, interdigitating, and irregular grain boundaries in quartz indicative of dynamic recrystallization. Note bimodal grain size distribution and patchy extinction; view is about 1.5 mm across. Middle left: Fractured, pulled- apart, and sericitized plagioclase porphyroclasts. Undulose nature of foliation resulting from shear bands crossing from upper left to lower right (example indicated by arrows) and foliation wrapping around porphyroclasts. Note back-rotation of porphyroclasts above dextral shear band indicated. Plane light; view is about 7 mm across. Middle right: Kinked and sericitized muscovite. Note fractured and extended apatite at bottom of photograph; direction of extension is slightly inclined counterclockwise to direction of main foliation. Fold and direction of porphyroclast extension both indicate dextral shear. Plane light; view is about 1.5 mm across. Lower left: dextral shear band (indicated by arrows) and inclined quartz foliation. View is about 7 mm across. Lower right: Pulled apart and dextrally rotated plagioclase porphyroclast. View is about 1.5 mm across.

AGE OF STRIKE-SLIP DISPLACEMENT

The age of strike-slip displacement is not well constrained. The strike-slip zone offsets the postmetamorphic Purcell Fault. The latest phase of metamorphism in the Cariboo Mountains in the hanging wall of the Purcell Fault occurred at 135 $\pm$ 4 Ma (Currie, 1988). Mylonites are crosscut by down-to-the-west shear fractures; these are probably part of a network of down-to-the-west normal faults which includes the North Thompson-Albreda fault. Motion on the North Thompson-Albreda fault is inferred to have occurred between 51 and 45 Ma (Sevigny et al., in press). Therefore, strike-slip is constrained to between 135 and 51 Ma.

CONCLUDING COMMENT

The role of the SRMT in Cordilleran deformation is the focus of ongoing controversy. Published interpretations range from up to 1500 km of strike-slip displacement between bodies of basement gneiss in the SRMT (Chamberlain and Lambert, 1985; Lambert and Chamberlain, 1988), to little or no strike-slip and late brittle normal faulting (e.g. Campbell, 1968; Murphy, 1984; McDonough and Simony, 1988a, 1989; Price and Carmichael, 1986). This variety of interpretations reflects, in part, the paucity of outcrop in the floor of much of the SRMT. The newly discovered dextral mylonites in the SRMT indicate that dextral strike-slip can no longer be completely ruled out, but the magnitude is likely to be small. This discovery will very likely add fuel to an already lively debate.

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Preliminary results of stratigraphy, structure, and metamorphism in the southern Scrip and northern Seymour ranges, southern Omineca Belt, British Columbia

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Scammell, R.J., Preliminary results of stratigraphy, structure, and metamorphism in the southern Scrip and northern Seymour ranges, southern Omineca Belt, British Columbia; in Current Research, Part E, Geological Survey of Canada, Paper 90-1E, p. 97-106, 1990.

Abstract

Five composite stratigraphic units in the area correlate with the five traditional subdivisions of the Horsethief Creek Group. F_1 and F_2 folds are coaxial with a penetrative east-west sillimanite grade mineral lineation. West-trending upright F_3 folds are postmetamorphic. Metamorphic conditions reached or exceeded the breakdown of muscovite + quartz. Synmetamorphic and west-dipping zones of shear give top-to-the-east motion (probably related to east-vergent thrusting on the Monashee Décollement) or top-to-the-west motion of unknown origin. Down-to-the-west normal motion occurred after initial uplift on rare, discrete, west-dipping, unannealed mylonitic shear zones, and local west-verging, postmetamorphic, noncylindrical folds. Passively intruded two-mica granitic rocks are abundant. Most granitic bodies postdate F_2 folding, but synkinematic and folded granites are present. Systematic brittle deformation is of minor importance.

Résumé

Cinq unités stratigraphiques hétérogènes de la région corrèlent avec les cinq subdivisions classiques du groupe de Horsethief Creek. Les plis F_1 et F_2 sont coaxiaux avec une linéation pénétrative de direction est-ouest des minéraux de faciès de sillimanite. Les plissements F_3 droits, de direction ouest, sont post-métamorphiques. Le degré de métamorphisme a atteint ou dépassé l'isograde de muscovite + quartz. Des zones de cisaillement synmétamorphiques, à pendage ouest, donnent un mouvement du sommet vers l'est (probablement lié au charriage de vergence est sur le décollement de Monashee) ou un mouvement de sommet vers l'ouest d'origine inconnue. Un mouvement normal de la base vers l'ouest s'est effectué après un soulèvement initial sur des zones de cisaillement mylonitiques non recuites, rares, distinctes, à pendage ouest; et des plissements locaux non cylindriques, post-métamorphiques, de vergence ouest. Des roches granitiques à deux micas, intrusives d'une façon psssive, sont abondantes. La plupart des massifs granitiques sont postérieurs au plissement F_2, mais des granites syncinématiques et plissés sont présents. Une déformation fragile systématique est de faible importante.

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INTRODUCTION

The architecture and crustal evolution of once mid-crustal rocks of the Kootenay Terrane exposed in the southern Scrip and northern Seymour ranges (Fig. 1, 2) was investigated. Stratigraphy, structure, and metamorphic trends are well defined in the northern Scrip Range, in the Selkirk Mountains to the southeast of Mica Dam, and in rocks of the Monashee Terrane to the south. The study area lies between these regions, and thus may reveal relationships between them. In particular, this study explores relationships between horizontal compression and extension, peak metamorphism of Jurassic age to the east, post-Early Cretaceous to the north, and unknown age to the south, magmatism, and tectonic unroofing. The data and conclusions will be used to interpret LITHOPROBE seismic reflection data collected to the south (Fig. 1) along a line which crosses these rocks in a poorly exposed region.

The area has previously been mapped on a reconnaissance scale (Wheeler, 1965) and in local detail in the vicinity of the Ruddock Creek Pb-Zn deposit (Fyles, 1970). This paper presents some preliminary results of three months of mapping (1:20 000) in the area during 1988 and 1989.

REGIONAL GEOLOGICAL FRAMEWORK

The region comprises two major terranes, the pericratonic Kootenay Terrane, and the Monashee Terrane. These two terranes are separated by the Monashee Decollement (Read and Brown, 1981). The Monashee Terrane consists of Early Proterozoic paragneiss and orthogneiss, and an unconformable, dominantly metasedimentary cover of Late Proterozoic to possibly Early Cambrian age (Journeay, 1986; Scammell, 1986; Parrish and Scammell, 1989; Höy and Godwin, 1988). Kootenay Terrane rocks north of the study area are continental terrace rocks of Late Proterozoic age (Raeside and Simony, 1983).

The Monashee Terrane displays at least three folding events. In Kootenay Terrane and North American continental terrane rocks north and southeast of the study area, the structure is dominated by pre-peak metamorphic, southwest- to west-verging, 50 km-scale recumbent nappes (Scrip Nappe of Raeside and Simony, 1983; Cairns Nappe of Brown and Lane, 1988). These are overprinted by synmetamorphic and postmetamorphic folds. It is not clear how these structures in these regions are related.

In Monashee Terrane, metamorphic assemblages of a Barrovian facies series are overprinted by inverted (hot-side-up) Buchan-type assemblage zones (Journeay, 1986; Scammell, 1986) of unknown age. In Kootenay Terrane to the north and east of the study area mineral assemblages record Barrovian-type metamorphism with peak metamorphic pressures of 6-8.5 kb (e.g. Simony et al., 1980; Raeside, 1982). Timing of the metamorphic peak is documented to be Jurassic in the Selkirk Mountains to the east (Leatherbarrow, 1981). In marked contrast, it is suggested to be <100 Ma in the northern Monashee Mountains to the north (Sevigny et al., 1989). The study area physically links these areas with differently aged metamorphic peaks, and should provide some clues as to how they are related.

Kootenay Terrane and North American continental terrace rocks have been transported on the Monashee Decollement eastward over Monashee Terrane (see Brown et al., 1986). The Monashee Decollement is inferred to root to the west, and have an episodic history of thrusting at least as old as Late Cretaceous (Lane et al., 1989) and as young as Early Tertiary (Journeay and Parrish, 1989). Major normal faults include the west-dipping Eagle River and North Thompson faults, and east-dipping Columbia River fault. These faults have Eocene displacements (Parrish et al., 1988; Sevigny and Simony, 1989).

Intrusive granitic rocks are very common in this region. Their ages include Paleozoic, Middle Jurassic, Mid-Late Cretaceous, Late Cretaceous to Early Eocene, and Early-Middle Eocene (Read and Wheeler, 1976; Okulitch, 1985; Carr, 1989; Sevigny et al., 1989).

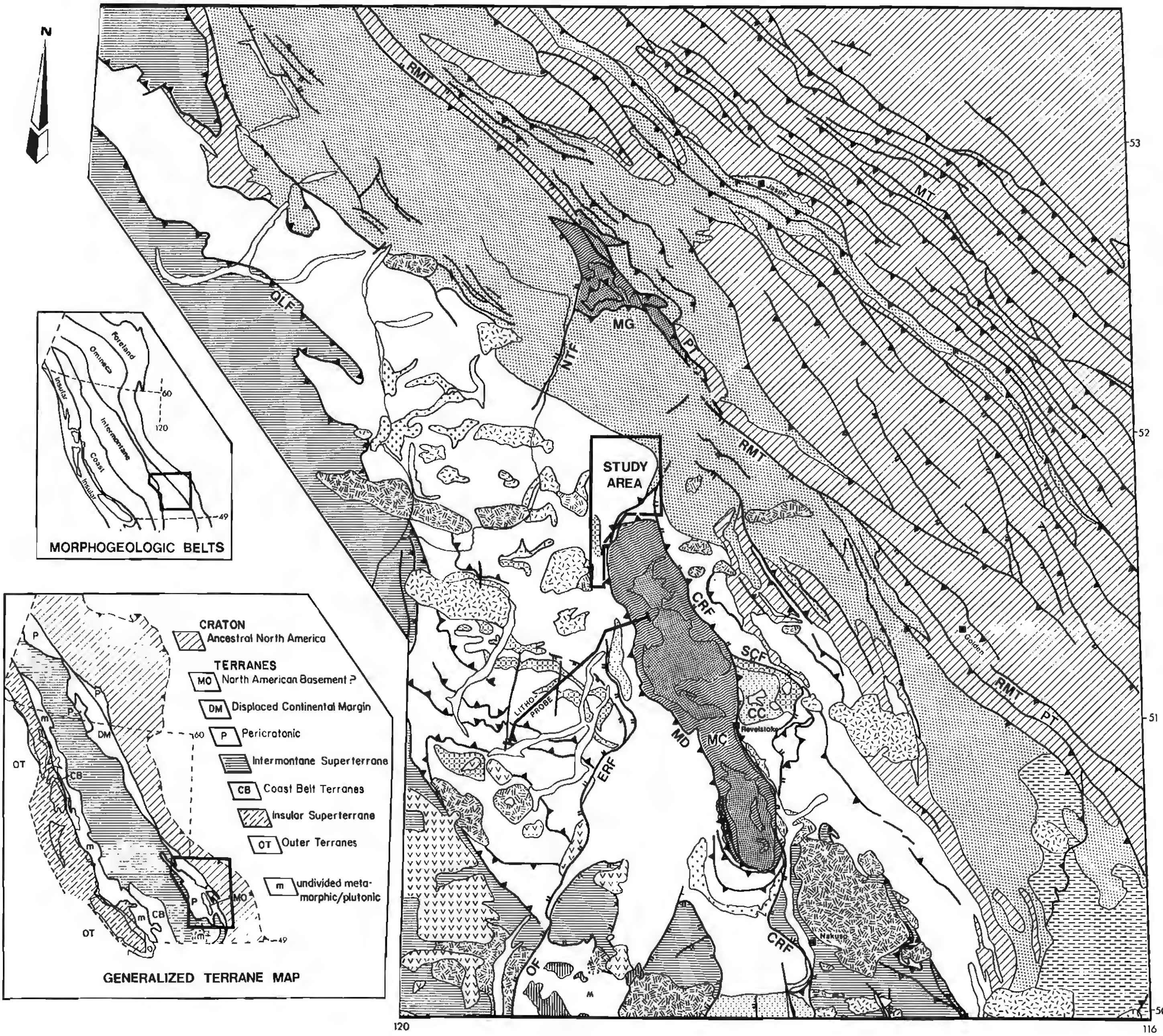
STRATIGRAPHY

The study area is dominantly underlain by a metamorphosed heterogeneous sequence of stratified rocks. Metamorphic rock types include pelitic, quartzofeldspathic, siliceous, calcareous, calc-silicate, mafic, and ultramafic varieties. Scammell and Dixon (1989) delineated five composite stratigraphic units. Mapping in 1989 has extended these units into the Seymour Range and north to the alpine ridges south of Pat Creek (Fig. 2). The succession is described in ascending structural order.

Unit 1, 600-2000 m thick, lies above a splay of the Monashee Decollement (termed MD2, Journeay and Brown, 1986; J.M. Journeay, pers. comm., 1987). It comprises interlayered pelitic ($>50\%$ mica) and semipelitic (5-50% mica) schist with minor quartzofeldspathic gneiss, amphibolite, calc-silicate and ultramafic rock. Within the central part of unit 1, a discontinuous subunit (0-500 m thick) is composed of interlayered calc-silicate gneiss, amphibolite, and sporadic boudins of brown weathering ultramafic rock (J.M. Journeay, pers. comm., 1987).

Unit 2 is composed of centimetre-scale impure marbles and calc-silicate gneiss with minor pelite, calcareous pelite, and semipelite schist. It is locally discontinuous, up to 10 m thick, and has interlayered contacts with bounding units. This unit is one of two regionally important calcareous marker units.

Unit 3 is >1000 m thick and locally as thick as 2300 m. It is dominated by an interlayered succession of semipelitic schist, amphibolite, and hornblende gneiss. Subordinate rock types include pelitic schist, calc-silicate, quartzofeldspathic gneiss, quartzite, rare ultramafic pods, and quartz pebble to boulder paraconglomerate interlayered on a centimetre- to metre-scale. Quartzofeldspathic rocks commonly display biotite and biotite-garnet seams. Pelitic schist is generally relatively aluminosilicate-poor. Although commonly discontinuous, some subunits dominated by amphibolite and rusty pelitic schist can be traced for several kilometres. Amphibolite gneiss (garnet, biotite) ranges from a few millimetres up to five metres thick. Most are pulled apart. Contacts are either sharp or gradational. Discordant amphibolite sheets have not been observed. Tex-



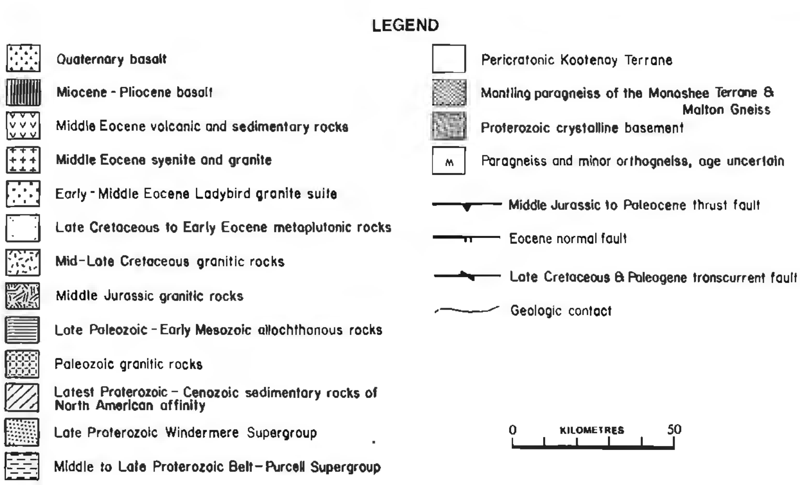


Figure 1. Location of the study area and regional geology (modified after Wheeler and McFeely, 1987, Parrish et al., 1988). Faults: OF = Okanagan Fault, PT = Purcell Thrust fault, MT = McConnel Thrust fault, MD = Monashee Decollement, ERF = Eagle River fault, SCF = Standfast Creek fault, CRF = Columbia River fault, QLF = Quesnel Lake fault, NTF = North Thompson fault. Metamorphic complexes: MC = Monashee Complex (or terrane), CC = Clachnacudainn Complex, MG = Malton gneiss complex. RMT = Rocky Mountain Trench.

tures include finely layered to massive varieties. Thin interlayered marble and calc-silicate horizons occur near the top of unit 3. The unit is capped by a 5-30 m thick horizon of sillimanite-rich rusty pelitic schist.

Unit 4, a distinctive unit that has been mapped from the Seymour Range to south of Pat Creek, is a second composite calcareous marker horizon in the region. It overlies rusty pelitic schist of unit 3 along an interlayered contact. The unit is 50-1000 m thick and dominantly impure marble and calc-silicate with subordinate rusty pelitic schist, semipelitic schist, quartzofeldspathic gneiss, quartzite, and ultramafic boudins. Thickness and rock types vary along strike; consequently the internal stratigraphy of this horizon is not known in detail. A variety of 10 cm to 1 m thick marbles are present. Some can be traced along strike for several kilometres. They range from massive, pure, grey to white weathering marble, to impure grey-and buff-weathering fetid marble. Accessory phases in impure marble includes quartz, diopside, plagioclase, garnet, graphite, and epidote.

The Ruddock Creek Pb-Zn sulphide horizon is inferred to be one of the structurally highest subunits of unit 4. This discontinuous, stratiform, sulphide-bearing subunit is generally 2-5 m thick, and is well described by Fyles (1970). Ultramafic rocks are found above and below the sulphide-bearing subunit. These ultramafic rocks are typical of all ultramafic rocks found sporadically throughout the five mapped units. They are found as metre- to 10 metre-scale, foliated to massive, and fine- to very coarse-grained layer-parallel pods. Ultramafic rocks are composed of orthopyroxene, clinoamphibole, olivine, chlorite, talc, and serpentine. Discontinuous, metre-scale marble horizons mark the top of unit 4.

Unit 5 is at least 300 m thick, and possibly much thicker as its upper limit has not been determined. It is dominated by buff- and rust-coloured semipelitic schist and quartzofeldspathic gneiss interlayered with biotite pelitic schist, garnet-biotite seams, metre-scale amphibolite, and rare ultramafic rock. Amphibolites form $<15\%$ of the mapped part of this unit, and tend to be massive units of metre-scale thickness that are laterally continuous for up to 100 m. Thinner and laminated varieties of amphibolite and hornblende gneiss of the type found in unit 3 are not present in the lower part of unit 5. Some quartzofeldspathic gneiss contains dispersed feldspar and quartz grains 0.2-2 cm in diameter which indicate that these rocks are metamorphosed pebble-granule conglomerates.

Correlation

Based on initial mapping in the southern part of the area, and comparison with published descriptions of rocks of the Horsethief Creek Group (e.g. Raeside and Simony, 1983), Scammell and Dixon (1989) correlated these five units with the five traditional subdivisions of the Horsethief Creek Group. South of Pat Creek, calcareous unit 4 lies within 4 km along strike of the middle marble division mapped by Raeside (1982) at the headwaters of Pat Creek (Fig. 2). Bounding units in the two areas display similar rock types and stratigraphic relationships (Scammell and Dixon, 1989). In view of these similarities, and of the knowledge that the style of regional deformation has generated panels of rock with lateral continuities of >10 km, rocks of the study area and the late Proterozoic Horsethief Creek Group can be correlated with confidence. Units 1-5 correlate with the lower clastic, lower carbonate, semipelite-amphibolite, middle marble, and upper clastic divisions respectively.

The above correlation and the geometry of units in the map area indicate that an essentially upright panel (complicated by later folding, see below) of Horsethief Creek Group underlies the region from the hinge of the Scrip Nappe to the northwestern flank of Monashee Terrane. This strengthens the case for correlations, based on stratigraphic similarity, of stratigraphy that lies in fault contact on the southern and western flanks of the Monashee Terrane, with Horsethief Creek Group rocks (Journeay, 1986; Carr, 1989; Johnson, 1989).

STRUCTURE

Four phases of folding are identified. F_1 and F_2 synmetamorphic folds are penetrative throughout the area, whereas postmetamorphic F_3 and F_4 folds are relatively localized. Widespread brittle deformation and narrow discrete ductile shear zones play a minor role in the strain history.

Macroscopic folds

Interpretation of the geometry of macroscopic folds in the study area is in a preliminary stage. Folds present are 10 km-scale structures delineated mostly by marker horizons. The coaxial nature of folding and the difficulty in determining the order of mesoscopic parasitic folds severely restrict the use of mesoscopic fold vergence in determining the geometry of larger structures.

From the Seymour Range to Ruddock Creek in the southern part of the study area, the whole stratigraphic sequence lies in a west- to northwest-dipping panel in the immediate hanging wall of the Monashee Decollement (Fig. 2). At Ruddock Creek, the Pb-Zn horizon of unit 4 outlines a kilometre-scale type-3 (Ramsay, 1967) interference pattern (Fyles, 1970). Our mapping of calcareous layers lower in unit 4 confirms this structure. The initial orientation of the F_1 structure is not known. It is folded by several kilometre-scale reclined F_2 folds with gently plunging fold axes trending west-northwest.

In the northwestern part of the area, rocks of units 3 to 5 lie in a gently to moderately southwest-dipping upright panel (Fig. 2). These rocks are similar in orientation to correlative horizons of the Horsethief Creek Group mapped to the northwest (Raeside, 1982). There they are inferred to lie in the overturned limb of the Scrip Nappe, reoriented by F_2 folding into an upright, gently to moderately southwest-dipping orientation. The above observations and correlations imply that the axial surface trace of the Scrip Nappe lies in the Pat Creek valley, and that the trace of the inferred underlying synclinal mate should lie southwest of this panel.

Although calcareous marker horizons are common in the northeastern part of the area, and unit 4 is recognized to

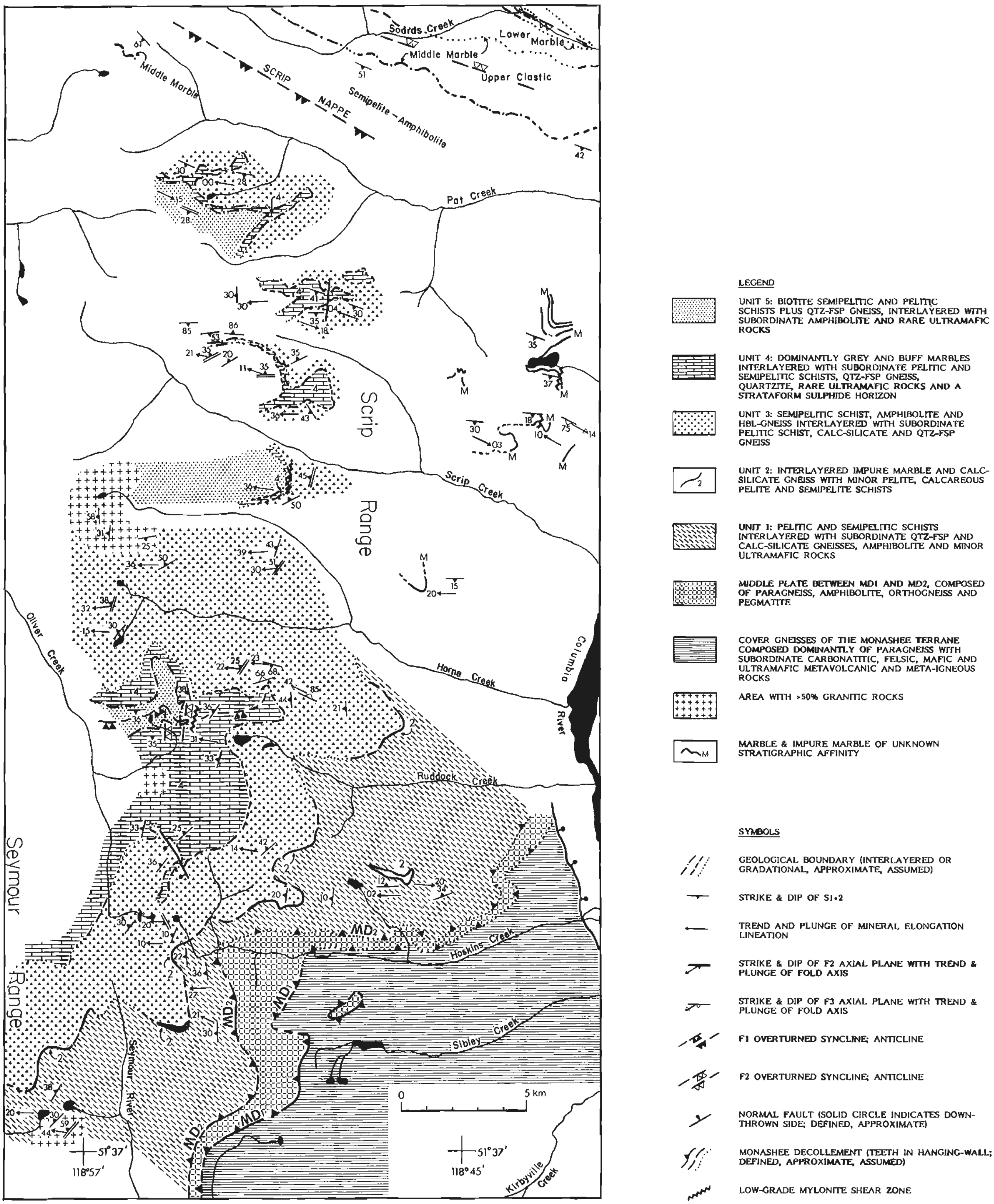


Figure 2. Preliminary geological map of the study area. Located approximately in the bounds of the study area shown in Figure 1. Geology north of Pat Creek is after Raeside (1982); the marble layer of unknown stratigraphic affinity north of Horne Creek, and south of Scrip Creek, is after Wheeler (1965); the location of the Monashee Decollement (MD2) is after J.M. Journeay (pers. comm., 1987).

strike across the whole area, refinement of correlations and further orientation data are required before the macroscopic structure between Ruddock Creek and the ridge north of Scrip Creek can be outlined with confidence.

Mesoscopic folds, foliations, and lineations

Four phases of folding are identified at the mesoscopic scale. F_1 folds are preserved as rare centimetre- to decimetre-scale rootless isoclinal folds, and more commonly in type 3 interference patterns with F_2 folds. A regional metamorphic foliation is marked by schistosity and gneissic layering. It is defined by the dimensional preferred orientation of phyllosilicate and prismatic mineral grains, plastically deformed grains, and concordant lenses of leucosome. It generally parallels primary compositional layering in the metasedimentary rocks, and the limbs of F_1 and F_2 folds. This composite foliation plane is inferred to be a product of transposition and/or synkinematic recrystallization during folding of F_1 by F_2 folds. It is designated S_{1+2}. Dipping moderately to the southwest to northwest, poles to S_{1+2} define a girdle pattern in stereographic projection (Fig. 3a). This dispersion can be explained by superposed F_3 folding about an axis plunging gently to the west (Fig. 3e).

F_2 folds generally fold a phyllosilicate foliation. In some cases, phyllosilicate-rich layers display a crenulation cleavage from the folding of the F_1 axial planar foliations by F_2 folds. Axial-planar biotite is locally developed. F_2 folds are close to isoclinal, similar in geometry, and exhibit angular and curved hinges. They have gently to moderately dipping axial planes (S_2, Fig. 3c). Similarly to S_{1+2} in orientation, poles to S_2 plotted in stereographic projection display a weak girdle of points due to F_3 folding. F_2 fold axes, L_2, generally plunge gently to the west-northwest, and less commonly, gently to the east-southeast (Fig. 3c). Some scatter in L_2 data is a product of localized F_3 and F_4 folding. L_2 is invariably parallel to a strong mineral lineation (LS, Fig. 3b). This lineation is defined by the parallel alignment of sillimanite fibres, mica aggregates, hornblende, and quartz rods and is best developed in pelitic and quartzofeldspathic units. The collinear nature of LS and L_2 throughout the region indicates that F_2 folding and sillimanite growth were synchronous.

A third phase of folding is recognized locally throughout the region. Noncylindrical F_3 folds generally have upright axial planes (S_3, Fig. 3e), and axes plunging gently to the west or southeast (L_3, Fig. 3e). F_3 folds are broad, open structures. They are commonly disharmonic in pelitic/quartzofeldspathic multilayers. S_3 folds exhibit axial-planar crenulations of S_{1+2} in phyllosilicate-rich layers. Where F_3 fold axes are inclined to LS, they are folded. Phyllosilicates, sillimanite, and feldspars are bent or broken; the third phase of deformation is therefore postmetamorphic. No megascopic F_3 folds have been documented.

Rare noncylindrical, overturned, east- and west-verging, late metamorphic to postmetamorphic folds reflect a fourth phase of folding. These folds are similar in fabric relationships to F_3 folds, but differ in orientation. They are also spatially associated with discrete brittle and ductile shear zones.

Mylonitic fabrics

Mylonitic fabrics are locally developed throughout the area. In the southern part of the region, both upper-plate-to-the-east and -west, high grade mylonitic shear zones occur parallel to the westerly dipping S_{1+2} foliation. Kinematic indicators are visible on surfaces viewed perpendicular to S_{1+2}, and parallel to LS. C-surfaces (Berthe et al., 1979) are defined by seams of intergrown and aligned biotite and sillimanite, and S-surfaces (op. cit.) are defined by inclined foliation and mica fish. Shear bands (op. cit.) are common. These kinematic indicators are visible at both mesoscopic and microscopic scales.

Metre-scale, low grade proto-ultramylonitic shear zones were observed in four localities. These shear zones dip moderately to steeply northwest to southwest, and are localized in pegmatitic granitoid rocks and impure marble horizons. They deform granitic rocks which truncate S_{1+2} foliation and F_2 folds. Unannealed quartz ribbons and brittly deformed feldspars are observed in thin section. Asymmetry of rare muscovite mica fish, foliation fish, and shear bands indicate down-to-the-west normal motion. Dip-slip displacements range from several metres to tens of metres.

Brittle deformation

A systematic array of joint sets is present throughout the region. Joints range in size from areas of a few centimetres, to tightly spaced arrays defining metre-thick fracture zones with areas of several thousand square metres. The most prominent array consists of steep east- and west-dipping conjugate sets (Fig. 3f). They are compatible with east-west extension. Basaltic dykes tend to be intruded along this system of joints.

Commonly throughout the area, fractures parallel to the above joint array are decorated with imbricate quartz-fibre lineations indicating both down-to-the-west and down-to-the-east normal displacement. Dip-slip displacements are generally on the order of centimetres, but can be hundreds of metres (Fyles, 1970). The latter are marked by metre-thick fracture zones.

Regional relationships

Folds in the study area are similar in style, trend, and fabric relationship to those mapped to the north (Raeside, 1982; Raeside and Simony, 1983; Sevigny and Simony, 1989). East- and west-verging, high grade shear zones in the study area are similar in metamorphic grade, mineral assemblages, and fabrics to mylonitic rocks documented in Monashee Terrane (Scammell, 1986), and to those associated with late east-vergent thrusting on the Monashee Decollement (MD2, J.M. Journeay, pers. comm., 1987). Top-to-the-east shear zones in the study area are most likely associated with east-vergent thrusting of the region over Monashee Terrane. The significance of high grade, down-to-the-west normal motion is not presently known.

The presence of west-dipping, normal displacement, low grade ductile shear zones in the central part of the study area suggests that unroofing of this high grade terrane was accomplished by motion on faults of similar dip and kinematics. The low grade of these shear zones indicates that west-directed ductile unroofing outlasted east-directed unroofing (if any) in this region. Rocks in the hanging wall of the Eagle River fault, or the North Thompson fault, or high grade rocks now situated farther to the west, may have formed the cover to this region.

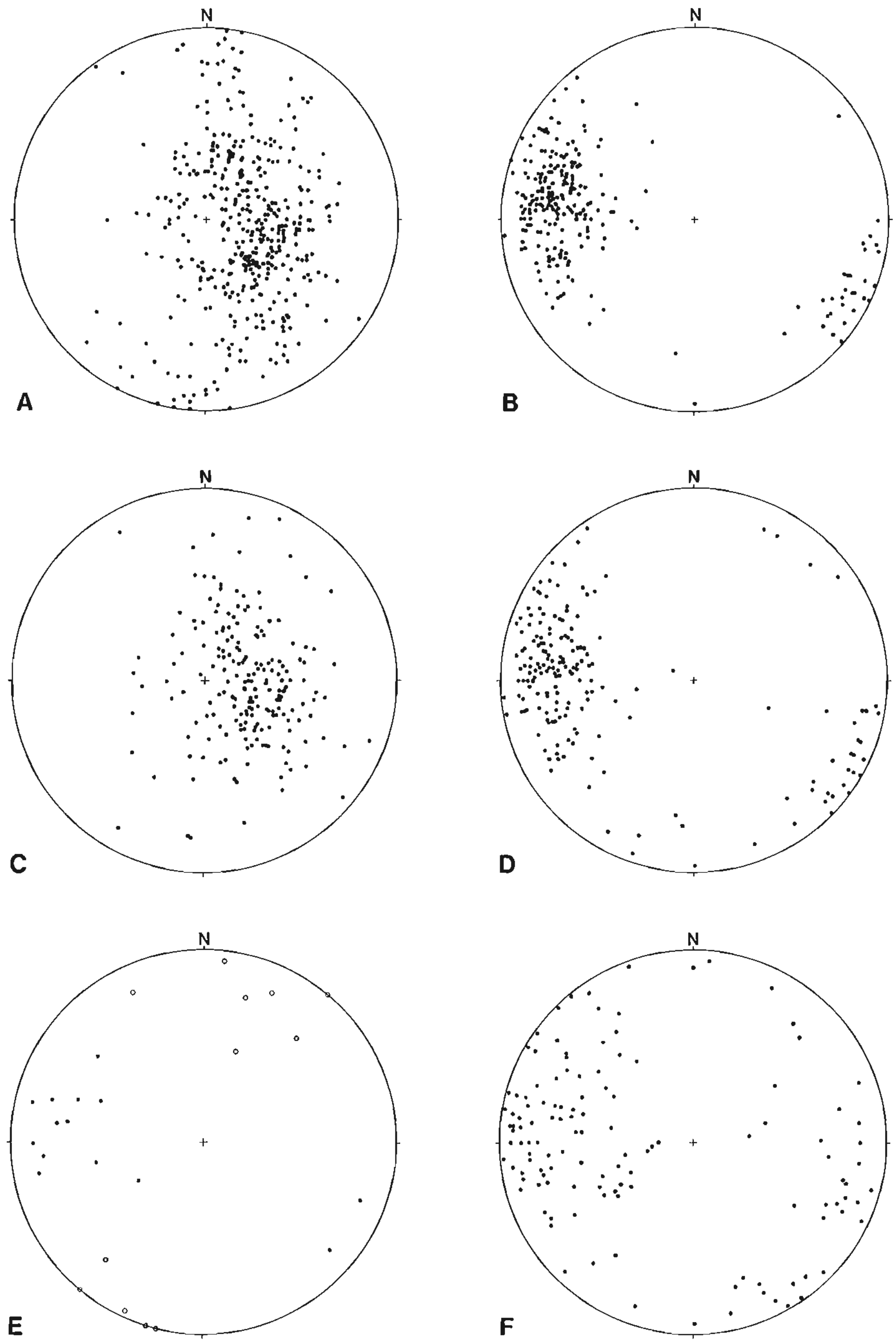


Figure 3. Lower hemisphere equal area stereonet compilation of planar and linear fabric element data. A) Poles to S_{1+2} foliation; n = 414. B) Mineral elongation lineation, LS; n = 248. C) Poles to F_2 axial planes, S_2; n = 197. D) F_2 fold axes, L_2; n = 190. E) Open circles are poles to F_3 axial planes, S_3, n = 12; dots are F_3 fold axes, L_3; n = 15. F) Poles to joint planes, N = 128.

METAMORPHISM

Most rocks in the area are migmatites as defined by Ashworth (1985). Processes of migmatization have dominantly affected pelitic and semipelitic rocks, and to a lesser degree, amphibolites which display leucocratic and melanocratic layers. Migmatites are usually irregularly layered parallel to S_{1+2}. Less commonly, leucosome disrupts this foliation, enveloping rafts of melanosome. Leucosome compositions include trondhjemite, granodiorite, and granite. Magmatic injection, anatexis, and metamorphic differentiation contribute to migmatite generation in the area.

Metamorphic mineral assemblages in pelitic rocks comprise quartz-plagioclase-potassium-feldspar-garnet-sillimanite-muscovite. Sillimanite is most commonly intergrown with biotite. It can be found as centimetre-thick mats on discrete surfaces. Garnet is generally subhedral with poikilitic cores. Muscovite is present either intergrown with biotite in S_{1+2} foliation planes, or as mica fish in mylonitic rocks associated with LS, the sillimanite mineral lineation. Muscovite in this form may be prograde remnants, or mimetic retrograde products. The second form of muscovite crosscuts S_{1+2} foliation, and is therefore interpreted to be secondary or retrograde in origin. The presence of granitic melt, potassium-feldspar, and sillimanite throughout the region indicate that these rocks equilibrated above the muscovite-out isograd. The possible presence of prograde muscovite indicates conditions were close to the muscovite-out isograd. Late muscovite may be related to intrusion of late granites. Future assessment of the physical conditions of metamorphism will be performed using electron microprobe determinations of the mineral compositions of these metamorphic assemblages.

GRANITIC ROCKS

Leucocratic granitoid rocks are ubiquitous, and are estimated to comprise ca. 30 % of the rocks in the region. They can comprise 100 % and up to 80 % of the exposure at mesoscopic and macroscopic scales respectively. They can be found as (i) sheared discontinuous concordant bodies, (ii) sheared discordant bodies, (iii) deformed and undeformed pods with gradational boundaries in other granitoid rocks, and (iv) undeformed crosscutting dykes and pods, indicating prekinematic, synkinematic, and postkinematic intrusion.

Granitic rocks generally appear to range in composition from quartz monzonite to granite. Biotite and/or muscovite are always present. Minor phases include garnet, sillimanite, and tourmaline.

Local unambiguous relationships appear to be regionally consistent. The oldest granitoid rocks are medium grained, and are generally but not invariably, foliated parallel to S_{1+2}. They are present as centimetre- to metre-thick layer-parallel pods that are commonly folded by F_2 folds. Rare lineated sillimanite is present. These granites comprise $< 10\%$ of the granitic rocks.

Coarse grained to pegmatitic (> 1 order of magnitude grain size contrast with host rocks), foliated muscovite-biotite granite is present as layer-parallel, metre-thick sheets. It may be related to the above-mentioned medium grained granitic rocks. These rocks are crosscut by medium grained, biotite and muscovite-biotite granitic rocks that are distinguished in the field by their light to medium grey, and less commonly buff colours. These grey granitic rocks contain up to 25 % biotite. Their fabrics range from subhedral-granular to weakly foliated and/or lineated. They can be layer-parallel, but most often are found as 10 m-scale stocks which cut across the S_{1+2} foliation. Intrusions of this rock several hundred metres thick occur in the Seymour Range. Regionally they make up $< 10\%$ of the granitic rocks.

The youngest and most common granitic rocks are found as either subhorizontal or nearly upright sheets that can be traced for hundreds of metres, and display thicknesses in the range of one to ten metres. They are generally discordant with respect to S_{1+2} foliation and LS lineation. Textures are subhedral-granular. Pegmatitic varieties locally contain biotite and feldspar crystals up to 30 cm in length. Decimetre-thick, discontinuous layering is defined by laterally continuous medium grained to pegmatitic domains. These domains are gradational into each other. These granitic intrusions which commonly parallel the S_{1+2} foliation, can generally be traced into areas where they are discordant with respect to S_{1+2} foliation and LS lineation. Intrusion along the layering gives the appearance that these granitoid rocks have been folded by F_2 folds. Often these nonfoliated, apparently folded granitic rocks can be traced into areas where they display discordant contacts. They crosscut all of the above granitic rocks. Based on these observations, intrusion and crystallization of these granitic melts are inferred to have outlasted F_2 folding.

None of the above granitic rocks show contact metamorphic effects on host rocks implying intrusion into a relatively hot sequence. Passive intrusion is indicated by the fact that the S_{1+2} foliation maintains a consistent attitude both close to, and distant from the intrusions, and by the presence of similarly oriented rafts of the layered sequence within the granitic rocks.

Regional relationships of granitic rocks

Granitic rocks are ubiquitous in the high grade tectonostratigraphic sequences that structurally overlie exposures of deeper crustal level rocks in the region. To the north of the area, Sevigny et al. (1989) document relatively rare 100.4 $\pm$ 0.3 Ma old synkinematic granites, and more common 63 $\pm$ 1 Ma old postkinematic granites. Similar granites are common in the hanging wall of the Monashee Decollement along the western flank of the Monashee Terrane (e.g. Journeay, 1986; Johnson, 1989). Carr (1989) also documents extensive ca. 55-59 Ma old Ladybird leucogranite in the hanging wall of the Monashee Decollement to the southwest of the Monashee Terrane. The granitic suites identified in the field area are most likely members of the above dated suites. A U-Pb chronological study of these granitic rocks has been initiated.

SUMMARY

Kootenay Terrane rocks between Monashee Terrane and the axial surface trace of the Scrip Nappe are multiply-deformed migmatitic metasediments. Five composite stratigraphic units strike across the region from west of the Monashee Terrane to the Scrip Nappe. They are composed dominantly of metamorphosed pelitic, quartzofeldspathic, and carbonate sedimentary rocks, and subordinate mafic and ultramafic intrusive and/or extrusive rocks. They are correlated with the five traditional late Proterozoic Horsethief Creek Group subdivisions, and thus establish a stratigraphic link with North American continental terrace sedimentary prism.

Mesoscopic rootless isoclinal folds and a macroscopic synformal structure inferred from stratigraphic symmetry are F_1 folds. Superposition of synmetamorphic F_2 folds has generated macroscopic and mesoscopic coaxial (type 3) interference patterns. F_1 and F_2 fold axes parallel a penetrative, west- and east-trending, sillimanite grade mineral lineation. Mesoscopic postmetamorphic west-trending open F_3 folds, and noncylindrical overturned F_4 folds are locally developed.

High grade, upper-plate-to-the-east and -west shear zones are present. East-vergent motion is most likely related to east-vergent thrusting of the Kootenay Terrane over the Monashee Terrane. The significance of the high grade, down-to-the-west normal shear zones is not known. Low grade, down-to-the-west normal motion continued after initial uplift of the region. Brittle structures within the study area are a minor component of the deformation history.

Metamorphic grade throughout the area peaked at or above the breakdown of muscovite. Layer-parallel and discordant, medium grained to pegmatitic granitic rocks comprise ca. 30% of the rocks in the region. Most of these two-mica granitoid rocks are post-F_2 folding. Synkinematic and folded granites are present but relatively rare.

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Studies of the Kluane and Nisling assemblages in Kluane and Dezadeash map areas, Yukon

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Abstract

The structure and metamorphic character of four areas underlain by the Nisling assemblage and the Kluane Schist were studied in Dezadeash and Kluane map areas. The protoliths of the two assemblages differ. Their metamorphic grade is locally similar, and both terranes preserve sillimanite-bearing rocks. There is no evidence for a large east-dipping thrust that would place undivided metamorphic rocks on the Kluane Schist in the east. The contact of the Kluane Schist with the Nisling assemblage is obscured by younger plutons.

The Nisling assemblage occupies a larger part of Dezadeash map area than previously thought. The Kluane Schist is in fault contact with the Dezadeash Group, and no metamorphic gradation is seen. The Dezadeash turbidites do not appear to be the protolith of the Kluane Schist.

Résumé

La structure et le caractère métamorphique de quatre régions dont le sous-sol est constitué par l'ensemble de Nisling et les schistes de Kluane ont été étudiés dans les régions cartographiques de Dezadeash et Kluane. Les roches mères des deux ensembles sont différentes. Leur degré de métamorphisme est, par endroits, similaire et les deux terranes conservent des roches sillimanitifères. Il n'existe aucun indice d'un grand chevauchement, à pendage est, qui placerait des roches métamorphiques non divisées sur les schistes de Kluane à l'est. Le contact des schistes de Kluane et de l'ensemble de Nisling est caché par de jeunes plutons.

L'ensemble de Nisling occupe une plus grande partie de la région cartographique de Dezadeash qu'on ne le pensait auparavant. Les schistes de Kluane sont en contact par faille avec le groupe de Dezadeash et on ne remarque aucune gradation métamorphique. Les turbidites de Dezadeash ne semblent pas être les roches mères des schistes de Kluane.

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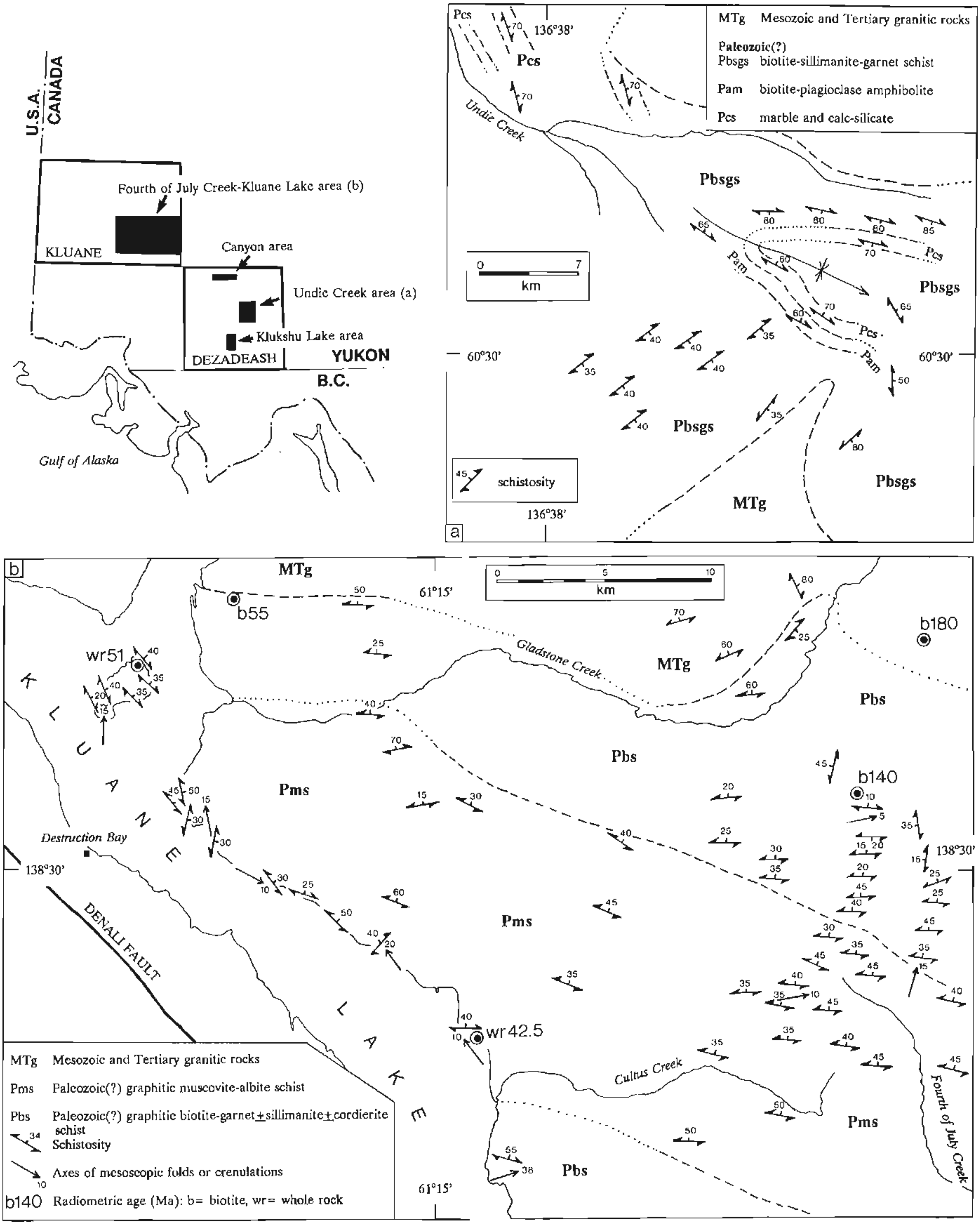

Figure 1. Location of Kluane and Dezadeash map areas and the four study areas, and results of reconnaissance mapping in the Undie Creek and Fourth of July Creek-Kluane Lake areas. K-Ar dates from Tempelman-Kluit and Wanless (1975) and Farrar et al. (1988).

INTRODUCTION

As part of the continuing study of the structural and metamorphic history of the oldest tectonic components of the northern Cordillera in southwestern Yukon, fieldwork was carried out in parts of Kluane Lake and Dezadeash map areas during the summer of 1989 (Fig. 1). The Proterozoic to lower Paleozoic Nisling assemblage, the oldest known in the area, records pre-Triassic sillimanite -grade metamorphism and penetrative deformation in a supracrustal succession of passive continental margin character. The Nisling assemblage may be an uplift of displaced or autochthonous Precambrian North American basement, or part of an exotic circum-Pacific continental assemblage (Erdmer, 1989). The Nisling assemblage is juxtaposed on its west side with the Kluane Schist (Tempelman-Kluit, 1976), a metamorphic terrane of unknown age and affinity interpreted as a metamorphosed part of the Late Jurassic and Early Cretaceous Dezadeash Group to the west (Eisbacher, 1975, 1976).

A reconnaissance during previous fieldwork suggested that the metamorphic grade of parts of the two schist assemblages is similar, although protoliths are distinct, and that the regional distribution of rocks characteristic of each assemblage is poorly constrained. A large area of separate, undivided metamorphic rocks exposed between the two assemblages was interpreted by previous workers to be in the hanging wall of an east-dipping thrust (unit "m", shown to the east of Kluane Lake on Wheeler and McFeely's (1987) map). The interpretation of the Kluane Schist as a metamorphosed part of the Dezadeash Group is in conflict with mid-Jurassic isotopic ages for parts of the schist (e.g. Tempelman-Kluit and Wanless, 1975). In addition, the Dezadeash turbidites are only mildly affected by mid-Cretaceous and Early Tertiary granites west of Dezadeash Lake (G. Lowey, pers. comm., 1989), suggesting that young plutons of the Coast belt are high-level intrusions and that post-mid-Jurassic metamorphism was not of regional extent. The contact between the Kluane Schist and the Dezadeash Group was reported by previous workers to be gradational (e.g. Kindle, 1952; Eisbacher, 1975), although faulted during younger events. Reconnaissance showed that gradation is unclear, and that the contact could be interpreted in other ways.

In light of these observations, fieldwork was slated in 1989 to study the nature, structure, and metamorphism of rocks in four selected areas: between Fourth of July Creek and Kluane Lake, near Undie Creek, near Klukshu Lake, and near Canyon. The initial results of mapping in the first two areas are shown on Figure 1, and general observations are summarized below.

FOURTH OF JULY CREEK-KLUANE LAKE AREA

The contact between quartz-muscovite schist and quartz-biotite schist east of Kluane Lake, shown as a steep-dipping fault on Muller's (1967) map, is not a late or brittle structure. The schistosity on both sides is parallel and moderately northeast dipping in a regional homocline several kilometres thick that extends across strike from Cultus Creek to Gladstone Creek. The two schist assemblages are graphitic, and on the basis of field examination their metamorphic grade is slightly different.

The main rock type along Kluane Lake is silvery grey, graphitic muscovite-albite schist (Fig. 2). It is mainly medium grained, with fine- grained quartz veins parallel to the main schistosity that are locally isoclinally folded. A steep northeast-dipping cleavage that is axial planar to southwest-verging asymmetrical crenulations and small folds is exposed in many shoreline outcrops (Fig. 3). Fold axes have gentle northwest plunges. Apple-green actinolite boudins and layers a few tens of centimetres to a few metres in width are interfoliated with the muscovite schist in many places. Folded biotite granite dykes cut the schistosity in some outcrops.

The other rock type, mainly northeast of the headwaters of Fourth of July Creek, is biotite-sillimanite-cordierite-garnet schist which is commonly coarse grained and locally granitized near contacts with younger plutons. It has a blue-grey colour when fresh, due to the abundant graphite-bearing cordierite porphyroblasts which give it a knotted weathering surface. It weathers to a distinctive purplish-brown or deep wine colour. Where the maximum-phase assemblage (biotite-sillimanite-cordierite-garnet) is not present, the assemblage biotite-garnet-sillimanite is common. Local evidence of layer-parallel shear and extension is preserved in both schist assemblages (Fig. 4, 5, 6).

Two small crosscutting serpentinized peridotite plugs, the largest a few tens of metres across, were seen in the schist.

UNDIE CREEK AREA

The rocks present around the headwaters of Undie Creek are similar to those of the Nisling terrane exposed near Kusawa Lake to the southeast (Erdmer, 1989). The main rock types are biotite-sillimanite-garnet schist, diopside-grossular calc-silicate, and biotite amphibolite. Garnet in schist is invariably mauve. Migmatite or diatexite is common close to intrusive contacts (Fig. 7). For several hundred metres away from granite contacts, the schist is finer grained and sillimanite is absent, suggesting that the plutons imposed retrograde aureoles on the schist. Intrusive contacts are not sharp in outcrop, and are characterized by abundant layer-parallel injection of sills into the schist.

Both map-scale folds and mesoscopic crenulations that are moderately southeast- plunging are truncated by granite plutons of the Coast Plutonic Complex. This pre-intrusive, Mesozoic or older fabric is similar to that in the Nisling terrane along strike to the southeast in Yukon and British Columbia, where it is pre-Triassic in age.

KLUKSHU LAKE AREA

Thickly bedded epidote-chlorite meta-andesite exposed on Klukshu Lake (Kindle, 1952) is unlike any rock type occurring to the east in the Nisling terrane. It is intercalated with grey, locally phyllitic, siltstone and slate which are lithologically similar to parts of the Dezadeash Group to the west. The meta-andesite is lithologically similar to parts of the

Figure 2. Flaggy, northeast-dipping grey muscovite schist on Jacquot Island (Kluane Lake). View looking northeast across Kluane Lake.

Figure 3. Northeast-dipping crenulation cleavage in muscovite schist, east shore of Kluane Lake.

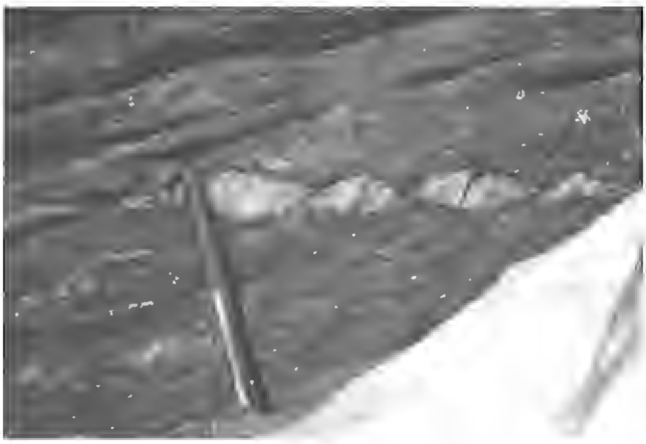

Figure 4. Extended quartz vein in biotite-garnet-cordierite-sillimanite schist near Fourth of July Creek.

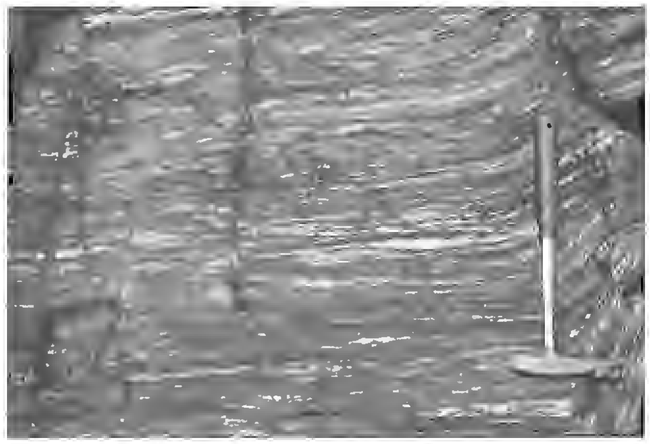

Figure 5. Muscovite schist showing strong quartz rodding, east shore of Kluane Lake.

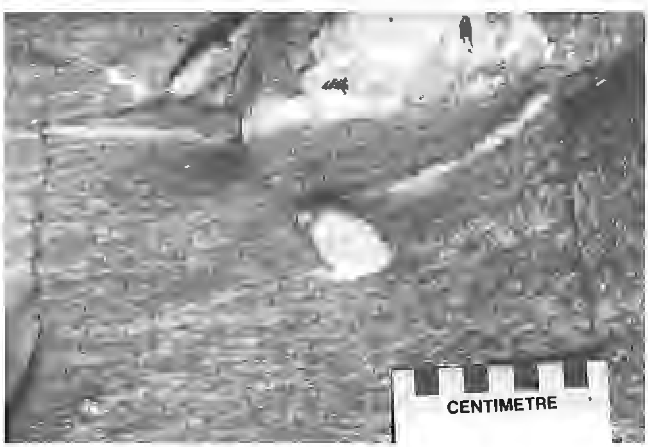

Figure 6. Rolled and attenuated quartz boudin in muscovite schist on the east shore of Kluane Lake.

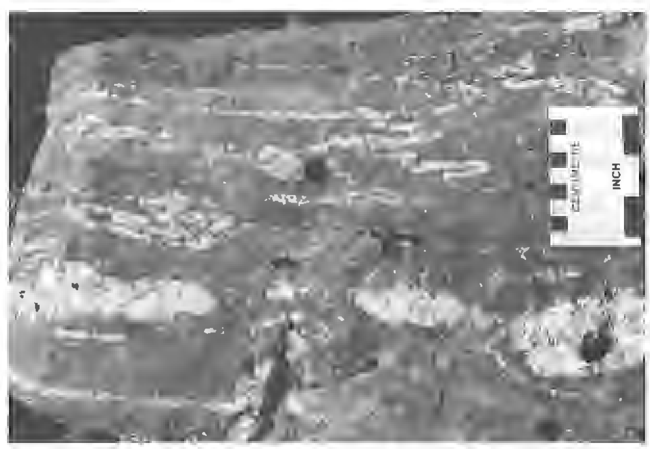

Figure 7. Sillimanite migmatite in biotite schist of the Nisling terrane near Undie Creek.

Triassic Lewes River Group and to other late Paleozoic volcanic strata of Laberge and Whitehorse map areas to the east, although no correlation is implied here. The meta-andesite and sedimentary strata are only slightly metamorphosed, in contrast to the commonly sillimanite-bearing Nisling rocks to the east. The amphibolite described to the southeast of Klukshu Lake by Kindle (1952) is interpreted as a contact- metamorphosed equivalent of the meta-andesite, at the margin of a granite of the Coast Plutonic Complex. It is distinct from the strongly foliated biotite-amphibolite layers of the Nisling terrane.

The band of schist reported by Kindle along Motherall Creek, 10 km south of Klukshu Lake, could not be found. Granite of the Coast Plutonic Complex underlies the area, and is juxtaposed along a steep, young fault against unmetamorphosed grey siltstone and greywacke of the Dezadeash Group to the west. New roadcuts resulting from highway development in 1989 show that the grey turbidites of the Dezadeash Group are not metamorphosed and preserve fine sedimentary structure within 50 m of the contact. Abundant subhorizontal slickensides and breccia zones are seen in the Dezadeash Group in the vicinity of the contact.

The interpretation by Kindle (1952) of the slate and marble assemblage exposed on the east bank of the Alsek River as a metamorphosed equivalent of the Dezadeash Group is invalidated by the location of these rocks (originally mapped as undivided Yukon Group) west of both the Duke River and Denali faults with respect to the Dezadeash Group.

FUTURE RESEARCH

These reconnaissance results show that schists of different protolith occur in the Kluane Schist, and that the metamorphic grade and structural style of parts of the Kluane Schist and the pre-Triassic Nisling terrane are similar. It is unlikely that most of the Kluane Schist is the metamorphosed equivalent of the Dezadeash Group. Its possible tectonic affinity with the Nisling terrane needs testing. The distribution, contact relations, protolith age, metamorphic age, and structural and P-T evolution of these rock assemblages are still unclear, and are the object of ongoing study.

ACKNOWLEDGMENTS

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Metamorphic rocks in the Florence Range, Coast Mountains, northwestern British Columbia

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Abstract

The Florence Range (NTS 104M/8) is mainly underlain by four fault-bounded lithological units of penetratively deformed and variably metamorphosed rocks of probable pre-Early Jurassic age. These are, from north to south: the Boundary Ranges metamorphic suite (interpreted to be oceanic), the foliated Hale Mountain granodiorite, Wann River gneiss (interpreted as metamorphosed volcanic rocks), and Florence Range metamorphic suite (interpreted as a metamorphosed continental margin assemblage). Contacts between these units are thought to be faults or shear zones. The lithological units, and structures within them, are unconformably overlain by volcanic rocks of probable Mesozoic age and are bounded by late Mesozoic plutons of the Coast Plutonic Complex on the west and the Llewellyn fault on the east.

Résumé

La majeure partie du sous-sol du chaînon Florence (SNRC 104M/8) est constituée de quatre unités lithologiques, limitées par des failles, de roches à déformation pénétrative et à degré de métamorphisme variable, d'âge probablement antérieur au Jurassique inférieur. Il s'agit, du nord au sud, des unités suivantes: la série métamorphique de Boundary Ranges (qui, serait de nature océanique), les granodiorites feuilletées de Hale Mountain, les gneiss de Wann River (qui seraient des roches volcaniques métamorphiques) et la série métamorphique de Florence Range (qui serait un ensemble de marge continentale métamorphisée). On pense que les contacts entre ces unités sont des failles ou des zones de cisaillement. Les unités lithologiques et les structures qu'elles renferment sont recouvertes en discordance par des roches volcaniques, probablement mésozoïques, et sont bordées par des plutons du Mésozoïque supérieur du complexe plutonique côtier à l'ouest, et par la faille de Llewellyn à l'est.

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INTRODUCTION

Fieldwork during 1989 focused on 1:20 000 scale mapping of the previously undivided metamorphic rocks of the Florence Range (59°15'-59°25'N, 134°10'-134°30'W; NTS 104M/8) of the Coast Mountains of northwestern British Columbia (Fig. 1). These rocks have been included in the Nisling Terrane, which is interpreted to be a displaced continental margin assemblage of unknown origin (Wheeler and McFeely, 1987). Nisling Terrane rocks resemble continental margin rocks of western North America (e.g. Windermere Supergroup). However, east of the Florence Range there are oceanic rocks of early Paleozoic to early Mesozoic age that separate the Nisling Terrane from North America. Either the Nisling Terrane is a rifted fragment of North America or a fragment of another continent. In addition the precise age of the pre-Late Triassic metamorphism of the terrane is uncertain. These are subjects that will be addressed by this study.

GENERAL GEOLOGY AND PREVIOUS WORK

The Florence Range lies between the southern ends of Tagish and Atlin lakes (Fig. 1, 2). Metamorphic rocks are bounded on the west by undeformed, probably late Mesozoic, granitic and granodioritic intrusions of the Coast Plutonic Complex. To the east, metamorphic rocks are separated by the Llewellyn fault from Upper Triassic volcanic rocks of the Stuhini Group (Stikine Terrane) and undeformed Mesozoic plutonic rocks (Christie, 1957; Werner, 1978; Bultman, 1979; Fig. 2).

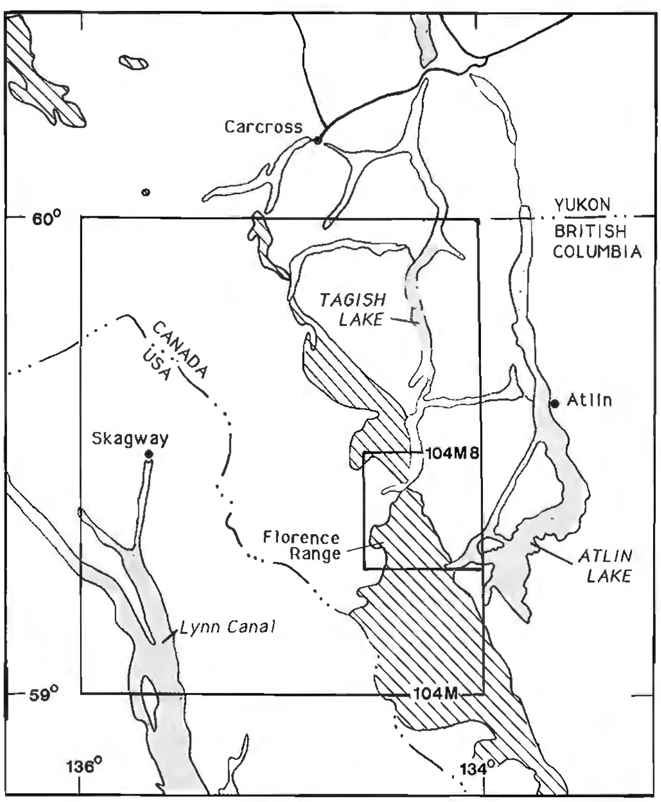

Figure 1. Location map of the Florence Range. Lined areas indicate metamorphic rocks of the Nisling Terrane (modified from Wheeler and McFeely, 1987).

Mapping at 1:250 000 by Christie (1957) outlined the regional extent of exposed metamorphic rocks in map area 104M. He divided the metamorphic rocks into units 1a (micaceous quartzite, hornblende-quartz-feldspar gneiss, amphibolite, schist, and limestone), and 1b (chlorite schist, feldspar-chlorite gneiss, amphibole gneiss, and limestone). Other than mapping the locations of large carbonate layers in the latter group of rocks, the two units were not subdivided.

More detailed mapping was conducted in the Florence Range by Bultman (1979), who mapped the eastern margin of the Florence Range at reconnaissance scale, and Werner (1977, 1978), who mapped the metamorphic rocks exposed south of the Wann River and south of Willison Creek at 1:30 000 scale (throughout this manuscript "the Wann River" refers to the Wann River above Nelson Lake). Neither study subdivided the metamorphic rocks, although Werner mapped carbonate bands and major axial surface traces (Werner, 1978).

Metamorphic rocks continuous with those of the Florence Range extend northwest to the B.C.-Yukon border (Fig. 1). They have been mapped at 1:50 000 scale as the Boundary Ranges metamorphic suite (Mihalynuk and Rouse, 1988a,b; Mihalynuk 1989; Mihalynuk et al., 1989). Lithologies include chlorite-actinolite schist, biotite-plagioclase-quartz schist, chlorite schist, graphitic schist, and minor marble, pyroxene-plagioclase schist, impure metaquartzite, and orthogneiss. The orthogneiss bodies include altered and deformed leucogranite and quartz diorite, Bighorn granite, and Hale Mountain hornblende-biotite granodiorite (Mihalynuk and Rouse, 1988b; Mihalynuk et al., 1989).

North of the B.C.-Yukon border metamorphic rocks are exposed as isolated pendants in Mesozoic plutons (Wheeler, 1961; Doherty and Hart, 1988). There they comprise biotite-muscovite-quartz-feldspar schist, chlorite-rich biotite-granite gneiss, quartzite and minor quartz-mica schist with rare amphibolite bands, and foliated hornblende and hornblende-biotite granodiorite, quartz diorite, and quartz monzonite (Doherty and Hart, 1988).

The nature of the basement to the metasedimentary rocks of map sheet 104M/8 on is uncertain. The ages of deposition, deformation, and metamorphism of the metamorphic rocks are inferred to be pre-Late Triassic, because clasts similar to metamorphic rocks of the Florence Range occur in the Upper Triassic Stuhini Group at the southern end of Atlin Lake (Bultman, 1979). The Boundary Ranges metamorphic suite is interpreted to have been deposited, deformed, and metamorphosed before Early Jurassic time, because sediments containing Early Jurassic fossils have been mapped as unconformably overlying the Boundary Ranges metamorphic suite north of the Florence Range (Mihalynuk and Rouse, 1988a,b).

LITHOLOGICAL SUBDIVISIONS

In the Florence Range, the metamorphic rocks can be divided into four units (Fig. 2): the Boundary Ranges metamorphic suite, Hale Mountain granodiorite, Wann River gneiss, and Florence Range metamorphic suite. The rocks

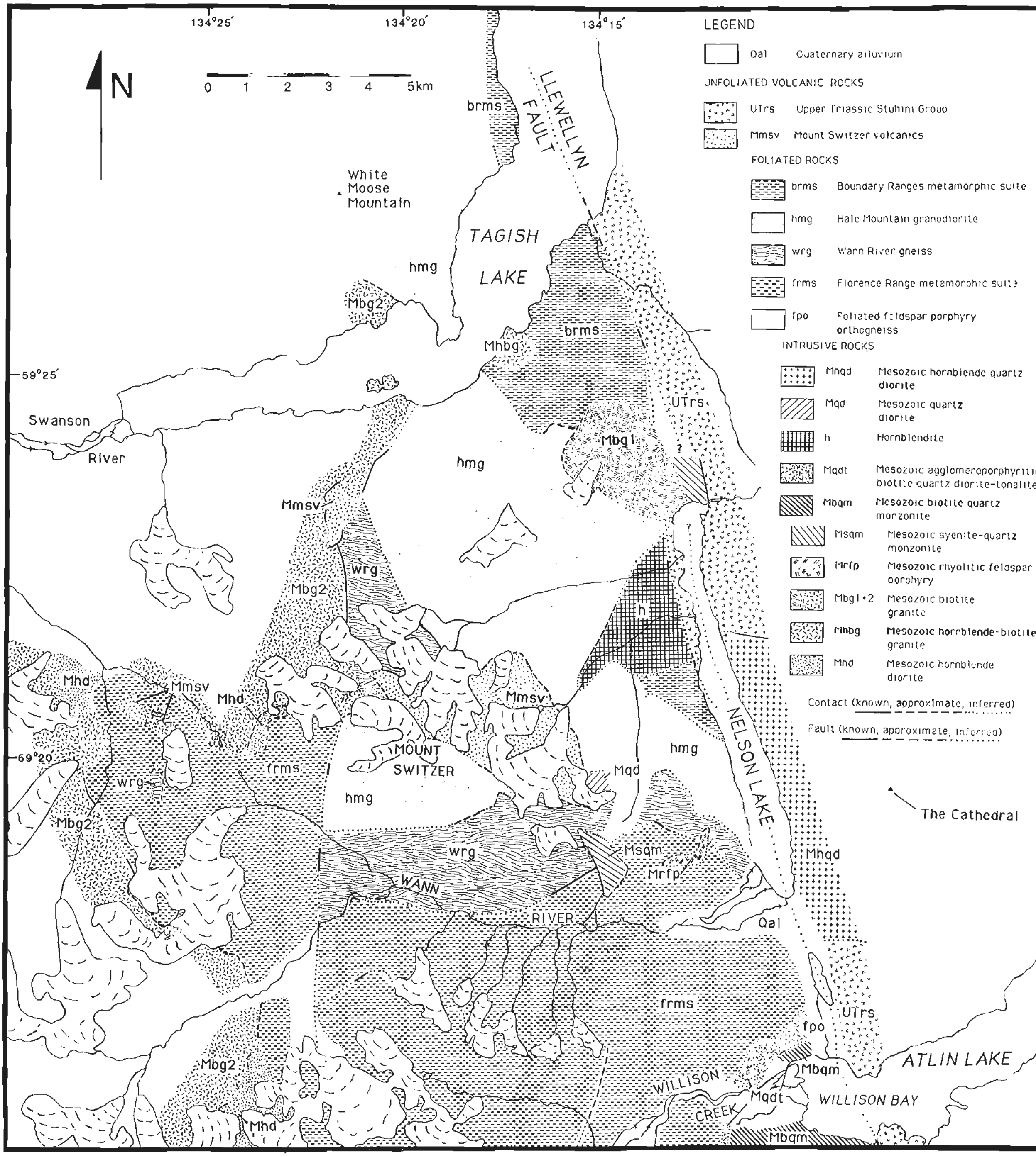

Figure 2. Geological map of the Florence Range of map area 104M/8. The locations of some contacts were generously provided by M.G. Mihalynuk of the British Columbia Geological Survey Branch, who mapped in the 104M/8 and 104M/9E map areas during the 1989 field season.

are cut by undeformed intrusive rocks and are overlain by undeformed volcanic rocks of probable Mesozoic age.

Boundary Ranges metamorphic suite

The Boundary Ranges metamorphic suite (Mihalynuk and Rouse, 1988a) is exposed along the western and southern shores of Tagish Lake. It is primarily composed of chlorite schist and chlorite-actinolite schist, with minor chlorite-pyroxene schist, discontinuous carbonate layers, quartzite, layer-parallel 1-20 cm thick felsic layers, and orthogneiss. The orthogneiss varies in composition from granite to diorite.

Possible protoliths for the chlorite-bearing schists include pelitic, calcareous marine sedimentary, volcanic, and reworked volcanic rocks. Felsic layers may be tuffs or felsic flows and the quartzite may be metamorphosed sandstone or chert. A possible tectonic environment for the deposition of these sediments is a volcanic arc setting. If the protolith for the quartzite is a sandstone, then the arc may have formed near or on continental crust.

The contact between the Boundary Ranges metamorphic suite and the structurally overlying Hale Mountain granodiorite is characterized by interlayering, on a decimetre-scale, of both units. This layering may reflect shearing along the contact, and therefore this contact is tentatively interpreted to be a shear zone.

Hale Mountain (hornblende-biotite) granodiorite

The Hale Mountain hornblende-biotite granodiorite (Mihalynuk, 1989; Mihalynuk et al., 1989) is exposed north of the Wann River in the Florence Range, and on White Moose and Hale mountains to the north. It is characterized by medium- grained plagioclase phenocrysts in a fine-grained groundmass of hornblende with minor biotite, chlorite, and epidote. Metre-scale compositional layering is ubiquitous. More felsic layers (0.5-3 m thick) are coarse grained and occur only in the structurally highest exposures.

In some areas the granodiorite has a strong foliation and lineation, but in other areas fabrics have not been observed. However, phenocrysts are recognizable throughout this unit. Undeformed potassium feldspar-bearing pegmatites characteristically crosscut the Hale Mountain granodiorite.

Wann River gneiss

The Wann River gneiss is exposed in the Wann River valley (above Nelson Lake), structurally above the Hale Mountain granodiorite and below the Florence Range metamorphic suite (Fig. 2), and as small occurrences in thrust slices of the Florence Range metamorphic suite. It exhibits a distinctive millimetre- to decimetre-scale compositional layering (Fig. 3). The composition varies from dioritic (20 % hornblende) to gabbroic (50 % hornblende), with minor biotite and epidote. The small scale and the gradational nature of compositional layering in the gneiss suggests that layering is a primary fabric. Therefore the gneisses are interpreted to be metavolcanic rocks of broadly intermediate composition. Unlike the Hale Mountain granodiorite, the Wann River gneiss does not contain plagioclase phenocrysts; felsic lithologies are medium- grained, compositional layering occurs on a smaller scale, and the gneisses are everywhere strongly foliated and crosscut by plagioclase-bearing pegmatites.

The contact zone between the Hale Mountain granodiorite and the Wann River gneiss is either abrupt, or an interlayered boundary that is continuous over at least 100 m, or a brecciated and pegmatite-flooded zone about 200 m wide (M.G. Mihalynuk, pers. comm., 1989). The foliation is more strongly developed in both units toward the contact, indicating that the contact is sheared. On the ridge west of Nelson Lake a 20 m- thick succession of metasedimentary rocks are preserved at the contact between the Hale Mountain granodiorite and the Wann River gneiss.

Figure 3. Well- layered hornblende gneiss of dioritic composition from the Wann River gneiss unit.

Florence Range metamorphic suite

The Florence Range metamorphic suite outcrops on the slopes north of the Wann River and continues to the south, beyond the southern edge of map area 104M/8 (Souther, 1971; Werner, 1977, 1978; Fig. 1, 2). It structurally overlies the Wann River gneiss. The contact, where observed, is marked by an abrupt lithological change in a zone of strained, well- layered gneiss. This zone is tentatively interpreted to be a shear zone.

The Florence Range metamorphic suite is composed of semipelitic, pelitic, carbonate, amphibolitic, and calc-silicate rocks, with minor quartzite and graphite-bearing pelitic and semipelitic rocks. Semipelitic and pelitic layers (biotite, quartz, plagioclase, muscovite, garnet, kyanite, sillimanite, graphite) are 0.1-30 m thick. Quartzite layers are commonly impure ($\pm$biotite) and reach 3 m in thickness. Amphibolite ($\pm$garnet) is associated with carbonate rocks. Carbonate and amphibolite are interlayered on a scale of 0.1-20 m and form layers that are continuous for up to several kilometres along strike. Such layers are thicker and increasingly abundant toward the west, whereas semipelitic layers become more common toward the east. Calc-silicate rocks are composed of calcite with varying amounts of tremolite, diopside, actinolite, grossular garnet, and anorthite, and form layers 0.02-1 m in thickness. A feldspar-porphyry orthogneiss was found north of the mouth of Willison Creek.

The rock types in Florence Range metamorphic suite suggest a continental margin setting. Amphibolites are interpreted to be metamorphosed flows, tuffs, and reworked tuffs.

Mount Switzer volcanics

The Florence Range metamorphic suite is unconformably overlain by the undeformed Mount Switzer volcanics, which are preserved on Mount Switzer. Clasts of granite, biotite schist, foliated amphibolite, and quartzite are preserved in the conglomerate overlying the unconformity.

Plutonic rocks

Two hornblendite bodies, one of which is too small to map at the scale of Figure 2, were found in the Florence Range. Hornblendite also occurs in map area 104M/10 (Mihalynuk, 1989). Although undeformed, they are not necessarily entirely posttectonic, because the absence of foliation within the hornblendite may reflect the extreme competence of this lithology. South of the Wann River, garnet-bearing hornblendite crosscuts layering and foliation within the Florence Range metamorphic suite, indicating that the hornblendite must have been intruded after the first phase of deformation that affected the Florence Range metamorphic suite, but before the Florence Range metamorphic suite was metamorphosed. North of the Wann River, foliation in the Hale Mountain granodiorite wraps around xenoliths of hornblendite.

Cathedral Mountain is primarily underlain by hornblende quartz diorite to granodiorite. The biotite quartz monzonite that outcrops north and south of Willison Creek is not part of the Cathedral Mountain batholith, as was previously reported by Bultman (1979). The quartz monzonite does, however, intrude the variably deformed, medium- grained agglomeroporphyritic biotite quartz diorite-tonalite that outcrops north and south of the mouth of Willison Creek. The absence of a penetrative foliation indicates that this intrusion postdates deformation that produced a penetrative fabric in the Florence Ranges metamorphic suite. The fabrics that have been developed may be related to later faulting.

The syenite to quartz monzonite unit intrudes the contact between the Wann River gneiss and the Florence Range metamorphic suite west of the southern end of Nelson Lake. It is medium- grained and locally trachytic. Also exposed west of the southern end of Nelson Lake is rhyolitic feldspar porphyry that intrudes the Florence Range metamorphic suite and the Wann River gneiss.

Medium- to coarse-grained biotite granite outcrops on the northern shore of Nelson Lake and on the ridge to the northwest. The intrusive contact between the biotite granite and the Florence Range metamorphic suite was examined on the ridge, but no contacts have been found in the valley below. Where the granite is exposed on Nelson Lake, it lies on the eastern side of a splay of the Llewellyn fault and is therefore thought to intrude the Llewellyn fault in part. It may also intrude the Upper Triassic Stuhini Group, which is exposed to the east.

Hornblende-biotite granite exposed on the southern shore of Tagish Lake has not been shawn to be the same as, or continuous with the granite that outcrops north of Nelson Lake, described above. No contacts with surrounding country rock have been observed.

The hornblende diorite (with minor biotite) that outcrops in the southwestern quarter of map area 104M/8 is cut by granites that are exposed to the west and intrudes the Florence Range metamorphic suite. It commonly contains inclusions of the Florence Range metamorphic suite.

Medium- to coarse-grained biotite granites of the Coast Plutonic Complex intrude the western margin of the Florence Range metamorphic suite, the Wann River gneiss, the Hale Mountain granodiorite, the Boundary Ranges metamorphic complex, and the hornblende diorite.

STRUCTURE

Folds

Mesoscopic folds have been observed only in the strongly layered Boundary Ranges and Florence Range metamorphic suites. The absence of mesoscopic folds in the Hale Mountain granodiorite and the Wann River gneiss may reflect the limited range in competence within these units. However, they are folded by large-scale open folds.

The Boundary Ranges metamorphic suite has experienced at least three phases of folding (Mihalynuk et al., 1989). A pre-existing metamorphic fabric is commonly isoclinally folded by mesoscopic folds, and carbonate layers outline refolded folds. Areas of planar layering are interpreted to be located on the limbs of large-scale folds. However, the lack of continuous distinctive rock- types, such as carbonate layers, makes the mapping of large-scale structures in this suite difficult.

Within the Florence Range metamorphic suite a pre-existing layer-parallel metamorphic foliation is folded by folds, commonly northeast-verging, that range in size from crenulations (0.5-3 cm) to megascopic folds (Fig. 4). Variations in the vergence of these folds are attributed to later faulting or folding, some examples of which have been seen in outcrop.

Faults and shear zones

The contacts between the Boundary Ranges metamorphic suite, the Hale Mountain granodiorite, the Wann River gneiss and the Florence Range metamorphic suite are interpreted to be faults or shear zones. The contacts are offset by later, steeply- dipping, approximately north-south striking faults (Fig. 2, 5), and truncated by the Llewellyn fault.

Figure 4. Synform outlined by carbonate from southwestern corner of the study area. Note that schist layers are more competent than the surrounding carbonate and therefore deform by faulting as well as folding.

Faults within the Florence Range metamorphic suite are interpreted as northeast-verging thrust faults (Fig. 2, 6) that place more carbonate-rich thrust sheets over thrust sheets dominated by semipelitic rocks. At the base of some of the thrust sheets, the Wann River gneiss occurs. The initial relationship between the Florence Range metamorphic suite and the Wann River gneiss is not known.

The contacts between the Wann River gneiss and Hale Mountain granodiorite, and the Hale Mountain granodiorite and Boundary Ranges metamorphic suite are characterized by strongly developed ductile fabrics. They are in part thought to be shear zones.

Steep, roughly north-south striking faults are recognized in the southern half of the field area where a variety of lithologies are present (Fig. 2, 5), and in the southwest, where different lithological units are juxtaposed (Fig. 2).

The Llewellyn fault (Bultman, 1979) truncates all penetrative fabrics and structures in the metamorphic rocks of the Florence Range. The sense and amount of displacement on the Llewellyn fault is uncertain. The juxtaposition of metamorphic rocks on the west and unmetamorphosed rocks on the east is indicative of east-side-down relative movement and/or strike-slip movement.

Figure 5. View toward the northeast of carbonate layers in the Florence Range metamorphic suite that outline northeast-verging folds that are truncated below by a northeast-verging thrust fault.

Figure 6. Carbonate layer offset by late, steeply dipping fault. View north of ridge north of Willison Creek.

METAMORPHISM

The metamorphic grade varies within the Florence Range from greenschist to transitional greenschist-amphibolite facies in the Boundary Ranges metamorphic suite, to upper amphibolite facies (sillimanite/fibrolite) in the Florence Range metamorphic suite. Garnet occurs sporadically in schist and amphibolite, and kyanite (with muscovite alteration) rarely occurs in schist. Garnets are commonly chloritized.

CONCLUSION

The previously undivided metamorphic rocks of the Florence Range are grouped into four fault-bounded, lithologically distinct subdivisions: the Boundary Ranges metamorphic suite, Hale Mountain granodiorite, Wann River gneiss, and the Florence Range metamorphic suite.

Rock types that make up the Boundary Ranges metamorphic suite in the Florence Range correspond to Christie's (1957) unit 1b and are continuous with a northwest-trending belt of metamorphic rocks that extend as far north as the B.C.-Yukon border (Mihalynuk and Rouse, 1988a; Mihalynuk et al., 1989). Protoliths for these metamorphic rocks may have formed in a volcanic arc setting with possible continental influence. They lack abundant quartzose and carbonate rocks that are typical of the Nisling Terrane. However, the Florence Range metamorphic suite comprises rock types typical of a continental margin setting. The suite closely resembles Christie's (1957) unit 1a and metamorphic rocks of the Nisling Terrane exposed north of the British Columbia-Yukon border (Wheeler, 1961; Wheeler and McFeely, 1987; Doherty and Hart, 1988, rather than the Boundary Ranges metamorphic suite.

The relationship between Florence Range and Boundary Ranges metamorphic suites is unclear, as they are separated by the Hale Mountain granodiorite and the Wann River gneiss. The Boundary Ranges metamorphic suite may be a distal equivalent of the Florence Range metamorphic suite, it may have some affinity with the Stikine Terrane, or it could be allochthonous to both.

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Radiolarian age determinations from the Canadian Cordillera

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Cordey, F., Radiolarian age determinations from the Canadian Cordillera; in Current Research, Part E, Geological Survey of Canada, Paper 90-1E, p. 121-126, 1990.

Abstract

This report summarizes previous results and gives brief descriptions of units sampled during the summer of 1989. 250 samples of chert, tuff, argillite, sandstone, and conglomerate were collected from the Pacific Rim, Wrangell, Bridge River, Stikine, and Cache Creek (southern, central, and northern) terranes. All samples collected contain radiolarians and, if they yield identifiable faunas, will give new biostratigraphic data with possible structural implications.

Résumé

Le présent rapport résume des résultats antérieurs et décrit sommairement des unités échantillonnées au cours de l'été de 1989. On a prélevé 250 échantillons de chert, de tuf, d'argilite, de grès et de conglomérat provenant des terranes de Pacific Rim, de Wrangell, de Bridge River, de Stikine et de Cache Creek (sud, centre et nord). Tous les échantillons prélevés renferment des radiolaires et, s'ils contiennent des faunes identifiables, donneront de nouveaux âges biostratigraphiques accompagnés de conséquences structurales possibles.

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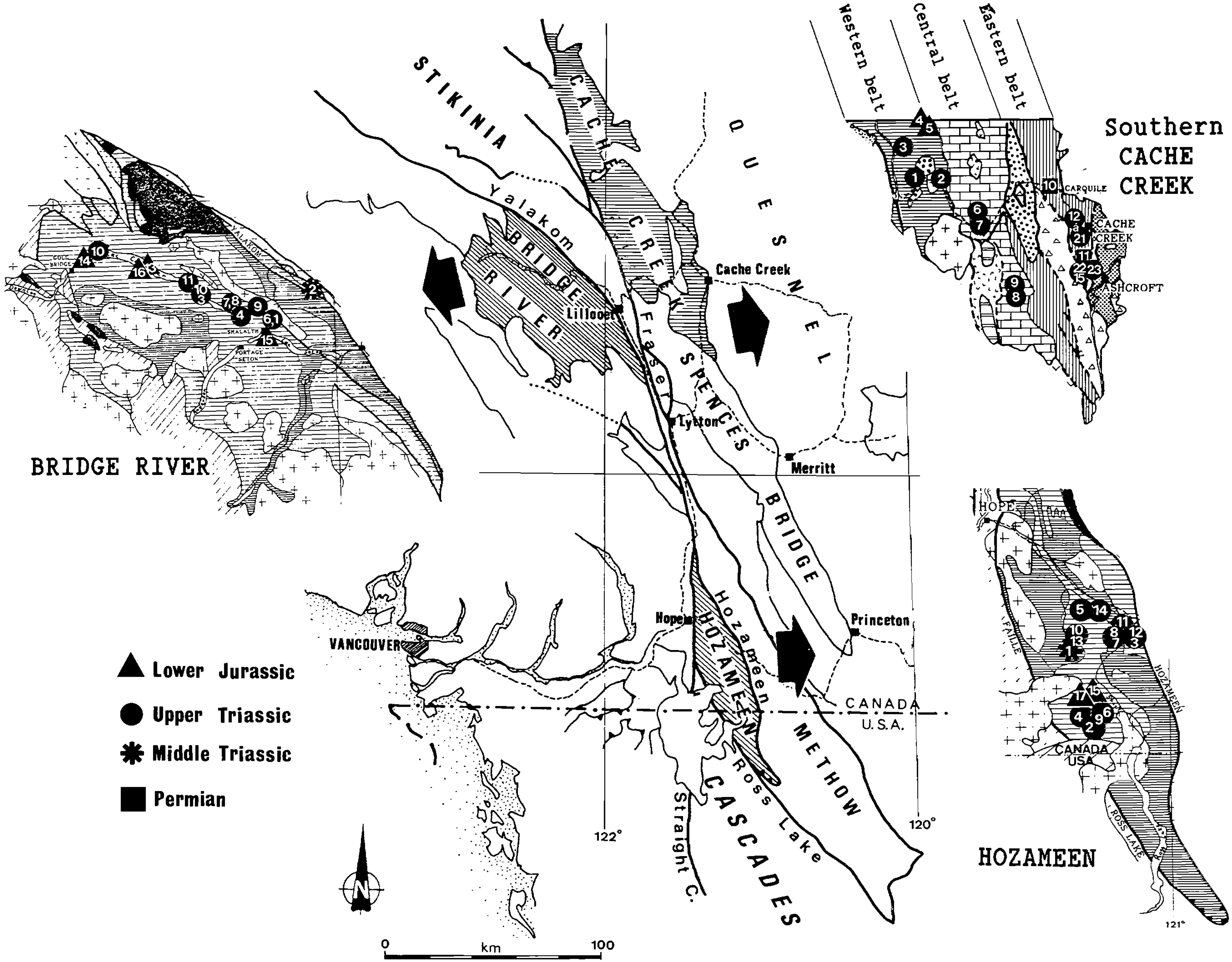

Figure 1. Synthesis of radiolarian dates obtained by the author in the Cache Creek, Bridge River, and Hozameen complexes in southern British Columbia (after Cordey, 1988).

INTRODUCTION

Previous work has shown the importance of paleontological dates in understanding the geology of the Canadian Cordillera. Since the 1970s, radiolarians have become a valuable tool, especially in rock units where siliceous lithologies are widespread. After a study of radiolarians in ophiolitic complexes of southern British Columbia (Cordey, 1988), it was clear that these complexes, and other units, could be profitably studied elsewhere in the Canadian Cordillera. Fieldwork, supported by the GSC, was done during the summer of 1989 in co-operation with several mapping teams of both the GSC and the British Columbia Geological Survey (BCGS). Geological units of southern, central, and northwestern British Columbia, and southern Yukon containing radiolarian faunas were sampled. This report gives a summary of results obtained to date.

CACHE CREEK COMPLEX

Radiolarians were identified from 23 localities in southern British Columbia (Cordey, 1988; Fig. 1). Early Permian to Late Triassic ages from cherts of the eastern and central belts of the complex are consistent with previous studies (e.g. Orchard, 1981, 1984; Monger and McMillan, 1984). In the western belt of the Cache Creek Complex, tuffaceous argillites yielded Early or Middle Jurassic radiolarians (Cordey et al., 1987), suggesting that the Cache Creek basin in southern British Columbia may have closed later than previously thought. Cherts in the western belt were resampled during 1989, particularly in the Pavilion-Kelly Creek area, near the Jurassic localities of Cordey et al. (1987).

The Cache Creek Complex extends for more than 1000 km along the length of the Canadian Cordillera. The minimum age of the Cache Creek in central and northern British Columbia is unknown and has important tectonic implications. In 1989 fieldwork was extended to selected areas in central and north British Columbia (Fig. 2) to attempt to answer this question. Near Fort St. James, along Stuart Lake, exposures of limestone and chert have been dated with conodonts by M.J. Orchard (pers. comm., 1989) as Pennsylvanian and Norian. Several new samples were collected from a cliff exposure where a section of grey ribbon chert contains radiolarians.

In co-operation with H. Gabrielse, a section of limestone overlain by chert interbedded with greywacke and sandstone at the head of Canyon Creek, northwest of Dease Lake, was sampled. This section lies in the northeastern facies belt of the "Atlin terrane" of Monger (1975).

A chert sample from this section was processed in the base camp using the methods in Cordey and Krauss (1990) and yielded Mesozoic, probably Triassic, radiolarians.

In co-operation with M. Bloodgood (BCGS), exposures of grey and red radiolarian-bearing cherts were sampled on the ridge of Sentinel Mountain, south of Atlin, where Permian and Pennsylvanian radiolarians were collected by J.W.H. Monger and identified by D.L. Jones (pers. comm. to Monger, 1975).

In co-operation with J. Jackson (University of Arizona), sections have been studied in the central facies belt of the Atlin terrane, between Atlin and Teslin lakes in the southern Yukon. On a ridge east of Teenah Lake, thick sections of grey radiolarian cherts interbedded with siliceous argillites and sandstones are exposed. A sample of chert collected by J. Jackson on this ridge yielded "Rhaetian" conodonts (M.J. Orchard, pers. comm., 1989). This locality extends the youngest age of the Cache Creek strata in northern British Columbia and provides the youngest age obtained on pure chert in the entire Cache Creek Complex. On the ridge situated west of Spawn Lake, a similar succession is observed, and numerous radiolarian-bearing cherts were collected. One sample was processed in the base camp and yielded Late Triassic radiolarians (including *Capnuchosphaera* sp.). Contiguous exposures contain radiolarians which could be of different ages.

BRIDGE RIVER COMPLEX

Middle Triassic to Early Jurassic ages on radiolarians were reported from 16 localities from cherts in the Bridge River Complex by Cordey (1986, 1988) (Fig. 1). This range is consistent with previous work (Cameron and Monger, 1971; Monger, 1977; Potter, 1983). Cherts of the Hozameen complex (Fig. 1) gave similar results in terms of both age and in proportions (Fig. 3): Triassic represents about 75 % of the localities and Jurassic 25 %. This study supports the correlation formulated previously between the two complexes (Cameron and Monger, 1971; Monger, 1977, 1985, 1986; Potter, 1983).

Fieldwork in 1989 is summarized in Figure 2. In co-operation with P. Schiarizza (BCGS) and his team working in Tyaughton Creek map area (Schiarizza et al., 1989), the author visited and collected several undated exposures of the complex. Previous data came from exposures along the main roads along or near Carpenter Lake. We reached exposures in the areas of Gun Lake and Noaxe and North Cinnabar creeks, i.e. in the northwestern part of the complex. Every sample collected contains visible radiolarians, some of them well enough preserved to reveal internal structures.

Cherts in the Bridge River Complex are generally sequences of grey ribbon chert that are similar to the grey cherts of Hozameen complex and the western and eastern belts of the Cache Creek Group. It appears that several different types of chert are present in the Bridge River Complex, including green, red, and brown lithologies; some show evidence of tuffaceous sedimentation.

STIKINE ASSEMBLAGE

In co-operation with R.G. Anderson (GSC), D. Brown (BCGS), and M. Gunning, the author examined several units of the Stikine assemblage in the Ambition Mountain massif located a few kilometres east of Stikine River in the Coast Mountains.

Near the Scud Glacier toe, a sequence of red massive chert overlies a black chert member and pyroxene-bearing wackes. Permian conodonts have been identified by M.J. Orchard in the red chert member. Radiolarians were collected from all three units. Radiolarians are also present in

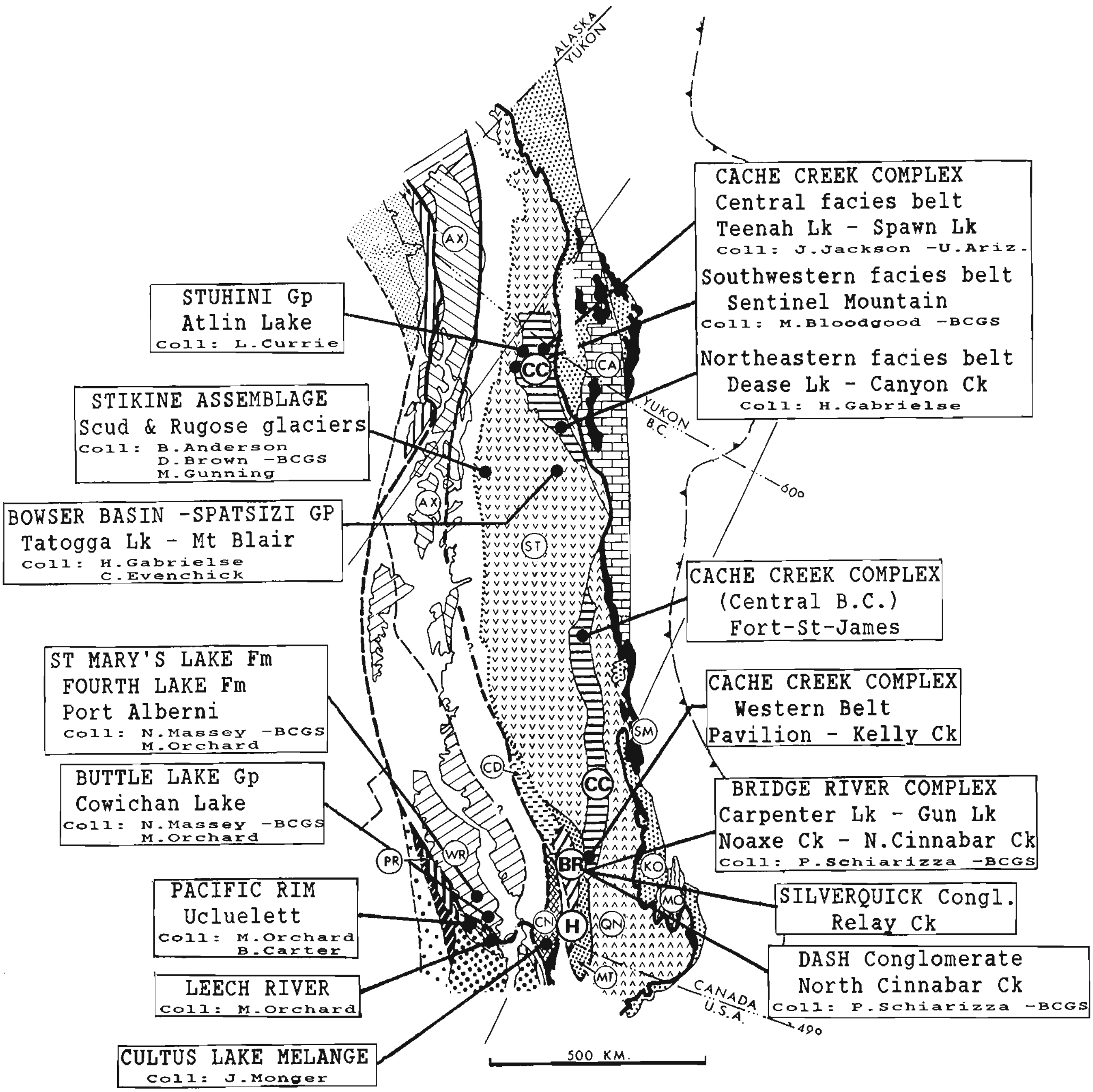

Figure 2. Fieldwork 1989 - synthesis of geological units. Terrane map after Price et al. (1985). In each frame is mentioned geological unit, location, and collaborators.
Terranes: PR: Pacific Rim; WR: Wrangellia; AX: Alexander; CN: Chilliwack-Nooksask; CD: Cadwallader; BR: Bridge River; H: Hozameen; MT: Methow; ST: Stikinia; CC: Cache Creek; QN: Quesnellia; SM: Slide Mountain; KO: Kootenay; MO: Monashee.

a sequence of green siliceous siltstone near the toe of the Rugose Glacier.

VANCOUVER ISLAND

Fieldwork was undertaken with M.J. Orchard and in co-operation with N. Massey (BCGS) on the Paleozoic Buttle Lake Group. The Paleozoic Sicker Group was revised by Muller (1980) who proposed four subdivisions (Nitinat, Myra, sediment-sill unit, Buttle Lake). Work in progress by Massey, replaces the Sicker Group with the new Buttle Lake Group, subdivided, in ascending stratigraphic order, into Fourth Lake Formation (including the Shaw Creek Member), Mount Mark Formation, and Saint Mary's Lake Formation.

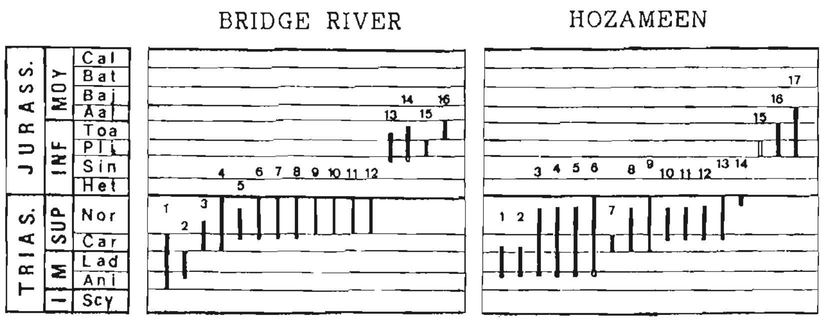

Figure 3. Comparison of ages from radiolarian associations in chert samples of Bridge River and Hozameen complexes. The numbers indicate the localities from older to younger in each complex (Cordey, 1988).

The Shaw Creek Member was formerly included in the now-abandoned "Cameron River Formation" (Massey and Friday, 1989). It is interpreted as the base of the Buttle Lake Group and has been previously dated as Early Mississippian on both conodonts and radiolarians. The type section, near Cowichan Lake, is a remarkable section of thinly laminated green cherty tuffs. Radiolarians and conodonts were observed on the outcrop, and many samples were collected.

Near Cameron River, the basal conglomerate of the Fourth Lake Formation contains abundant chert- pebbles with radiolarians. The conglomerate is older than mid-Pennsylvanian, the oldest dated Mount Mark limestone, but the pebbles are of unknown origin.

A section of grey to green to pink ribbon chert of the Leech River Group is exposed on the western side of Saanich Inlet along Duncan-Victoria Highway. This section is associated with pillow- basalts of unknown age. One sample was collected although radiolarians were sparse and poorly preserved.

A few pebbles of red chert were collected on a beach near the town of Ucluelet. Although they were not found in place, they contain exceptionally well preserved radiolarians whose ages are probably Late Jurassic or Cretaceous. These pebbles almost certainly were derived from outcrops of the Pacific Rim Complex.

CHERT-PEBBLE CONGLOMERATES

Chert-pebble conglomerates are scattered throughout the Intermontane Belt in southern British Columbia. Some of these strata are known to be mid-Cretaceous to Late Cretaceous in age. Several of the chert pebbles have yielded ages ranging from Late Devonian to Early Jurassic (Cordey, 1986, 1988). These studies showed that clasts from conglomerates may give ages that do not presently exist in the strata presently exposed in the unit. For example, a Late Devonian age was determined for a pebble in a conglomerate located near Lytton, apparently associated with either the Bridge River Complex or with the Cache Creek Complex. However, those units are not known to be older than Permian and Carboniferous respectively. Chert-pebble conglomerates can thus yield important data and should therefore be collected carefully and systematically.

Chert-pebble conglomerate, thought to have been derived from the Cache Creek Complex, is widespread in the Bowser Basin of north-central British Columbia. Currie (1984) reported Triassic pebbles (radiolarians identified by E.S. Carter) from the Ashman Formation conglomerate in the northeastern Bowser Basin. Three exposures of conglomerate from the Bowser Lake Group were sampled in the Mount Blair area (east of Tatogga Lake), with H. Gabrielse and C.A. Evenchick.

A pebble of green cherty tuff was collected by L. Currie (Carleton Geoscience Centre) from the Stuhini Group, at the southern end of Atlin Lake in northern British Columbia. The age of the conglomerate is Late Triassic but the provenance of the pebbles is unknown.

In the Bridge River region, chert pebbles from conglomerates associated with the Bridge River Complex were collected with P. Schiarizza (BCGS). Chert pebbles from the Dash conglomerate in northern Cinnabar Creek were also collected. Chert pebbles from the Silverquick conglomerate in Relay Creek were processed in the field and produced Late Triassic and Jurassic radiolarians respectively.

CONCLUSIONS

250 samples were collected during the summer of 1989. The quality of radiolarian faunas should be variable because of differences in depositional facies, tectonic history, and metamorphism. Previous investigations showed that careful collecting and processing generally allows the recovery of identifiable radiolarians. Conodonts are occasionally associated with radiolarians in cherts and can help improve the precision of the dating (Orchard, 1986).

Some faunas should give us information about units of unknown or uncertain age. Some data should help to develop and test structural hypotheses related to mapping projects. Chert- pebbles from conglomerates might give ages that will improve our knowledge of the ages of the conglomerate or show the presence of other geological units since eroded. Finally, the data may aid in testing and evaluating hypotheses about provincialism and paleolatitudinal differentiation of radiolarians.

ACKNOWLEDGMENTS

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Brown, B. Carter, L. Currie, C.A. Evenchick, B. Gaba, H. Gabrielse, M. Gunning, S. Irwin, J. Jackson, G. Jordan, M. Journeay, P.T. Krauss, R. MacDonald, N.W.D. Massey, M. Mihalynuk, and P. Schiarizza.

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A field technique for identifying and dating radiolaria applied to British Columbia and Yukon

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A field technique for identifying and dating radiolaria applied to British Columbia and Yukon; in Current Research, Part E, Geological Survey of Canada, Paper 90-1E, p. 127-129, 1990.

Abstract

With a mobile laboratory, the radiolarian specialist is able to process samples and identify radiolarian fauna in the field. In summer 1989 a few age dates were obtained in base camps in the Bridge River and Cache Creek complexes in southern and northwestern British Columbia. This technique can be applied concurrently with field mapping programs to provide a quick understanding of the stratigraphy and structure of geological units, and eventually influence mapping schedules. This approach is particularly useful in complexes with large amounts of siliceous, argillaceous and/or tuffaceous sedimentary strata.

Résumé

À l'aide d'un laboratoire mobile, le spécialiste en radiolaires peut analyser des échantillons et déterminer une faune de radiolaires sur le terrain. Au cours de l'été de 1989, on a obtenu quelques âges dans des camps de base, dans les complexes de Bridge River et de Cache Creek, dans le sud et le nord-ouest de la Colombie-Britannique. Cette technique peut être appliquée conjointement avec des programmes de cartographie sur le terrain de façon à permettre une compréhension plus rapide de la stratigraphie et de la structure des unités géologiques et, éventuellement, peut influer sur les programmes de cartographie. Elle est particulièrement utile dans des complexes renfermant de nombreuses couches sédimentaires siliceuses, argileuses ou tufacées.

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INTRODUCTION

Techniques for extracting microfossils from siliceous sedimentary rocks appeared in the seventies (Hayashi, 1969; Dumitrica, 1970; Pessagno and Newport, 1972) and are now commonly used by specialists. All methods used hydrofluoric acid, with variations in etching time and concentration of solutions. Processing of siliceous rocks generally requires a complete chemical laboratory with a fume hood and other equipment. For example, residue quality is enhanced by the use of cleaners baths (H_2O_2 or NaOH, sometimes warm), and the processing of argillaceous rocks requires alternation of acid concentrations and ultrasonics (Cordey, 1988). Also, radiolarian identification requires scanning electron microscopy when their preservation is poor. It appears nevertheless that processing of cherts, radiolarian identification and dating can be successfully accomplished in the field.

DESCRIPTION OF THE TECHNIQUE

Table 1 summarizes the different stages of field processing. Main precautions lie in hydrofluoric acid transportation, which can be pre-diluted to minimize risk of damage in the event of leakage. In the field, a few samples are processed simultaneously in small plastic beakers with lids and placed in an open, isolated area. An etching concentration of 2 to 4 % HF is sufficient, with duration varying from 4-24 hours between two residue recoveries. Lids and low concentration avoid acid fumes. After processing, the acid must be neutralized and stored in plastic tanks, prior to proper disposal in a chemical laboratory. Finally, the examination of residues requires the use of a binocular microscope (magnification x80), supplemented by an artificial light source in the absence of adequate daylight.

Table 1. Diagram summarizing radiolarian study stages in the field.

OUTCROP	Collection	Selection of good samples lithologies: cherts, tuffs, argillites, greywackes, sandstones, conglomerates
	Processing	HF 2 to 4% during 4 to 24 hrs. each sample or pebble must be processed separately
BASE CAMP or FLY CAMP	Residue recovery	selected sieves (12.80.200)
	Residue examination	selection of good specimen a good binocular microscope is necessary
	Identification and dating of radiolarians	
	Dating the sample	common age range of radiolarians

RESULTS OBTAINED

This method was used for the first time during 1989 summer fieldwork, when the senior author collected extensively for radiolarians in the Canadian Cordillera. Although it has not been attempted systematically for every geological unit studied, success was obtained for samples from the Bridge River complex in southern British Columbia, and in the Cache Creek complex in northwestern British Columbia.

Bridge River complex

In co-operation with P. Schiarizza (B.C. Geological Survey) and his team working in Tyaughton map area (Schiarizza et al., 1989), the laboratory was set up in the base camp located at the Gun Creek Ranch. A few pebbles collected 24 hours earlier from the Silverquick conglomerate at Relay Creek were processed. Radiolarian fauna was extracted and identified in two pebbles. These were dated respectively as Late Triassic (based on **Capnodoce** sp.), and probably Jurassic (on associations of nassellarians). These ages are consistent with those obtained from chert of the Bridge River complex (Middle Triassic to Lower Jurassic), from which the pebbles of the Silverquick conglomerate are thought to have originated.

Cache Creek complex

In co-operation with M. Bloodgood (B.C. Geological Survey) and her mapping team working out of Atlin, we processed several samples collected in the northwestern part of the Cache Creek complex in British Columbia and Yukon. One sample of grey chert collected with J. Jackson (University of Arizona, Tucson) in the central facies belt of the complex yielded Upper Triassic radiolarians (including **Capnuchosphaera** sp.). A second sample of grey chert collected in the northeastern facies belt with H. Gabrielse yielded Mesozoic (probably Triassic) radiolarians (on nassellarian forms).

SUMMARY

The radiolarian processing technique can be used in the future on a larger scale. It seems perfectly adapted to exhaustive study of ages of ophiolitic or accretionary complexes which possess large amounts of radiolarian-bearing siliceous sedimentary rocks (ribbon cherts, tuffs, argillites, sandstones, conglomerates). Age ranges of these complexes tend to be better established as biostratigraphic studies progress. Nevertheless, large areas remain undated and poorly correlated. Radiolarian-bearing sedimentary rocks processed in the field can possibly help to: 1) better understand the stratigraphy in poorly-documented areas or geological units; 2) underline presence and/or structure of tectonic slices in accretionary complexes; and, 3) orientate mapping schedules and priorities without delay.

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Mesozoic stratigraphy and setting for some mineral deposits in Iskut River map area, northwestern British Columbia

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Anderson, R.G. and Thorkelson, D.J., Mesozoic stratigraphy and setting for some mineral deposits in Iskut River map area, northwestern British Columbia; in Current Research, Part E, Geological Survey of Canada, Paper 90-1F, p. 131-139, 1990.

Abstract

In Iskut River map area, bimodal or intermediate to mafic Upper Triassic Stuhini Group volcanic rocks change to basinal dun feldspathic greywacke and siltstone to the east and northeast. Upper Triassic Stuhini Group grades into Lower Jurassic Hazelton Group near Treaty Creek.

The Hazelton Group comprises three heterogenous volcanogenic formations. Unuk River Formation, composed of andesitic breccia, tuff and marine siliceous siltstone is the oldest. Heterogeneous maroon to green volcanic conglomerate, breccia and greywacke of the Betty Creek Formation overlie it. The youngest is Mount Dilworth formation felsic tuff and tuff breccia, an important regional marker representing the climactic volcanic event of Hazelton volcanism.

Lower Middle Jurassic Salmon River Formation (of Bajocian ? age) includes: eastern siliceous shale and tuff turbidite; medial felsic and mafic pillowed lava, shale and limestone that hosts the Eskay Creek deposit; and a speculative western facies of andesitic, calc-alkaline volcaniclastic rocks.

Résumé

Dans la région cartographique de la rivière Iskut, des roches volcaniques de nature bimodale ou variant de neutre à mafique, du groupe de Stuhini du Trias supérieur, deviennent des grauwackes et des siltstones feldspathiques brun foncé de bassin à l'est et au nord-est. Le groupe de Stuhini du Trias supérieur passe progressivement au groupe de Hazelton du Jurassique près du ruisseau Treaty.

Le groupe de Hazelton comprend trois formations hétérogènes d'origine volcanique. La formation d'Unuk River, constituée de brèches andésitiques, de tufs et de siltstones siliceux marins, est la plus ancienne. Des conglomérats, des brèches et des grauwackes volcaniques et hétérogènes dont la couleur varie de rouge foncé à vert, de la formation de Betty Creek, recouvrent la première. La plus jeune, la formation de Mount Dilworth constituée de tufs et de brèches tufacées, est un important horizon marqueur régional, représentant l'événement volcanique culminant du volcanisme de Hazelton.

La formation de Salmon River (bajocienne ?) du Jurassique inférieur et moyen comprend: des shales silicieux et des turbidites tufacées à l'est; des laves en coussins médianes de nature felsique et mafique, des shales et des calcaires qui renferment le gisement d'Eskay Creek; et un faciès ouest hypothétique de roches volcanoclastiques, andésitiques et calco-alcalines.

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INTRODUCTION

As part of a continuing regional mapping program in Iskut River map area (NTS 104B), the 1989 summer's fieldwork focused on the eastern third of the map area (Fig. 1). Mesozoic strata and intrusions dominate the geology and host many of the mineral deposits under development, including the Eskay Creek deposit.

Abrupt facies change characterizes volcanogenic successions; Upper Triassic and Jurassic strata in the eastern third of the map area are good examples (Fig. 2). Between the eastern margin of the map area (130°W) and the Unuk River, Upper Triassic and Jurassic rocks grade from volcanic-rich to less volcanic, more distal and more basinal equivalents.

An important lower Middle Jurassic facies change occurs between the Bowser River and Snippaker Mountain. Eastern siliceous shale and tuff turbidite change to pillowed lava, micrite and siliceous shale that in turn appear to be equivalent to hornblende-phyric andesitic breccia farther west. The pillow lava-sedimentary rock sequence hosts the apparently stratabound Eskay Creek precious metal deposit. Similar strata extend 50-65 km north and south from Eskay Creek (Fig. 1).

Anderson (1989) presented a preliminary Paleozoic and Mesozoic stratigraphic, plutonic and structural framework for the map area. This report highlights Triassic and Jurassic strata and incorporates results from companion studies (e.g. Anderson and Bevier, 1990; Smith and Carter, 1990).

UPPER TRIASSIC STUHINI GROUP

In Iskut River map area, stratigraphy and compositional range of volcanic rocks define western and eastern facies of the Upper Triassic Stuhini Group (Anderson, 1989; Fig. 2). In the western facies, sedimentary rocks underlie a bimodal volcanic suite. Eastern facies sedimentary rocks interfinger with intermediate and mafic volcanic rocks but are more common at the top of the succession (e.g. the Storie, McTagg and Treaty creeks anticline). From the Stikine River east and from McQuillan Ridge northeast, both facies grade to a sequence dominated by sedimentary rocks, mainly dun feldspathic greywacke and orange and black siltstone.

Western facies

Coralline limestone and polymict cobble conglomerate characterize western facies sedimentary rocks (Fig. 2). Chert, limestone, greywacke and shale dominate the fragments. Breccia, felsic tuff, shale, and micrite overlie the limestone and conglomerate. Conodonts in the limy rocks are Late Triassic (Carnian or Norian) in age (Anderson, 1989; Orchard *in* Lefebure and Gunning, 1989). Near Snippaker Mountain, the lower sedimentary member changes from mainly limestone and polymict conglomerate to feldspathic greywacke and siltstone (Lefebure and Gunning, 1989), a common lithology farther northeast.

Western facies Stuhini Group volcanic rocks extend from the Stikine River at least as far east as Snippaker Mountain (Lefebure and Gunning, 1989). Aphyric felsic tuff interfingers with siliceous and limy shale of the lower sedimentary member. Towards the top, laminated dun mafic and white felsic tuff are common. Coarse pyroxene phenocrysts characterize mafic and felsic flows. Decimetre-scale growth faults are uncommon in the volcanic rocks. Synvolcanic earthquake activity associated with the bimodal volcanism may have triggered turbidity flows represented by the basal chert-limestone coarse clastic deposits.

Eastern facies

Eastern facies Stuhini Group rocks lack the thick, lower limestone and felsic volcanic rocks of the western facies (Fig. 2). Brown to dun feldspathic, locally calcareous greywacke and siltstone are distinctive. The thin-bedded, wavy-laminated siltstone and fine grained greywacke weather orange and black. Polymict pebble to boulder conglomerate, shale, and rare, thin coralline and crinoidal limestone are subordinate. Conglomerate and breccia fragments include: common aphyric to hornblende- or clinopyroxene-phyric andesite and basalt; lesser grey to black chert and brown feldspathic greywacke; uncommon grey limestone; and rare felsic volcanic fragments. Shale rip-up clasts are common in greywacke members of distinctive Tabe and Tae turbiditic facies.

Volcanic rocks in the eastern facies are generally intermediate or mafic. Dark green hornblende- or clinopyroxene-phyric andesitic and basaltic volcanic conglomerate and breccia are typical. Autobreccia is common. On McQuillan Ridge, a Late Triassic pluton intruded greenish grey aphyric to plagioclase$\pm$hornblende porphyritic tuff and subordinate siltstone (Anderson and Bevier, 1990). ***Halobia***-bearing sedimentary rocks occur with the volcanic rocks (Grove, 1986). The McQuillan Ridge sequence, considered Norian to Sinemurian by Alldrick et al. (1989), is part of the Upper Triassic Stuhini Group.

TRIASSIC TO JURASSIC TRANSITIONAL UNIT

Volcanogenic members of the Upper Triassic Stuhini Group and Lower Jurassic Hazelton Group have not been seen in contact. A Triassic-Jurassic (210 +24/-14 Ma) U-Pb zircon date for basal andesite of the Unuk River Formation (Brown, 1987) in the Stewart area hints at the possibility of a gradational contact between volcanic strata of the two groups.

A gradational contact is recognized between the sedimentary, basinal facies of the Stuhini Group and a condensed section of Hazelton Group volcanic rocks (Fig. 2). Near the headwaters of the Unuk River and Treaty Creek, Alldrick and Britton (1988) mapped a distinctive sedimentary unit (their unit 8b near Rounsefell and Atkins glaciers) that marks the gradational contact. A similar, but thinner unit occurs south of John Peaks.

Thin-bedded, orange-brown to black, wavy-laminated siltstone and shale, containing ***Monotis(?)*** or ***Halobia(?)***, are the uppermost recognizable Stuhini Group. Upsection, siltstone becomes more siliceous and contains increasingly abundant greywacke and conglomerate. Discontinuous lenses of pebble to boulder conglomerate occur with coarse greywacke. The clast-supported conglomerate is mainly volcanic clast-bearing (plagioclase$\pm$hornblende porphyry andesite and dacite). Soft sediment deformation features such as convolute bedding, channel scour and rip-up clasts characterize the coarse clastic rocks and record high energy influx of coarse detritus into a siltstone basin. The uppermost laminated siliceous siltstone, fine grained greywacke, and minor grey micrite, medium- to coarse- grained greywacke and local matrix- to clast-supported conglomerate suggest a return to relatively quiescent conditions. Plagioclase-phyric dacite sills and flows(?) interfinger with the uppermost strata and underlie a condensed Hazelton Group section of lava and volcanic breccia.

The Unuk River-Treaty Creek section appears to be structurally and stratigraphically continuous. A stratigraphic break between Triassic and Jurassic rocks may exist within the conglomerate-rich member but no extra-basinal fragment lithologies occur.

Granitoid- and dacite-bearing polymict conglomerate and greywacke distinguish the transitional unit south of John Peaks. The unit is thinner and the depositional break more abrupt than at the Unuk River-Treaty Creek section. Granitoid- and dacite-bearing polymict conglomerate and greywacke overlie Stuhini Group siltstone and volcanic conglomerate. As at Treaty Creek, a condensed section of Hazelton Group rocks overlies the polymict conglomerate. At John Peaks, the polymict conglomerate may represent an unconformity at the base of the Hazelton Group.

The transitional unit between Unuk River and Treaty Creek is undated but Sinemurian fauna (H.W. Tipper, pers. comm., 1985) occur in apparently similar rocks along strike to the southeast. Correlative strata may be part of unit lJwp mapped by Read et al. (1989) south of Leroy Creek. Thorkelson (1988) described similar conglomerate within or at the top of the Stuhini Group north of Cartmel Mountain in the Spatsizi map area (NTS 104H) to the northeast.

LOWER JURASSIC HAZELTON GROUP

Two-part subdivision of Jurassic rocks in the region has a long history. McConnell (1911), Schofield and Hanson (1922) and Hanson (1929) called a lower volcanogenic unit the Bear River formation. The Nass and Salmon River formations of McConnell (1910, 1911 and 1913), Schofield and Hanson (1922) and Hanson (1929, 1935) included upper, volcanic-poor, sedimentary strata. Hanson (1935) included the volcanic and sedimentary rock divisions in the Hazelton Group.

Grove (1986) adopted, refined and expanded the earlier nomenclature. Lower Jurassic Unuk River and Betty Creek formations were defined by Grove to include the lower volcanogenic strata. The Middle Jurassic Salmon River Formation and Upper Jurassic Nass Formation [Bowser Lake Group, in part] encompassed the overlying sedimentary rocks. Alldrick and Britton (1988) and Alldrick et al. (1989) recently defined the Lower to Middle Jurassic Hazelton Group to incorporate volcanogenic rocks (Lower Jurassic Unuk River, Betty Creek and newly established Mount Dilworth formations) and sedimentary rocks (Middle Jurassic Salmon River Formation).

The coherent, three-part regional stratigraphy of Hazelton Group volcanogenic strata is best exposed in the eastern third of the Iskut River map area (e.g. Grove, 1971, 1986; Alldrick, 1983-1985, 1987; Brown, 1986, 1987; Alldrick and Britton, 1988 and this work; Fig. 2). Among the Unuk River, Betty Creek and Mount Dilworth formations, the Mount Dilworth formation is the most homogeneous regional marker and represents the penultimate volcanic event in Hazelton Group volcanism.

Unuk River Formation

White and dun, locally deformed, andesitic volcanic breccia, thin-bedded hyaloclastite and lava dominate Unuk River Formation in eastern Iskut River map area. The breccia's matrix typically weathers white in contrast to the recessive, dun pebble to boulder fragments. Aphyric, plagioclase- and chloritized hornblende-phyric andesite fragments are characteristic. Rare serpentinized olivine basalt occurs within the volcaniclastics.

West of the Bowser River, the volcaniclastics grade into a variegated sedimentary unit. Waxy, rusty or white weathering, thin-bedded, wispy-laminated siliceous siltstone dominates the unit; black weathering pebble conglomerate and greywacke are subordinate. South of Frank Mackie Glacier, a 10 m wide dyke of alkali-feldspar-phyric "Premier porphyry" andesite crosscuts the siltstone. The intrusive relations indicate that the sedimentary rocks are not Salmon River Formation as mapped by Grove (1986) and Alldrick and Britton (1988) but are equivalent to Alldrick and Britton's (1988) middle or upper argillite unit within the Unuk River Formation.

East of Salmon and Frank Mackie glaciers, the top of the Unuk River Formation is a $\pm$hornblende-alkali-feldspar porphyry flow at least 20 m thick. In the Salmon Glacier area, the flow is coeval and texturally similar to the 189-195 Ma Premier porphyry dykes and Texas Creek plutonic suite (Alldrick et al., 1986, 1987). The lava also may be the upper bounding stratum for many precious metal veins (Alldrick, 1985).

Betty Creek Formation

Betty Creek Formation contains maroon to green volcanic siltstone, greywacke, conglomerate, breccia and rare lava with common sedimentary structures and anastomosing ferruginous or jasperoid veins. It overlies the Unuk River Formation conformably; locally the contact is gradational. The epiclastic nature, maroon colour and abundant ferruginous veining distinguish Betty Creek Formation. Its members are massive, thick- or medium- bedded. Poorly sorted, pebble to boulder volcanic conglomerate and breccia contain matrix-supported grey, green and purple aphanitic and ($\pm$hornblende-) plagioclase-phyric andesite fragments. Finer grained rocks appear to be reworked crystal and/or lithic tuffs. Load casts, ball and pillow structures, flame structures, graded beds, planar cross-stratification, convolute bedding, centimetre-scale growth faults, and small scale Tabe turbidites provide unequivocal tops indicators. The top of the Betty Creek Formation comprises maroon,

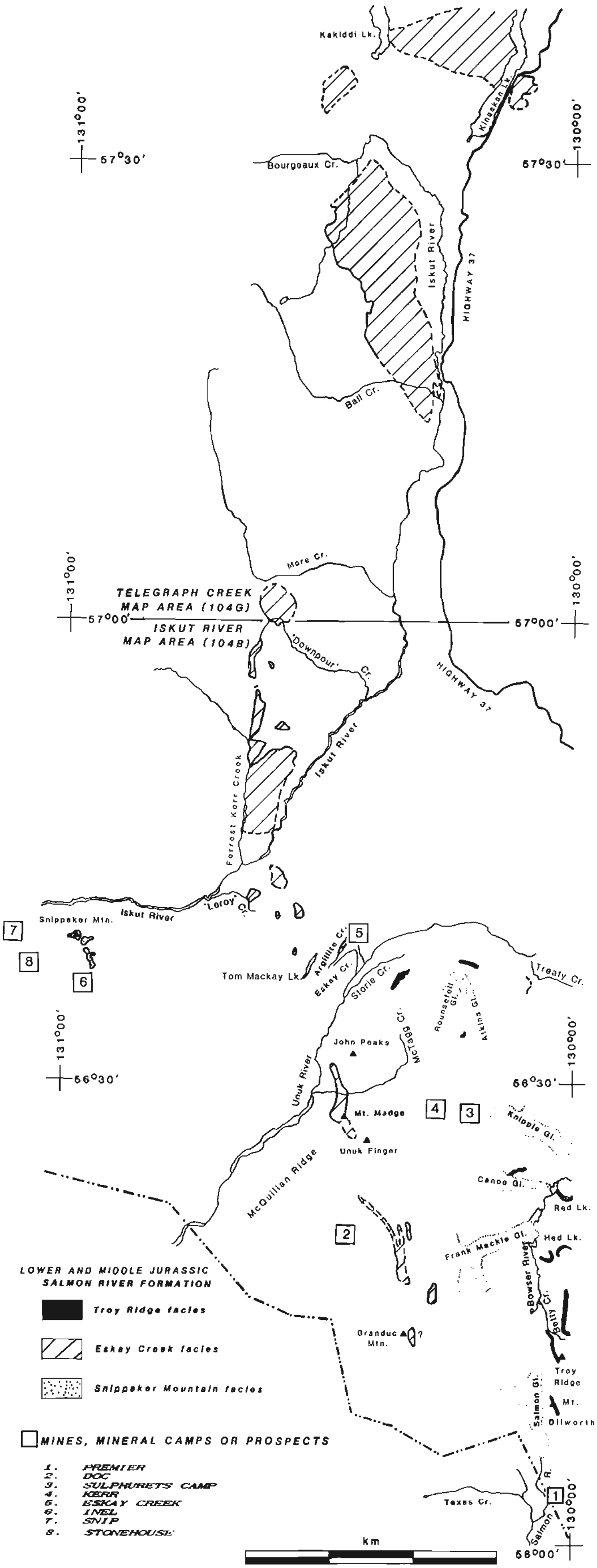

Figure 1. Index map for localities mentioned in text and distribution of Troy Ridge, Eskay Creek and Snippaker Mountain facies of Lower and Middle Jurassic Salmon River Formation.

dense to amygdaloidal, aphyric and serpentinized olivine-phyric basalt and andesite lava between Canoe and Knipple glaciers, and rare white limestone and black and white chert southeast of Hed Lake and south of John Peaks.

Mount Dilworth formation

Mount Dilworth formation is the least heterogeneous and most extensive marker within the Hazelton Group in eastern Iskut River map area (Fig. 2; Alldrick and Britton, 1988; Alldrick et al., 1989). Its white, maroon or green weathering, felsic tuff, tuff breccia and dust tuff are welded and non-welded and aphyric to sparingly plagioclase-phyric. Eutaxitic and spherulitic textures, flow-layering, and dacite to rhyolite composition distinguish it from locally similar Betty Creek Formation. Autobreccia is common. The "Fisheye sandstone" (spherulitic rhyolite) north of John Peaks (Grove, 1986) is probably part of this felsic member. West of the Bowser River, rare massive basalt dykes and flows(?) with bladed plagioclase or serpentinized olivine phenocrysts occur near the base. Mount Dilworth formation is no older than Pliensbachian in the Eskay Creek area (Smith and Carter, 1990) and is the product of the climactic, but penultimate, eruption of Hazelton volcanism. The widespread distribution and relatively consistent thickness of Mount Dilworth formation suggest that topographic relief was low during eruption of the ignimbrite and felsic lava.

LOWER AND MIDDLE JURASSIC SALMON RIVER FORMATION

Alldrick and Britton (1988) and Alldrick et al. (1989) included Lower to Middle Jurassic Spatsizi Group equivalent strata (Thomson et al., 1986; Anderson, 1989) and rocks of Middle and Upper Jurassic Bowser Lake Group within the Salmon River Formation. They considered it part of the Hazelton Group. We restrict the Salmon River Formation to upper Lower Jurassic and lower Middle Jurassic strata (Fig. 2) because of: 1) unconformable stratigraphic relations between rocks of Late Triassic (Stuhini Group) and late Early Jurassic age (Salmon River Formation) that indicate a marked depositional break; and 2) the known or suspected unconformity common at the base of the overlying Bowser Lake Group (Gunning, 1986; and this work). The usage is consistent with the sense of Schofield and Hanson's (1922) original description of the formation. A restricted Salmon River Formation also emphasizes one of its precious-metal rich subunits (Eskay Creek facies). The volcanic-poor nature, bounding unconformities and general basinal aspect (e.g. south to north thickening and change to distal facies) of Salmon River Formation could merit its exclusion from the volcanic-dominant Hazelton Group. However, the Salmon River Formation appears equivalent to the upper sedimentary part of the Hazelton Group formally defined farther south and east (Tipper and Richards, 1976; Fig. 2).

Salmon River Formation comprises two members (Fig. 2). A thin, belemnoid-rich, upper Lower Jurassic calcareous sandstone occurs at the base. The overlying lower Middle Jurassic member has three facies that form north-trending belts. The Troy Ridge facies, informally known as the "pajama beds," is a distinctive black siliceous, radiolarian-bearing shale and white reworked tuff turbidite that occurs in the east. Along and west of the Unuk River is a sequence of pillowed lava and limy to siliceous shale and siltstone of the Eskay Creek facies. This medial facies hosts the Eskay Creek prospect. The westernmost Snippaker Mountain facies consists of andesitic volcaniclastics (Lefebure and Gunning, 1989).

Upper Lower Jurassic unit

The unnamed lower member of the Salmon River Formation consists of rusty brown or green, fossiliferous, calcareous greywacke (Fig. 2). It forms a 60-100 cm thick unit between Mount Dilworth formation and the upper member of the Salmon River Formation. In the Storie Creek area (Gunning, 1986) and at Snippaker (unit lJs of Lefebure and Gunning (1989)), the unit is thick, well-dated (Toarcian) and unconformably overlies fossiliferous Upper Triassic Stuhini Group rocks.

In most places the unit is too thin to map. Along strike, it changes from sandy bioclastic limestone near Mt. Dilworth in the south to calcareous, less fossil-rich greywacke near Red Lake and Knipple Glacier. Northwest, at Storie Creek, it is thicker and siltstone-rich (Gunning, 1986). To the north and northwest, the Toarcian unit thickens (locally to more than 1500 m; Read et al., 1989) and becomes more basinal. Grey siltstone, subordinate greywacke and rare fossiliferous limestone characterize the basinal facies. It continues north-northwest along the ridge between More and "Downpour" creeks (*see* Read et al., 1989) to the region between More and Ball creeks and Iskut River in southeastern Telegraph Creek map area (Souther's (1972) unit 14).

The upper Lower Jurassic unit is important because it is the only consistently fossiliferous Jurassic unit in the map area. Belemnites are the most abundant fauna, but ***Weyla,*** trigoniid bivalves, scleractinian coral, gastropods, ammonites and bryozoa are uncommon. Association of belemnites and ***Weyla*** in a few localities indicate a Toarcian age (H.W. Tipper, written comm., 1985).

Lower Middle Jurassic unit

The unnamed upper part of the Salmon River Formation comprises three informal lower Middle Jurassic facies that form north-trending belts (Fig. 1, 2): the eastern Troy Ridge facies; the medial Eskay Creek facies that hosts precious metal veins at Eskay Creek; and the western Snippaker Mountain facies.

Troy Ridge facies

Black cherty, radiolarian-bearing shale and white reworked tuff make up the Troy Ridge facies (informally known as the "pajama beds"; Fig. 1, 2). The characteristic rhythmic alternation of thin (2-10 cm) shale and tuff beds suggests a turbiditic origin. The unit contains more shale to the north (e.g. north of the Knipple Glacier) and upsection towards the gradational contact with the basal unit of the Bowser Lake Group. Bed thickness and tuff:shale proportions

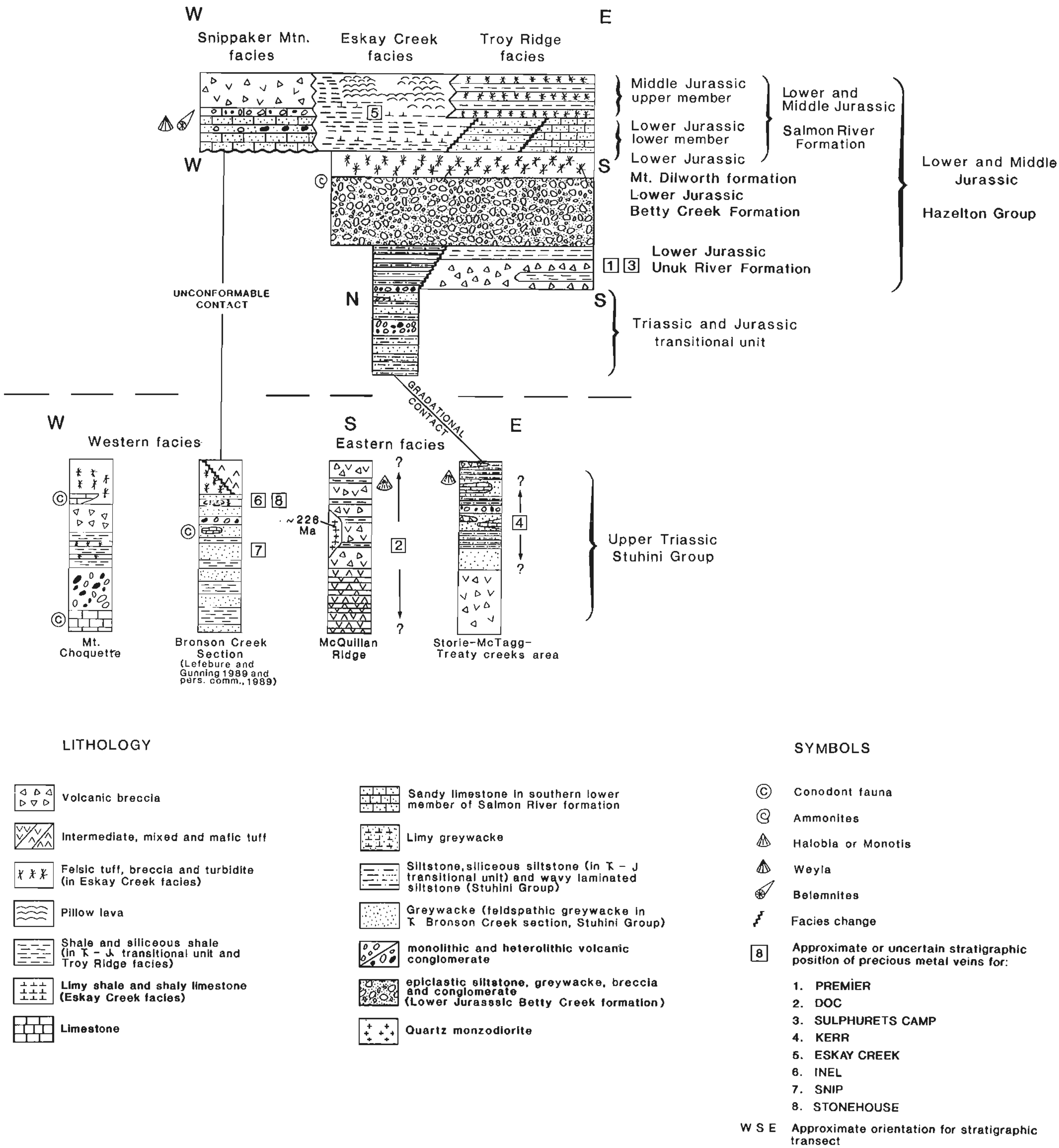

Figure 2. Schematic facies changes in Triassic and Lower and Middle Jurassic strata. Facies changes occur toward the east and northeast for Upper Triassic Stuhini Group and both south to north and east to west for Upper and Middle Jurassic Salmon River Formation in Iskut River map area.

increase near the thickest accumulations of the Hazelton Group, near more proximal facies of the Toarcian member and away from the Eskay Creek facies. The siliceous shale locally contains microscopic radiolaria(?) (Brown, 1987).

In eastern Iskut River map area, the Troy Ridge facies and the basal shale member of the Bowser Lake Group are the Salmon River Formation of Schofield and Hanson (1922), Grove (1986) Alldrick and Britton (1988), and Alldrick et al. (1989). We prefer to include the fossiliferous Toarcian sandstone and siliceous shale and tuff in the Salmon River Formation, distinct from the Bowser Lake Group. The division emphasizes the formation's bounding unconformities and its overall sedimentary nature. The Troy Ridge facies has strong lithological similarities to the Quock Formation of the Spatsizi Group in the Spatsizi map area (Thomson et al., 1986). It is also very similar to the Yuen Member of the Hazelton Group's Smithers Formation as defined by Tipper and Richards (1976) in the region to the south and southeast.

Eskay Creek facies

West of John Peaks, Mount Madge and Unuk Finger, limestone, limy or cherty siltstone and shale interfinger with and overlie thick pillow lava and pillow lava breccia (Fig. 1, 2). The cherty siltstone locally contains microscopic radiolaria(?) (J. Blackwell, pers. comm., 1989) and limy units are commonly rich in belemnites (G. McArthur, pers. comm., 1989). Pillow lavas are predominantly mafic and exceptionally well preserved at all scales (Read et al., 1989). Interpillow interstices are locally filled with limestone (Grove, 1986 and this work). Rare rhyolitic pillow lava and breccia interlayered with the pillow basalt occur on Mt. Madge and southeast of the Eskay Creek deposit. Syn-volcanic hydrothermal alteration of the lavas is locally characteristic (Souther, 1972) and in places, the felsic lavas may be silicified basalt.

Eskay Creek facies is stratigraphically equivalent to the Troy Ridge facies "pajama beds" farther east. The Eskay Creek facies overlies the Lower Jurassic Mount Dilworth formation and underlies the Middle Jurassic basal unit of the Bowser Lake Group of Bathonian age (Fig. 2; Gunning, 1986). Radiolarian-bearing limestone fragments in the basal conglomerates of the Bowser Lake Group yielded radiolaria of late Middle Toarcian to Early Bajocian age (Smith and Carter, 1990). The stratigraphic and geographic positions of Bajocian fauna reported by Donnelly (1976) from the horizon at Eskay Creek are uncertain. In Telegraph Creek map area, uppermost sedimentary rocks of Souther's (1972) unit 14, which underlie the pillowed lava sequence of unit 15, yielded Early Bajocian ammonites.

The Eskay Creek facies extends along strike 40-65 km north and south of Eskay Creek. In the north, up to 2000 m of pillowed aphyric to augite-phyric basalt lava interfinger with siltstone, tuffaceous wacke and conglomerate between Iskut River and Forrest Kerr Creek (unit Jvb of Read et al., 1989). East of Iskut River between Ball and Bourgeaux creeks and between Kinaskan and Kakiddi lakes, up to 2400 m of dark grey aphanitic to plagioclase-phyric basaltic-andesite pillow lava (unit 15 of Souther, 1972) forms a north-trending belt in southeastern Telegraph Creek map area (Fig. 1). In the south, the unit extends from Eskay Creek at least 40 km south along the western and southern flanks of John Peaks and Mt. Madge to at least Mount Pearson and perhaps to Granduc Mountain (Grove, 1986, unit 11a). South of Mt. Madge, pillowed lavas 200-2000 m thick interfinger with siltstone, chert, and limestone lenses (Grove, 1986).

The Eskay Creek precious metal veins are apparently stratabound within this horizon (G. McArthur, pers. comm., 1989). This and the richness of the Eskay Creek prospect suggest that exploration should be concentrated on the Eskay Creek facies.

Snippaker Mountain facies

Near Snippaker Mountain, plagioclase±hornblende-phyric andesite lavas and breccia (unit lmJv of Lefebure and Gunning, 1989) overlie Toarcian sandy limestone, limy conglomerate and limy sandstone in about the same stratigraphic horizon as the Eskay Creek and the Troy Ridge facies (Fig. 1, 2). If equivalent, the distribution of Middle Jurassic rocks might represent a presently west-facing calc-alkaline volcanic arc in the west with rift-facies submarine pillow lavas and sedimentary rocks in the middle and distal, basinal volcanogenic turbidite in the east. These belts may be the western part of the adjacent Spatsizi shale basin described by Thomson et al. (1986) and record early Middle Jurassic opening of a backarc basin just west of (and as a progenitor to) the Bathonian and younger Bowser Basin.

MIDDLE AND UPPER JURASSIC BOWSER LAKE GROUP

The Middle and Upper Jurassic Bowser Lake Group facies also change but less systematically than in the Salmon River Formation. Pencil shale generally occurs at the base of the Jurassic Bowser Lake Group along the Bowser River. Between Mt. Dilworth and Troy Ridge, basal greywacke and shale turbidite grade upsection to pencil shale and siltstone. Along the northern margin of Knipple Glacier, basal chert-pebble conglomerate and greywacke overlie the upper Lower Jurassic member of Salmon River Formation. Troy Ridge facies rocks are absent and the contact may be an unconformity.

The gradational contact between Salmon River Formation and overlying Bowser Lake Group is best exposed near Red Lake. Upsection, greywacke beds increase at the expense of fine grained tuff and cherty shale of the Troy Ridge facies. The shale:greywacke ratio also increases upsection.

The basal Bowser Lake Group is best dated in the westernmost outcrops between Tom Mackay Lake and Eskay Creek. Lowermost and uppermost members of the Tom MacKay Lake sequence yielded Bathonian to Callovian ammonites (P.L. Smith, pers. comm., August, 1989; Gunning, 1986). The basal unit is also the most coarse grained. It overlies the lower Middle Jurassic Eskay Creek facies in structural (and stratigraphic?) conformity. Basal siltstone, shale and minor greywacke changes upsection to resistant, metre-thick beds of white quartz arenite and chert pebble conglomerate. Black, grey and green chert, quartz and rare felsic volcanic (Mount Dilworth formation equivalent?) dominate the conglomerate fragments. The upper unit comprises rhythmically interbedded siltstone and fine grained greywacke.

Northeastward from Tom MacKay Lake to the northeast corner of Iskut River map area, monotonous Middle and Upper Jurassic greywacke and shale predominate. Felsite-, quartz- and chert-bearing pebble conglomerate is rare. The unit contains sparse Callovian, Oxfordian and Oxfordian to Kimmeridgian fauna (Anderson, 1989; Read et al., 1989).

MESOZOIC STRATIGRAPHY AND MINERAL DEPOSITS

Precious and base metal veins being developed in the area occur within Upper Triassic (e.g. Kerr, Doc, Inel, Snip and Stonehouse deposits), Lower Jurassic (e.g. Premier and Sulphurets deposits) and lower Middle Jurassic (e.g. Eskay Creek deposit) strata (Figs. 1, 2). For many deposits (e.g. Premier, Kerr, Inel and Snip), proximity to Early Jurassic calc-alkaline to alkaline plutonic and porphyritic intrusions, especially the alkali-feldspar porphyry variety (Premier porphyry), seems to be the main control. The host strata are of secondary importance.

An important exception is the Eskay Creek deposit. It appears to be stratabound within the siliceous to limy sedimentary rocks and pillowed lava sequence of the Eskay Creek facies of the Salmon River Formation. The backarc basinal environment inferred for the Eskay Creek facies rocks may be important in localization of Eskay Creek-type precious metal mineralization. If so, a newly-recognized lower Middle Jurassic stratigraphic horizon, extending more than 100 km from southeastern Telegraph Creek to southeastern Iskut River map areas, warrants closer inspection for other precious metal-rich prospects.

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A note on Mesozoic and Tertiary K-Ar geochronometry of plutonic suites, Iskut River map area, northwestern British Columbia

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Abstract

Thirteen new K-Ar hornblende (hb) and biotite (bi) dates, preliminary U-Pb zircon dates, and existing geochronometry suggest Late Triassic (213-226 Ma), Early Jurassic (189-196 Ma), Middle Jurassic (175-180 Ma) and post-tectonic, Eocene (44-62 Ma) plutonic episodes in Iskut River map area.

Plutons dated include: Late Triassic Seraphim Mountain pluton (213 ± 4 Ma, hb); Early Jurassic McLymont Creek pluton (189 ± 3 Ma, hb); Middle Jurassic Warm Springs Mountain pluton (159 ± 2 Ma, bi; date reset from ca. 177 ± 2 Ma U-Pb date for zircon) and Zippa Mountain syenite complex (98 ± 2 Ma, hb, and discordant mineral pair: 167 ± 4 Ma, hb, and 77 ± 1 Ma, bi); and Tertiary Saddle Lake pluton (62 ± 2 Ma, hb; 53 ± 1 Ma, bi; and mineral pair 58 ± 2 Ma, bi, and 54 ± 1 Ma, hb) and Great Glacier pluton (52 ± 1 Ma, bi, and mineral pair 53 ± 4 Ma, hb, and 51 ± 1 Ma, bi).

Résumé

Treize nouvelles datations au K-Ar sur hornblende (hb) et biotite (bi), des datations préliminaires obtenues par la méthode U-Pb appliquée aux zircons et la géochronométrie existante semblent indiquer qu'il y a eu dans la région cartographique de la rivière Iskut, des épisodes plutoniques au Trias supérieur (213 à 180 Ma) au Jurassique inférieur (189 à 196 Ma), au Jurassique moyen (175 à 180 Ma) et à l'Éocène post-tectonique (44 à 62 Ma).

Les plutons datés comprennent: le pluton de Seraphim Mountain du Trias supérieur (213 ± 4 Ma, hb); le pluton de McLymont Creek du Jurassique inférieur (189 ± 3 Ma, hb), le pluton de Warm Springs Mountain du Jurassique moyen (159 ± 2 Ma, bi; ancienne datation U-Pb sur zircon d'environ 177 ± 2 Ma) et le complexe syénitique de Zippa Mountain (98 ± 2 Ma, hb, et couple minéral discordant: 167 ± 4 Ma, hb et 77 ± 1 Ma, bi); et le pluton de Saddle Lake du Tertiaire (62 ± 2, hb, 53 ± 1 Ma, bi; et couple minéral 58 ± 2, bi, et 54 ± 1, hb) et le pluton de Great Glacier (52 ± 1 Ma, bi, et paire minérale de 53 ± 4 Ma, hb, et 51 ± 1 Ma, bi).

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INTRODUCTION

Granitic rocks in the Iskut River map area (NTS 104B) 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.

Plutonic styles established in earlier field mapping (e.g. Anderson, 1989) and the existing geochronometric database (Smith, 1977; Brown, 1987; Alldrick et al., 1986, 1987, 1989) suggested three probable and three speculative plutonic episodes. Compositional spectrum, mineralogy, texture, fabric, spatial proximity with coeval volcanics, and distinctive intraplutonic porphyritic dykes provide criteria to group plutons. Late Triassic Stikine, Early Jurassic Texas Creek, and Eocene Hyder plutonic suites occur in the map area (Fig. 1; Anderson, 1989; Woodsworth et al., 1989a,b).

Reconnaissance and detailed K-Ar and U-Pb dating is underway to complement the regional mapping of Iskut River map area (NTS 104B; e.g. Anderson, 1989; Anderson and Thorkelson 1990; Read et al., 1989) and biostratigraphic studies (e.g. Smith and Carter, 1990). K-Ar mineral dates reported here for hornblende and biotite from some plutons are the preliminary results. In a few cases, K-Ar dates compare closely with unpublished U-Pb isotopic data for zircon from the same sample; the U-Pb data will be published elsewhere. In all cases, the K-Ar dates provide minimum age estimates for the plutons.

New and earlier-determined isotopic ages for Iskut plutonic rocks confirm the Late Triassic, Early Jurassic and Tertiary (Eocene) episodes and show the importance of Middle Jurassic plutonism. The Mesozoic episodes are coeval with eruption of Upper Triassic Stuhini Group, Lower Jurassic Hazelton Group and lower Middle Jurassic Salmon River Formation (Eskay Creek facies) volcanic rocks.

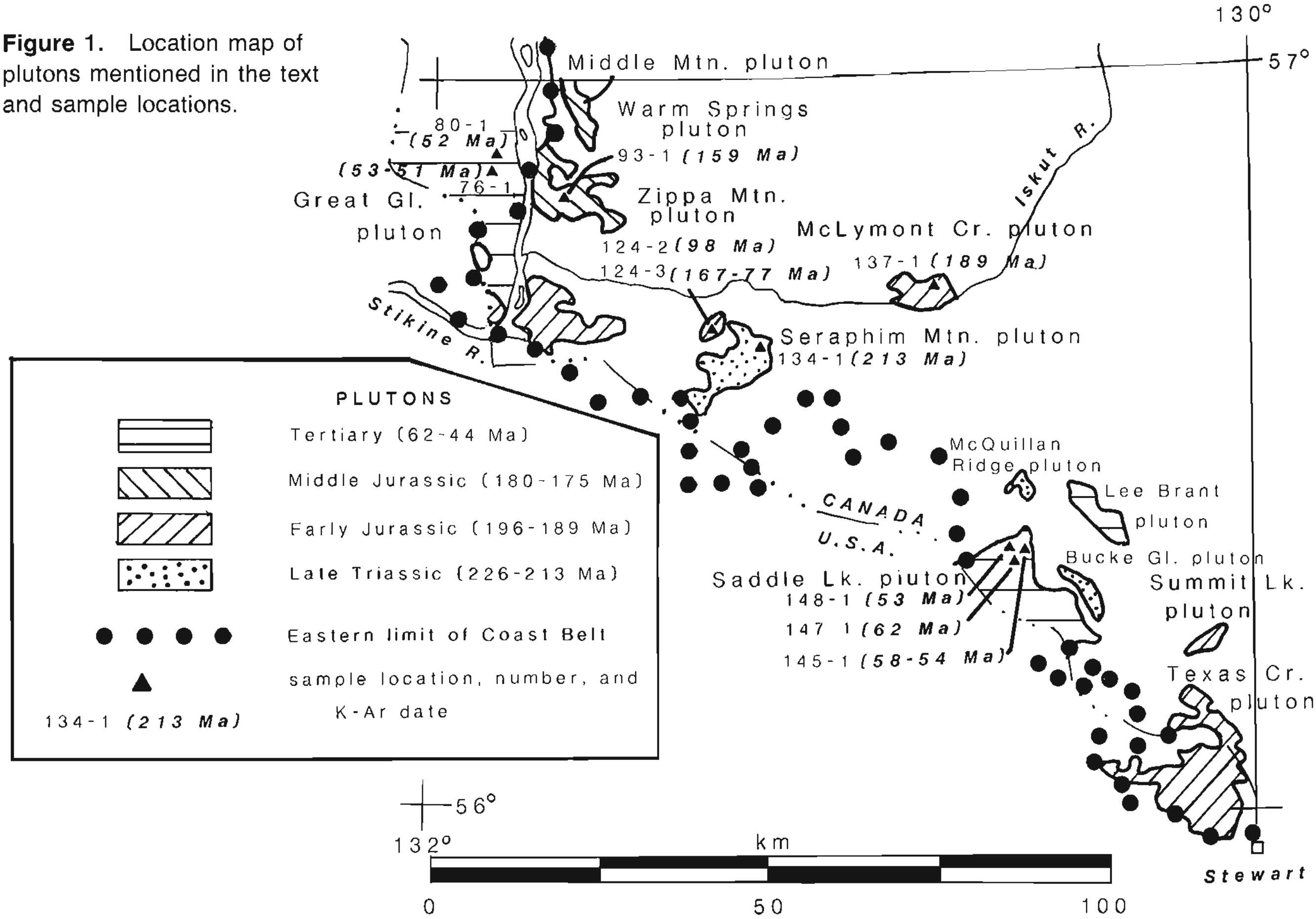
GEOLOGICAL SETTING

Anderson (1989), Thorkelson (1990) and Anderson (1990) reviewed the geological setting of the Iskut River map area. The outline of Mesozoic stratigraphy below emphasizes periods of Mesozoic volcanism.

Upper Triassic Stuhini Group and Late Triassic Stikine plutonic suite

Upper Triassic Stuhini Group volcanic and sedimentary rocks change from west to east and south to north and comprise two facies. Western facies rocks include basalt and rhyolite flows and tuff that overlie an important, well-dated coralline limestone. Chert and limestone boulder conglomerate, shale, and micrite are subordinate lithologies. Conodonts in the limestone marker and in micrite intercalated with the volcanic rocks are Late Triassic (locally Carnian and Norian; Anderson, 1989; Lefebure and Gunning, 1989). Coarse grained clinopyroxene ± hornblende phenocrysts are characteristic of the mafic and, locally, of the felsic flows.

Figure 1. Location map of plutons mentioned in the text and sample locations.



Eastern facies rocks lack the widespread limestone marker and the felsic volcanic rocks of the western facies. Brown to dun feldspathic and locally calcareous greywacke and orange and black siltstone characterize sedimentary eastern facies rocks. Polymict pebble and boulder conglomerate and shale are subordinate and thin coralline and crinoidal limestone rare.

Volcanic rocks in the eastern facies are generally intermediate or mafic. Dark green hornblende- or clinopyroxene-phyric andesite- and basalt-bearing volcanic conglomerate, breccia and autobreccia are typical. A Late Triassic pluton (see below) intruded greenish grey aphyric to plagioclase$\pm$hornblende porphyritic tuff and subordinate siltstone on McQuillan Ridge. The strata, considered Norian to Sinemurian by Alldrick et al. (1989), are part of the Upper Triassic Stuhini Group.

Anderson (1989) considered the Late Triassic Stikine plutonic suite to encompass clinopyroxene gabbro and diorite, biotite-hornblende quartz monzodiorite and biotite alkali-feldspar megacrystic quartz monzonite. The plutons were crosscut by east-trending mafic or bimodal dykes whose compositions vary systematically with that of cospatial, thick sequences of Stuhini Group volcanic host rocks. The dykes linked plutonism and volcanism (Anderson, 1989). Warm Springs and McLymont Creek plutons were included in the Stikine plutonic suite because of the characteristic dykes. However, they yielded Jurassic K-Ar or U-Pb dates and belong to the Middle Jurassic and Early Jurassic suites, respectively.

Lower Jurassic Hazelton Group and Early Jurassic Texas Creek plutonic suite

Lower Jurassic Hazelton Group comprises three internally heterogenous units. Andesitic breccia, tuff and marine siliceous siltstone of the Unuk River Formation are lowermost. Betty Creek Formation comprises heterogeneous maroon to green, epiclastic volcanic conglomerate, breccia and greywacke that conformably overly the Unuk River Formation. Welded or nonwelded, rhyolite to dacite tuff and tuff breccia of the Mount Dilworth formation are uppermost. The felsic rocks are an important regional marker and represent a climactic volcanic event of Hazelton Group volcanism.

A well dated, 20 m thick, $\pm$hornblende-alkali-feldspar porphyry lava of Early Jurassic age marks the top of the Unuk River Formation (Alldrick, 1985, 1987; Brown, 1987). It is coeval (see below) and texturally similar to the 189-195 Ma "Premier porphyry" dykes and Texas Creek plutonic suite (Alldrick et al., 1986, 1987; Brown, 1987). The flow may be the upper bounding stratum for many precious metal veins as well (Alldrick, 1985).

Texas Creek plutonic suite comprises calc-alkaline and alkaline varieties. Calc-alkaline biotite-hornblende quartz monzodiorite and granodiorite plutons are crosscut by alkali-feldspar-phyric andesite dykes (e.g. Texas Creek pluton and "Premier porphyry" dykes near Silbak Premier mine). Alldrick et al. (1986, 1987) and Brown (1987) documented the close spatial, temporal and genetic link among the Texas Creek plutonic suite, Unuk River Formation feldspar-phyric volcanic rocks, and Premier porphyry dykes. Prismatic hornblende and common, widespread alteration of mafic minerals and plagioclase to chlorite and epidote are characteristic of the calc-alkaline phases. Alkaline syenite and $\pm$plagioclase$\pm$hornblende-alkali-feldspar porphyry intrusions of the Texas Creek plutonic suite are widespread (e.g. in the Sulphurets-Kerr district) and associated with porphyry copper and gold deposits.

Anderson (1989) included Middle Mountain pluton and Katete Mountain pluton as western members of the Texas Creek plutonic suite because of similarities in composition, mafic mineralogy and alteration state. However, they lack Premier porphyry dykes. Preliminary U-Pb geochronometry shows a Middle Jurassic age for Middle Mountain pluton and an Early Jurassic age for Katete Mountain pluton.

Middle Jurassic Salmon River Formation, Eskay Creek facies volcanic rocks

Anderson and Thorkelson (1990) suggested that strata between the Lower Jurassic Mount Dilworth formation felsic tuff and Middle and Upper Jurassic Bowser Lake Group siliciclastic rocks encompass the Salmon River Formation. Three known or suspected facies belts characterize the formation. Turbiditic, alternating siliceous shale and tuff in the east (Troy Ridge member) change to a medial belt of intermediate to mafic pillowed lava, shale and limestone (Eskay Creek facies) that hosts the Eskay Creek deposit. A speculative western facies (Snippaker Mountain member) comprises hornblende-phyric, andesitic volcaniclastic rocks.

No Middle Jurassic plutons had been recognized in the Iskut River map area. Plutons with Middle Jurassic (170-180 Ma) isotopic dates are widespread (e.g. Anderson (1983), Holbek (1988) and Brown and Gunning (1989 and pers. comm., 1989)) and make up the Three Sisters plutonic suite to the north and northeast (Woodsworth et al., 1989a,b).

Tertiary Hyder plutonic suite of the Coast Belt

The eastern margin of the Coast Belt in Iskut River map area is mainly the intrusive contact between Tertiary plutons and Mesozoic country rock. Tertiary plutons (44-54 Ma) of the Hyder plutonic suite are more siliceous, biotite-rich, and less altered or deformed than Mesozoic plutons to the east (Buddington, 1929; Smith, 1977; Woodsworth et al., 1989a,b; R.G. Anderson and M.L. Bevier, unpub. U-Pb data). Monzogranite, quartz monzonite and granodiorite predominate but minor monzodiorite and microdiorite (e.g. Portland dyke swarm) occur. The plutons lack dykes and preserved volcanic equivalents. Tertiary plutons crosscut all regional structural fabrics and are post-tectonic. Locally, where Tertiary diorite-granite dykes (e.g. Portland dyke swarm) make up 30% of the outcrop, they suggest an equivalent crustal extension (Brown, 1987).

Fresh biotite monzogranite extends south from Mount Seraphim at least to Mount Dick (Kerr, 1948). Anderson (1989) included it as part of the Hyder plutonic suite because of its lack of alteration and dykes, mineralogy, and proximity to the inferred eastern Coast Belt margin. However, a Late Triassic hornblende K-Ar date suggests that the pluton is part of the Stikine plutonic suite.

Biotite minette dykes represent Oligocene-Miocene (18-25 Ma; Brown, 1987; Alldrick et al., 1987) ultrapotassic magmatism in the map area. Anderson (1989) included the fresh Zippa Mountain gabbro-syenite-quartz monzonite complex south of the Quaternary peralkaline Hoodoo Mountain volcano with the Oligocene-Miocene alkaline suite. The complex consists of sequentially-intruded gabbro, syenite, quartz monzonite and hornblende-plagioclase porphyry. New K-Ar dates for the two youngest phases are discordant, Jura-Cretaceous (167-77 Ma) but suggest a likely Jurassic age for the pluton.

GEOCHRONOMETRY

Previous Work

Smith (1977), Alldrick et al. (1986, 1987), and Brown (1987) established a K-Ar and U-Pb geochronometric framework for the southeastern part of Iskut River map area. Early Jurassic (193-195 Ma) Texas Creek and Summit Lake plutons were crosscut by Premier porphyry alkali-feldspar-phyric andesite dykes of about the same age (189-195 Ma). The feeder dykes are similar to an alkali-feldspar porphyry andesite flow, dated at 190 ± 2 Ma (U-Pb; Brown, 1987), at the top of the Hazelton Group's lowermost Unuk River Formation. Other dated andesite lavas of the Unuk River Formation suggest that the Hazelton Group may be as old as earliest Jurassic or latest Triassic (U-Pb date of 210 +24/-14 Ma; Brown, 1987). Brown (1987) dated the Monitor Lake rhyolite, part of the uppermost Mount Dilworth formation of the Hazelton Group, at 197 ± 14 Ma (U-Pb date).

K-Ar and Rb-Sr mineral dates for Jurassic units show resetting or are not easily interpretable. Dates range from 108-211 Ma (Smith, 1977; Brown, 1987) and mineral pairs are discordant. The oldest K-Ar date for hornblende from the Texas Creek pluton (211 ± 12; Smith, 1977) is consistent neither with the geological relationships nor with the U-Pb geochronometry described above.

K-Ar dates for sericite alteration of the units are similarly wide-ranging, 101-43 Ma, and interpreted as resulting from mid-Cretaceous regional greenschist metamorphism (Alldrick et al., 1987). An Eocene K-Ar date (45 ± 3 Ma) for biotite from lineated Texas Creek monzodiorite reflects resetting from nearby Hyder plutonic suite dyke swarms (Brown, 1987).

Eocene K-Ar and Rb-Sr dates for plutonic rocks range from 44-54 Ma (Smith, 1977; Brown, 1987; Alldrick et al., 1989). The data include few mineral pairs; two of three hornblende-biotite mineral pair dates reported by Smith (1977) for Davis, Hyder and Boundary plutons are concordant.

Analytical methods and data

Table 1 includes thirteen K-Ar dates for hornblende (hb) and biotite (bi) from ten samples. Analytical techniques for K determinations and Ar analysis follow Roddick and Souther (1987) and uncertainties are at two standard deviations. The samples collected by R.G. Anderson and J. Beekmann are representative of outcrops of plutons thought to be part of the Late Triassic Stikine plutonic suite (Warm Springs Mountain, and McLymont Creek plutons), Tertiary Hyder plutonic suite (Great Glacier, Saddle Lake, and Seraphim Mountain plutons) and the Oligocene-Miocene alkaline suite (Zippa Mountain pluton) based on field relations and plutonic style.

Uranium-lead isotopic analyses are underway for these and other samples. Preliminary U-Pb data for zircons from Mitchell Glacier (J. Mortensen and R. Kirkham, pers. comm., 1989), McLymont Creek, Warm Springs Mountain, Great Glacier and Saddle Lake plutons are available. Below, we discuss the best age estimates from this data compared with the new K-Ar dates.

K-Ar results

The K-Ar cooling dates for hornblende (hb) and biotite (bi) are surprising and contradict the field classification of plutons in the Iskut River region. However, there are few stratigraphic constraints on the age of the plutons sampled except that they postdate Late Triassic Stuhini Group volcanism. The dates suggest four plutonic episodes and their cooling intervals (Table 1 and Fig. 2):

1. Late Triassic (Seraphim Mountain pluton, 213 ± 4 Ma, hb)
2. Early Jurassic (McLymont Creek pluton, 189 ± 3 Ma, hb)
3. Middle Jurassic (Warm Springs Mountain pluton (159 ± 2 Ma, bi, date reset from ca. 177 ± 2 Ma U-Pb zircon date) and Zippa Mountain syenite complex (discordant 98 ± 2 Ma, hb, and mineral pair 167 ± 4 Ma, hb, and 77 ± 1 Ma, bi, dates)
4. Tertiary (Saddle Lake pluton (62 ± 2, hb, 53 ± 1 Ma, bi, and mineral pair 58 ± 2, bi, and 54 ± 1, hb) and Great Glacier pluton (52 ± 1 Ma, bi, and mineral pair 53 ± 4 Ma, bi, and 51 ± 1 Ma, hb).

DISCUSSION

Mesozoic plutons

The Late Triassic date for Seraphim Mountain pluton is surprising. The pluton is felsic, siliceous, massive, dyke-poor, fresh and near the east margin of the Coast Belt. It is similar to other well-dated plutons of the Tertiary Hyder plutonic suite (e.g. Great Glacier and Saddle Lake plutons). Seraphim Mountain pluton intruded western facies rocks of the Stuhini Group (Kerr, 1948). The felsic composition of western facies Stuhini Group tuff and volcaniclastic rocks is consistent with a felsic plutonic analog such as the Seraphim Mountain pluton. U-Pb data for Bucke Glacier pluton (about 221 Ma; R.G. Anderson and P. van der Heyden, unpub. data) and McQuillan Ridge pluton (about 226 Ma; M.L. Bevier, unpub. data) suggest that a northwest-trending belt of Late Triassic plutons occurs just northeast of the Coast Belt's eastern margin. These data show that it is difficult to separate Tertiary and Triassic felsic plutons near the Coast Belt's eastern margin on lithology and field relations alone.

McLymont Creek pluton is coeval with the alkaline variety of the Texas Creek plutonic suite. The pluton's alkali feldspar megacrysts suggest it is also texturally and compositionally similar to Texas Creek alkaline plutons. The east-southeast-trending mafic dykes that characterize the pluton must be Early Jurassic or younger. The K-Ar date for hornblende is similar to the preliminary age estimate from U-Pb analyses of zircon and sphene from the same sample (about 192 Ma).

Middle Jurassic ages for the Warm Springs and Zippa Mountain plutons are less easily interpreted. The date for the Warm Springs Mountain (159 ± 2 Ma, bi) is discordant with the best age estimate from U-Pb analyses on zircon from the same sample (about 177 Ma). Biotite may have been partially reset during intrusion of the Eocene Great Glacier pluton 5 km to the west. Discordant Late Jurassic dates from Middle Jurassic plutons are poorly understood but seem to characterize Middle Jurassic plutons in southwestern Telegraph Creek map area (Holbek, 1988; Brown and Gunning, 1989 and pers. comm., 1989) and along the Stikine Arch (Anderson, 1983). These plutons are part of the Three Sisters plutonic suite (Woodsworth et al., 1989a,b).

Table 1. New K-Ar data for hornblende and biotite from Mesozoic and Tertiary plutons in the Iskut River map area.

Sample number	G.S.C. lab Number	Locality	K (wt. %)	Rad. ^{40}Ar ((10^{-7}) cc/g STP)	% atmos. ^{40}Ar	Age (± 2 s.d.[1] Ma) (Hb = hornblende) (Bi = biotite)
			TERTIARY PLUTONS			
			SADDLE LAKE PLUTON			
ATB-85-145-1	4069	Saddle Lake[2]	0.582	12.34	17.0	53.7 ± 1.0 (Hb)
"	4070	"	6.34	144.9	54.0	57.9 ± 1.6 (Bi)
ATB-85-148-1	4071	Saddle Lake[3]	6.81	142.0	13.0	52.8 ± 1.2 (Bi)
ATB-85-147-1	4072	Saddle Lake[4]	0.454	11.05	58.0	61.6 ± 1.8 (Hb)
			GREAT GLACIER PLUTON			
AT-86-80-1	4073	Snowcap Mountain[5]	6.97	139.4	8.8	50.7 ± 0.8 (Bi)
"	4074	"	0.632	13.22	34.0	53.0 ± 4.3 (Hb)
AT-86-76-1	4075	Snowcap Mountain[6]	6.07	124.2	4.0	51.9 ± 0.8 (Bi)
			MIDDLE JURASSIC PLUTONS			
			ZIPPA MOUNTAIN PLUTON			
AT-86-124-3	4066	Zippa Mountain[7]	0.935	63.65	25.0	167 ± 4 (Hb)
"	4068	"	6.52	198.4	7.5	76.6 ± 1.3 (Bi)
AT-86-124-2	4080	Zippa Mountain[8]	0.977	38.4	9.2	98.4 ± 1.6 (Hb)
			WARM SPRINGS MOUNTAIN PLUTON			
AT-86-93-1	4076	Warm Springs Mountain[9]	6.89	445.6	1.1	159 ± 2 (Bi)
			EARLY JURASSIC PLUTONS			
			McLYMONT CREEK PLUTON			
AT-86-137-1	4078	McLymont-Forrest Kerr creeks[10]	0.781	60.54	5.4	189 ± 3 (Hb)
			TRIASSIC PLUTONS			
			SERAPHIM MOUNTAIN PLUTON			
AT-86-134-1	4077	Seraphim Mountain[11]	0.563	49.42	5.2	213 ± 4 (Hb)

1. s.d. = standard deviation.
2. hornblende-biotite quartz monzodiorite; UTM (zone 9) 6246040 N, 404870 E; 56°21'05" N, 130°32'22" W (NTS 104 B/7); 2.5 km north-northeast of east end of Saddle Lake, 7.8 km east-southeast of southwest end of Flory Lake, 5000' elevation; collected by J. Beekmann.
3. hornblende-biotite quartz monzodiorite; UTM (zone 9) 6246070 N, 402520 E; 56°21'04" N, 130°34'38" W (NTS 104 B/7); 2.0 km north-northwest of west end of Saddle Lake, 5.75 km southeast of southwest end of Flory Lake, 4300' elevation; collected by J. Beekmann.
4. hornblende-biotite quartz monzodiorite; UTM (zone 9) 6245160 N, 403600 E; 56°20'36" N, 130°33'34" W (NTS 104 B/7); 1.5 km north-northwest of west end of Saddle Lake, 7.0 km southeast of southwest end of Flory Lake, 5000' elevation; collected by J. Beekmann.
5. hornblende-biotite monzogranite; UTM (zone 9) 6309690 N, 327610 E; 56°54'03" N, 131°49'50" W (NTS 104 B/13); 2.8 km east of Snowcap Mountain, 3.8 km northeast from Icecap Mountain, 3500' elevation.
6. hornblende-biotite monzogranite; UTM (zone 9) 6305650 N, 326900 E; 56°51'51" N, 131°50'22" W (NTS 104 B/13); north of Great Glacier, 3.5 km southeast of Icecap Mountain, 4.5 km south-southeast of Snowcap Mountain, 2800' elevation.
7. biotite-hornblende quartz monzonite; UTM (zone 9) 6281070 N, 357080 E; 56°39'14" N, 131°19'53" W (NTS 104 B/11); 1.25 km west of Zippa Mountain, 4.65 km south-southwest of confluence of Zippa Creek and Iskut River; 4150' elevation.
8. hornblende-plagioclase porphyry dyke; UTM (zone 9) 6281070 N, 357080 E; 56°39'14" N, 131°19'53" W (NTS 104 B/11); 1.25 km west of Zippa Mountain, 4.65 km south-southwest of confluence of Zippa Creek and Iskut River; 4150' elevation.
9. biotite-hornblende quartz monzodiorite; UTM (zone 9) 6302700 N, 336200 E; 56°50'28" N, 131°41'07" W (NTS 104 B/13); just west of Warm Springs Mountain, 2.25 km east-southeast of The Knob; 4650' elevation.
10. hornblende-biotite monzogranite containing alkali-feldspar megacrysts; UTM (zone 9) 6288000 N, 392500 E; 56°43'32" N, 130°45'24' W (NTS 104 B/10); 6.75 km west-southwest of confluence of Forrest Kerr Creek and Iskut River, 4.75 km north-northeast of confluence of McLymont Creek and Iskut River; 4900' elevation.
11. hornblende-biotite quartz monzonite; UTM (zone 9) 6278000 N, 365350 E; 56°37'43" N, 131°11'42" W (NTS 104 B/11); at summit of Seraphim Mountain, 5524' elevation

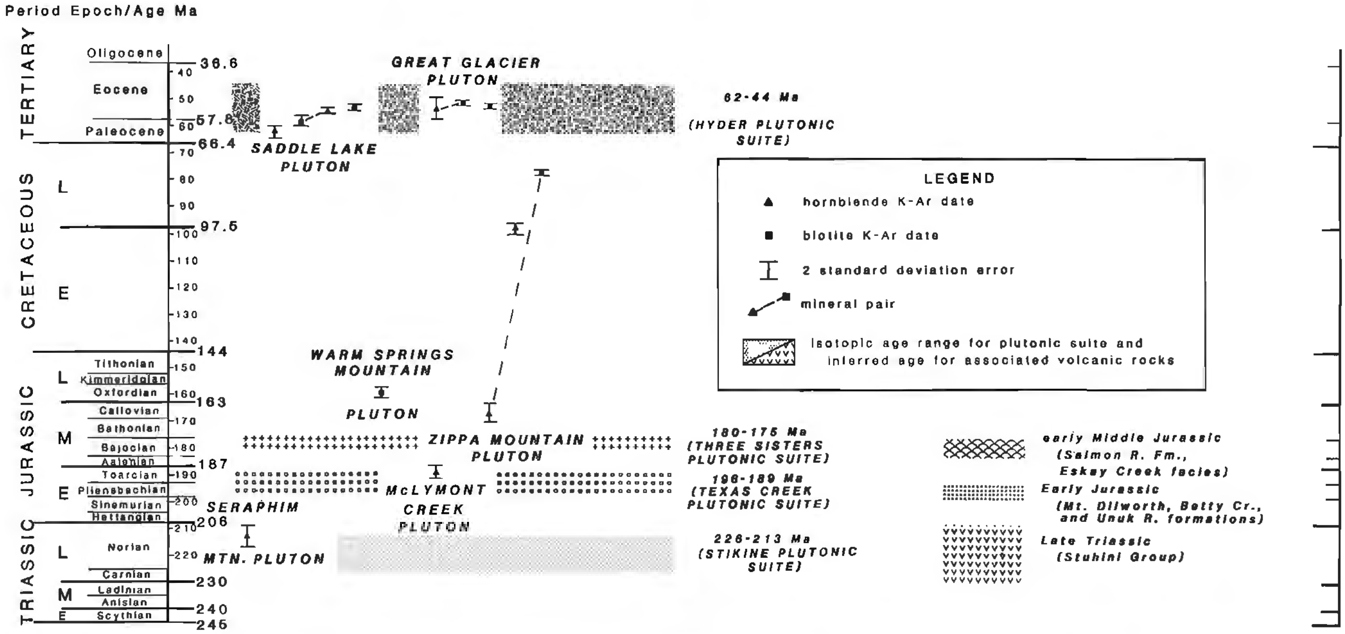


Figure 2. Mesozoic volcanic episodes in Iskut River map area and new K-Ar dates for Mesozoic and Tertiary plutons. Time scale is after Palmer (1983).

Anderson (1989) included the pluton near Middle Mountain with the Texas Creek plutonic suite. However, a preliminary U-Pb zircon date of about 177 Ma suggests that the pluton is also Middle Jurassic.

Hornblende (167 $\pm$ 4 Ma and 98 $\pm$ 2 Ma) and biotite (77 $\pm$ 1 Ma) K-Ar dates for the latest intruded phases of the Zippa Mountain complex are highly discordant but the cause of the discordance is unknown. The hornblende date should be regarded as a minimum estimate of the pluton's age. Well-dated alkaline (e.g. alkali-feldspar-phyric) plutons, phases or dykes in Iskut River and southwestern Telegraph Creek map area are commonly Early Jurassic. A preliminary U-Pb date for the Mitchell Glacier syenite of about 196 Ma (J. Mortensen and R. Kirkham, pers. comm., 1989) suggests the Early Jurassic alkaline magmatic province extends to the Sulphurets mineral camp. Zippa Mountain pluton's alkaline affinity and K-Ar dates suggest it is more likely part of the Early Jurassic Texas Creek plutonic suite than the Middle Jurassic Three Sisters suite.

New K-Ar dates and U-Pb age estimates and other geochronometry define Late Triassic, Early Jurassic and Middle Jurassic plutonic episodes in Iskut River map area. The episodes compare closely with the known or estimated age of Mesozoic volcanism represented by the Upper Triassic Stuhini Group, Lower Jurassic Hazelton Group and Middle Jurassic Salmon River Formation, Eskay Creek facies pillow lavas (Anderson and Thorkelson, 1990; Fig. 2).

Mesozoic K-Ar isotopic systematics for mafic minerals in plutonic rocks seem little affected by the mid-Cretaceous greenschist-grade metamorphic event interpreted from K-Ar whole-rock dates for alteration zones Alldrick et al. (1987). The Eocene thermal overprint, widespread in the Prince Rupert, Terrace and Whitesail Lake map areas to the south (van der Heyden, 1989), seems restricted to country rock near known or suspected Tertiary intrusions in Iskut River map area.

Tertiary plutons

The range of Tertiary dates from Saddle Lake (51-53 Ma) and Great Glacier (53-62 Ma) plutons overlaps or is slightly older than the age range (44-54 Ma) for the Hyder plutonic suite suggested by earlier studies (Smith, 1977; Alldrick et al., 1986, 1987). Dates for the Great Glacier pluton are concordant but the four dates for Saddle Lake pluton are slightly discordant. The K-Ar geochronometry for Saddle Lake pluton and Great Glacier pluton agrees within error with respective, preliminary U-Pb ages for zircon of 51 $\pm$ 1 and 56 $\pm$ 6 Ma.

SUMMARY

Late Triassic (213-226 Ma), Early Jurassic (189-196 Ma), Middle Jurassic (175-180 Ma) and Tertiary (44-62 Ma) plutonism is widespread in Iskut River map area. Mesozoic plutonism accompanied Late Triassic (Stuhini Group), Early Jurassic (Hazelton Group) and Middle Jurassic (Salmon River Formation, Eskay Creek facies) volcanism. The regional greenschist facies metamorphic event (Alldrick et al., 1987) seems to have had minimal effect on hornblende and biotite dates from Mesozoic plutons.

ACKNOWLEDGMENTS

B.C. Geological Survey Branch colleagues Derek Brown and Mike Gunning generously provided access to their unpublished data and stimulating discussions on the evolving geochronometric framework of northwestern British Columbia. Jim Mortensen and Rod Kirkham also provided unpublished data from the Sulphurets area for inclusion in the paper. Meticulous hand-picking of mineral separates by technical staff of the Geological Survey of Canada's Geochronology Laboratory provided the material for K-Ar analysis. Glenn Woodsworth's and Dirk Tempelman-Kluit's reviews of earlier versions of the manuscript helped improve the final version.

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Jurassic correlations in the Iskut River map area, British Columbia, and the age of the Eskay Creek deposit

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Smith, P.L. and Carter, E.S. Jurassic correlations in the Iskut River map area, British Columbia, and the age of the Eskay Creek deposit; in Current Research, Part E, Geological Survey of Canada, Paper 90-1E, p. 149-151, 1990.

Abstract

A high-resolution, Jurassic biochronology based on macro- and microfossils could have important applications in the tectonically complex Iskut map area where Jurassic sediments accumulated in a shallow marine, volcanically active area characterized by rapid lateral variations in facies. Reconnaissance work in the upper drainage of the Unuk River has produced Pliensbachian to possibly Oxfordian age ammonite and radiolarian faunas, some of which are anomalous with respect to currently available geological maps. The presence of Jurassic radiolarians in well-rounded clasts of Jurassic conglomerates may offer the opportunity of constraining periods of Jurassic tectonism.

Upper Pliensbachian ammonites were collected stratigraphically below the Eskay Creek deposit. A limestone clast from a conglomerate above the deposit has yielded radiolarians that indicate a Middle Toarcian to Early Bajocian age.

Résumé

Une biochronologie à la haute résolution du Jurassique, fondée sur des macrofossiles et des microfossiles, peut avoir d'importantes applications dans la région cartographique d'Iskut à tectonique complexe, où des sédiments jurassiques se sont accumulés dans une zone marine peu profonde, soumise à l'activité volcanique et caractérisée par des variations latérales rapides de faciès. Des travaux de reconnaissance dans la partie amont du bassin versant de la rivière Unuk ont permis d'obtenir des faunes d'ammonites et de radiolaires datant du Pliensbachien jusqu'à peut-être l'Oxfordien; certaines de ces faunes ne s'accordent pas avec les cartes géologiques déjà établies. La présence de radiolaires jurassiques dans des fragments bien arrondis de conglomérats jurassiques peut permettre de déterminer les périodes de phases tectoniques survenues au cours du Jurassique.

Les ammonites du Pliensbachien supérieur ont été recueillies dans un lieu situé stratigraphiquement en-dessous du gisement d'Eskay Creek. Un fragment de calcaire provenant d'un conglomérat gisant au-dessus du gisement a fourni des radiolaires qui indiquent un âge se situant entre la Thoarcien moyen au Bajocien inférieur.

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INTRODUCTION

Work on the Jurassic biostratigraphy of western Canada and the United States is beginning to produce an integrated biochronology based on both macro- and microfossils intended for use by regional mappers and for subsurface correlation. Some published results are available for the ammonites (e.g. Smith et al., 1988) and radiolarians (e.g. Carter et al., 1988) and numerous studies by ourselves and others are in progress. A resolution of one or two million years is indicated by the comparison of currently available biochronologic and geochronologic scales (e.g. Harland et al., 1989).

Correlations of Jurassic units in the Iskut map area have been hampered by structural complexity, rapid lateral variations in facies and particularly by the dearth of fossils. The lateral variations in facies result from pyroclastic and epiclastic deposits associated with andesitic stratovolcanoes interfingering with, and ultimately wedging out into marine, fine grained clastics. Preservation may also be influenced by sedimentary environment and the subsequent diagenesis of andesitic material. Preservation tends to be poorest and abundance lowest in proximal, coarser grained facies.

The need to improve biostratigraphic control by more detailed work has been given impetus by the presence of gold and silver mineralization thought to be confined to Lower Jurassic units (Anderson, 1989).

PRELIMINARY OBSERVATIONS

One of us (PLS) undertook three weeks of reconnaissance field work during the summer of 1989, visiting localities in the upper drainage of the Unuk River identified as potentially fossiliferous by various survey and exploration geologists (Fig. 1). We are currently in the early stages of processing our material and bringing together earlier collections but we offer the following preliminary observations.

Faunas appear to fall into three broad categories in terms of depositional setting:
a) Rich, mostly benthic faunas characteristic of local, epiclastic sandstone lenses commonly developed near the top of thick, very coarse epiclastic units. Overlying rocks are often shale and siltstone. Typically the faunas are dominated by the thick-shelled bivalve *Weyla,* trigonid bivalves, coleoids and scleractinian corals (solitary and compound). Pliensbachian and Toarcian ages are indicated. Near-shore trace fossils such as *Thalassinoides* are sporadically present.
b) Locally abundant ammonites in siltstone and mudstone units.
c) Limestone clasts in conglomerate, and rare carbonate concretions that yield radiolarians.

In the folded sequence of rocks exposed from the Calpine Camp in Eskay Creek westwards to Tom Mackay Lake (the Unuk River and Salmon River formations of Grove, 1986), faunal assemblages ranging in age from Pliensbachian to Oxfordian have been identified. Faunas beneath the main Calpine Camp gossan consist of a diverse Lower Jurassic benthonic fauna in the coarse sandstones of Donnelly's unit 1, overlain by ammonite-bearing siltstones of his unit 2 (Donnelly, 1976). The ammonites include *Tiltoniceras propinquum* (Whiteaves), *Protogrammoceras* spp., and *Lioceratoides?* sp. characteristic of the Carlottense Zone, the uppermost zone of the Pliensbachian. Ammonite aptychi were also collected. Breccio-conglomerates to the east of Argillite Creek, stratigraphically above the Calpine Camp gossan, contain well-rounded limestone clasts that yield radiolaria. The joint occurrence of *Elodium* cf. *nadaensis* Carter, *Emiluvia acantha* Carter, *Napora* sp., *Paronaella variabilis* Carter, *Parvicingula* sp., *Perispyridium* spp., *Protoperispyridium hippaensis* Carter, *Pseudocrucella sanfilippoae* (Pessagno), and *Stichocapsa* cf. *convexa* Yao indicates a late Middle Toarcian to Early Bajocian age. Superjacent localities between Argillite Creek and Tom Mackay Lake have yielded the Bathonian ammonite *Iniskinites* (Gunning, 1986) and possible Oxfordian perisphinctid ammonites, indicating a correlation with the Bowser Lake Group.

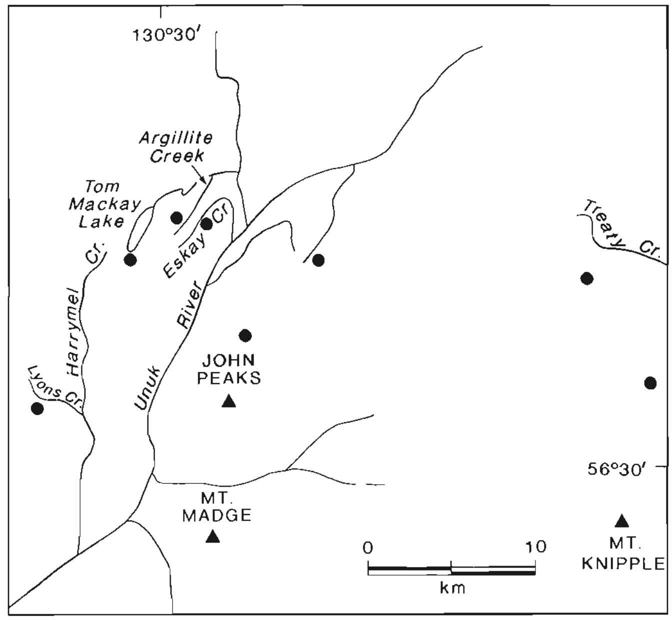

Figure 1. Reconnaissance area, Unuk River valley. Sites visited are shown by dots.

On the west side of the Unuk River in the headwaters of Lyons Creek, a thick shale and siltstone unit rests on silicified limestones and limestone breccias mapped as Triassic Takla Group (Grove, 1986). The limestones are stromatolitic and also contain irregular lenses riddled with trace fossils. Foraminifera and ostracodes from these carbonates have not yet been identified but the overlying shales contain an abundant Upper Pliensbachian ammonite fauna (Fig. 2).

Isolated faunas from localities to the east of the Unuk River valley, as far east as Treaty Creek are still being studied. Local rich benthic faunas are present, as are rare ammonite localities yielding Lower Jurassic forms such as *Dactylioceras.*

FUTURE WORK

Apart from completing the processing of material collected in 1989, we also hope to examine fossil material collected

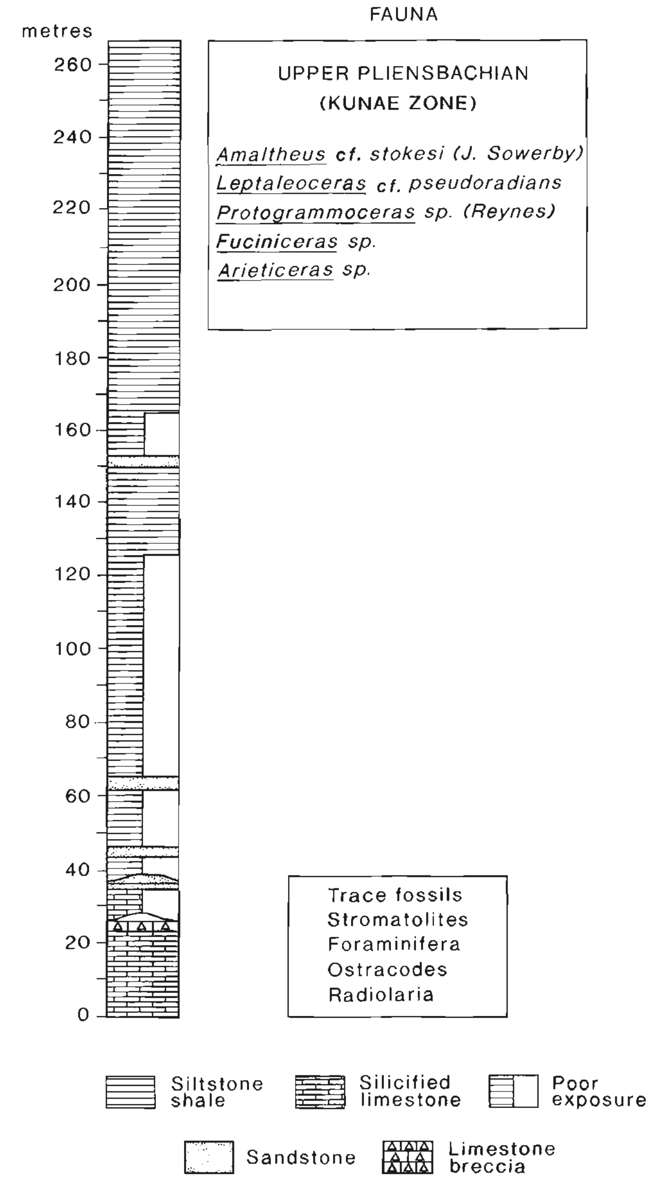

Figure 2. Stratigraphic Section, Lyons Creek.

by other workers. We are particularly anxious to gain access to microfossil material retrieved from well cores in what are suspected to be Jurassic units. We also plan a systematic study of conglomerate clasts in an attempt to identify the age and perhaps provenance of carbonate clasts. In coming field seasons the work will be expanded westward into the Iskut River valley and eventually north and northwestward into the Telegraph Creek and Tulsequah map areas.

ACKNOWLEDGMENTS

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Quaternary volcanic rocks of the Iskut River region, northwestern British Columbia

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Stasiuk, M.V. and Russell, J.K., Quaternary volcanic rocks of the Iskut River region, northwestern British Columbia; in Current Research, Part E, Geological Survey of Canada, Paper 90-1E, p. 153-157, 1990.

Abstract

Along and south of the Iskut River, Quaternary volcanic rocks occur as scattered monogenetic cinder cones and lava flows. These volcanic centres comprise the Iskut River volcanic rocks and include localities at: Iskut Canyon, Cone Glacier, Cinder Mountain, King Creek, Second Canyon Creek, and Lava Fork Creek. All the Iskut River volcanic rocks are basaltic. Representative samples for petrographic and chemical analysis were collected from these six volcanic centres.

Résumé

Le long de la rivière Iskut et au sud de celle-ci, on trouve des roches volcaniques d'origine quaternaire disséminées sous forme de cônes de scories et de coulées de lave monogéniques. Ces centres volcaniques sont constitués de roches volcaniques de la rivière Iskut et se manifestent aux endroits suivants : le canyon d'Iskut, le glacier de Cone, le mont Cinder, le ruisseau King, le ruisseau Second Canyon et le ruisseau Lava Fork. Toutes les roches volcaniques de la rivière Iskut sont basaltiques. Des échantillons représentatifs requis pour une analyse pétrographique et chimique ultérieure ont été prélevés dans ces six centres volcaniques.

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INTRODUCTION

The scattered collection of monogenetic cinder cones and lava flows which occur along and south of the Iskut River (NTS 104B, Fig. 1, 2) are described informally as the Iskut River volcanic rocks (Grove, 1974; Souther, in press a). The volcanic rocks are basaltic and were erupted between 8780 and 360 years BP (Elliott et al., 1981; Souther, in press a). The Iskut River volcanic assemblage includes at least six volcanic centres comprising lava flows and/or cinder cones. Iskut Canyon, Cone Glacier, Cinder Mountain, King Creek, Second Canyon Creek, and Lava Fork Creek centres were investigated during the 1989 field season. Figure 1 is a general geological map for northern British Columbia illustrating the distribution of other prominent Quaternary volcanic centres which are proximal to the Iskut River region, including Hoodoo Mountain (e.g., Souther, in press b), Mount Edziza (e.g., Souther and Hickson, 1984), and Level Mountain.

The 1989 field season represented a brief reconnaissance survey of the Iskut River volcanic centres. The main objectives were to visit as many centres as feasible, to make preliminary field measurements and to collect representative samples of the eruption products at each locality. The number of samples collected at each site reflects the quality (freshness) of the volcanic rocks exposed, the textural and mineralogical variability of eruptive units at each locality, and the volume of material erupted. Every effort was made to thoroughly sample all volcanic rock types and to ascertain the relative timings between distinct eruptive events. Samples were collected at each locality shown on Figure 2, except for Julian Lake.

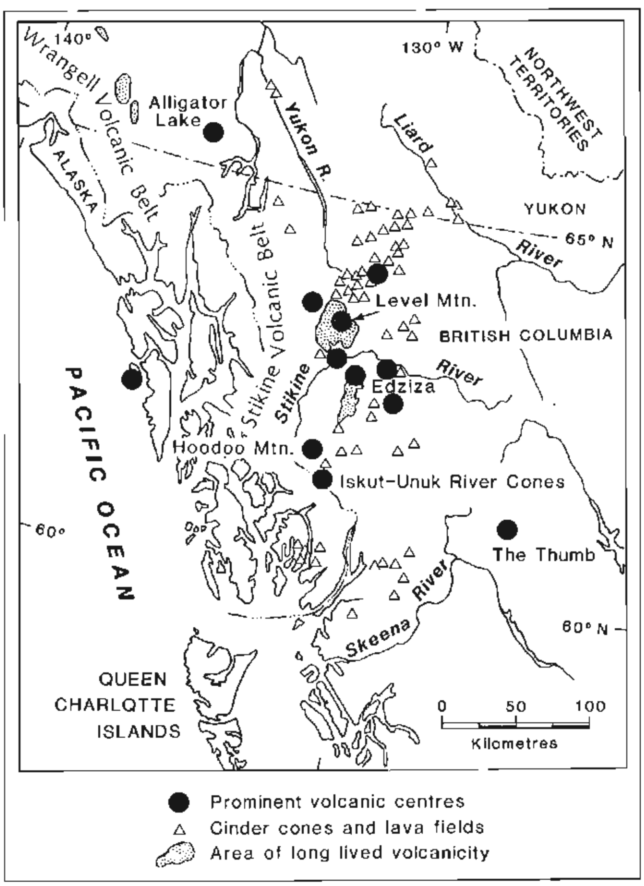

Figure 1. Regional geological map showing the distribution of Quaternary volcanic centres proximal to the Iskut River region. Other prominent centres include: ED-Mount Edziza, SR-Spectrum Range, HO-Hoodoo Mountain volcano, and LV-Level Mountain. Modified from Souther (1977).

Where samples were collected, sufficient material was taken to provide sample splits for J. Nicholls (University of Calgary). Over the next year the samples will be petrographically and chemically analyzed. The chemical and mineral chemical data can be used to clarify the tectonic affinity and mantle source region for the Iskut region magmas (e.g. Erdman, 1985; Bevier, 1989), evaluate the possible chemical relationships between each volcanic centre, and constrain the differentiation history of each volcanic centre.

SUMMARY OF FIELD OBSERVATIONS

Iskut Canyon

The Iskut Canyon cones and flows are located along the Iskut River, near the junction with Forrest Kerr Creek. They have been mapped by Grove (1974, 1986), Anderson (1989), and Read et al. (1989). The major lava flows erupted from a vent at about the 760 m level which is marked by the crater-like remnants of a cinder cone. The cone is approximately 50 m in diameter. The lava flows are exposed in the banks of the Iskut River and form canyon walls for approximately 15 km downstream and west of the vent area. Commonly, the canyon walls contain up to three separate flows. The lavas are well jointed and average 10 m in thickness. Blocky lava flow covers the flat valley floor as far as 7 km west of the vent. Lava blocks (0.3-1.0 m) on the flow tops form hummocky ridges which are arcuate-shaped and normal to the direction of lava flow. The ridges have a relief of about 3 m and a wavelength of 10 m.

Near the vent-area the lava is smooth surfaced and highly vesicular. The actual vent area is heavily vegetated due to overlying pyroclastic material (cinder and bombs) derived from the cinder cone. The single lava flow exposed in a downstream section of the Iskut River canyon wall is sparsely porphyritic with 1-5 mm diameter phenocrysts of olivine, plagioclase and pyroxene, and common 1 cm diameter intrusive xenoliths.

There are other possible vents exposed at the Iskut Canyon locality. Read et al. (1989) mapped one other vent, 2 km northwest of the main vent. There is another minor volcanic vent on the south bank of the Iskut River. These three vents appear to be aligned and may define a northwest-trending rift or fracture zone. Samples collected at the latter vent are lithologically similar to the main vent samples, suggesting that the vents were fed by the same magma and probably erupted contemporaneously.

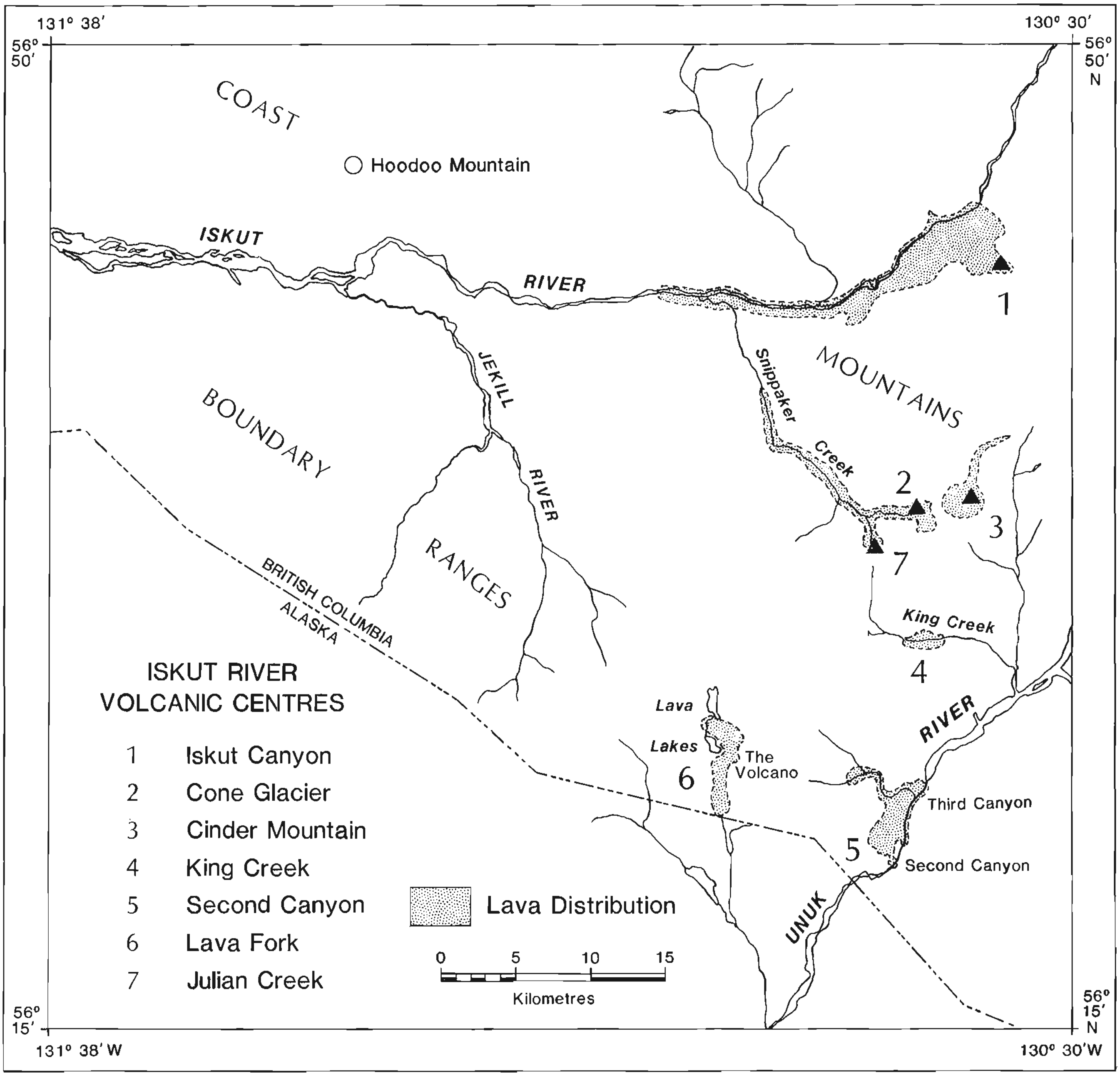

Figure 2. Location diagram for the Iskut River volcanic centres, including: Iskut Canyon, Cinder Mountain, Cone Glacier, Julian Lake, King Creek, Second Canyon, and Lava Fork (after Souther, in press a).

Cone Glacier

Near the nose of Cone Glacier two well-dissected cinder cones are well exposed with little or no vegetative cover, whereas the distal portions of the associated flows are commonly covered with vegetation. Souther (in press a) recognized a vent area west of Cone Glacier: a rugged volcanic edifice comprising pyroclastic deposits and lava flows. A second volcanic vent is located north and east of the glacier's toe. Both vents erupted basalt lavas; however the lavas are distinct lithologically and texturally.

The western cone has, as its lowest stratigraphic unit, pillowed lava flows. The flow material outcrops in cliffs which form the lower part of the cone's northern flank. In outcrop the pillow lavas are radially jointed and have pronounced glassy rinds and concentric vesicular zones. Phenocrysts of olivine and plagioclase are sparse and less than 5 mm across. The package of pillowed flow has a high thickness to length ratio which may be the result of lava erupting in contact with ice (Grove, 1974). Above the pillowed flow is a unit of brown-gold coloured volcanic breccia containing clasts of black fragmental basaltic glass, identifiable from the vesicular zones as pieces of pillowed lava. The matrix appears to be devitrified basaltic glass, suggesting that this unit is a hyaloclastite breccia. Above this unit are well bedded pyroclastic rocks which dip towards the surrounding valleys away from the proposed vent. Both units appear to be subaerial deposits.

North and northeast of the western cone is a series of four or more basalt lava flows. These basalt lavas overlie the pillow lava units described above. In places the flows have smooth, glacially striated upper surfaces, suggesting that the eruptions preceded or overlapped a (local?) period of glacial retreat. The flows are each 0.5-3 m thick, with scoriaceous and brecciated bottoms and tops. They contain megacrysts of euhedral, glassy feldspar up to 2.5 cm in length, phenocrysts of tawny-coloured olivine (2-3 mm in diameter) and less abundant crystals of vitreous, black orthopyroxene up to 5 mm in diameter.

The series of flows do not originate from the western cinder cone. They were traced toward Cone Glacier into glacial till, and then eroded blocks were followed uphill to the east. Well bedded pyroclastic units interbedded with flows are graphically displayed in a steep creek gully and dip away from the proposed eastern vent area. The flow material appears similar to the valley bottom flows; all the extruded material clearly flowed down a steep cinder cone flank.

Lava Fork

The Lava Fork lavas are the youngest volcanic rocks of the Iskut River region and may represent the youngest volcanic eruptions in Canada (Elliott et al., 1981; Souther, in press a). Age dates for the Lava Fork lavas (^{14}C) are as young as 360 ± 60 years BP (Souther, in press a). The Lava Fork volcanic rocks are spectacularly exposed and the lava field displays a wide variety of volcanological features including lava tubes and spatter vents.

The main lava flow appears to have erupted from a vent located on a ridge of crystalline basement rocks at an elevation of 1400 m. The vent area is delimited by a thick blanket of black, glassy pyroclastic tephra which mantles bedrock and glaciers to the north, east and south.

The main lava flows appear to have been erupted in a single period and were probably fed by lava fountains; the top surfaces of the lavas are occasionally covered with spatter agglutinate. The lava fountaining filled the cone with ponded lava, which eventually broke out, flowed across the ridge top, over a knob-like shoulder and down the steep valley wall to flood the valley floor below. This interpretation implies that the spatter bombs on the flanks of the cinder cone are part of the same eruptive activity that produced the lava flows.

The main lava flow lies within the Lava Fork valley and, by damming the Lava Fork Creek, formed the two Lava Lakes. The southern limb of the lava flow extends south along the valley bottom for at least 20 km into Alaska. Generally, the flow surface is scoriaceous and may have a thin veneer of clinkery material. Pahoehoe, ropy surfaces are common, and in places the lava flow has developed numerous, prominent lava channels with prominent levees (north fork of the lava at about 1040 m). The lava rivers commonly develop into lava tubes which subsequently have collapsed. In several localities, the lava tubes remain intact and are as high as 3 m and extend for 30-40 m.

Lavas near the vent area contain leucocratic, porous xenoliths, presumably derived from the underlying crystalline crustal rocks. The xenoliths are common and as large as 10 cm across. Basement rocks in the vicinity include foliated quartz granodiorite with biotite-quartz schist inclusions, quartz monzonite, and biotite-quartz schist. Assimilation may be an important process contributing to the chemical diversity of these lavas as the siliceous xenoliths are partially fused. Large xenoliths occur only in the lavas which are proximal to the vent area. In the lavas that are farther from the vent area, xenoliths are rarer and occur as smeared fragments 1 cm or less in diameter. Two hypotheses to explain this observation are that: i) there was a significant fluctuation in the amount of xenolithic material being entrained by the magma during eruption, or ii) the reduction in frequency and size of xenoliths is the result of assimilation occurring as the lavas flowed away from the vent-area.

Cinder Mountain, King Creek, Second Canyon Creek

Souther (in press a) proposed a vent area in the snowfields east of the Cone Glacier volcanic rocks, visible only as outcrops of limited extent protruding above the snow. One of the exposed outcrops comprises a succession of lava flows and pyroclastic deposits which dip gently south. The lowest unit is hyaloclastite breccia which envelopes at least two scoriaceous, well-jointed basalt dykes (lavas?) of irregular thickness. A third basalt flow occurs higher in the section.

In the King Creek area, a 50 m thick, valley-bottom exposure of pillowed lava is overlain by volcanic breccia. Overlying the pillow lava and exposed to an elevation of about 1070 m (the proposed vent), is volcanic breccia. Bedding in the breccia dips into King Creek valley.

Volcanic rocks in the Second Canyon Creek volcanic rocks comprises a single blocky lava flow, through which metamorphic basement rocks protrude, and a number of dissected cinder cones (Souther, in press a).

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Eastern margin of the Central Gneiss Complex in the Shames River area, Terrace, British Columbia

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Abstract

The eastern margin of the Central Gneiss Complex (CGC) west of Terrace has been interpreted as a high angle fault, an east-verging thrust, and a denudational fault. Along Shames River, deformed granitoid rocks of the CGC are in steep fault contact with volcanic and sedimentary strata of Stikinia. Extensional ductile movement in the CGC was directed northeast near Shames River, and northeast and southwest near Exstew River. East of Shames River, earlier northeast-vergent thrusting gave rise to an inverted metamorphic sequence of amphibolite over greenschist facies volcanic and plutonic rocks. This region thus records a change from compressional to extensional tectonics, and shows similarities to metamorphic core complexes elsewhere.

Résumé

La marge orientale du complexe gneissique central (CGC), située à l'ouest de Terrace, a été interprétée comme une faille à fort pendage, une faille inverse à vergence est et une faille de dénudation. Le long de la rivière Shames, des roches granitoïdes déformées du CGC sont en contact par faille à fort pendage avec des roches volcaniques et sédimentaires de Stikinia. Un mouvement ductile d'extension dans le CGC avait une direction nord-est près de la rivière Shames, et nord-est et sud-ouest près de la rivière Exstew. À l'est de la rivière Shames, de vieux charriages à vergence nord-est ont donné naissance à une séquence métamorphique renversée d'amphibolites sur des roches volcaniques et plutoniques à faciès schiste vert. Cette région fournit la preuve d'un changement d'une tectonique de compression à une tectonique d'extension et montre des similarités avec des complexes à noyau métamorphique ailleurs.

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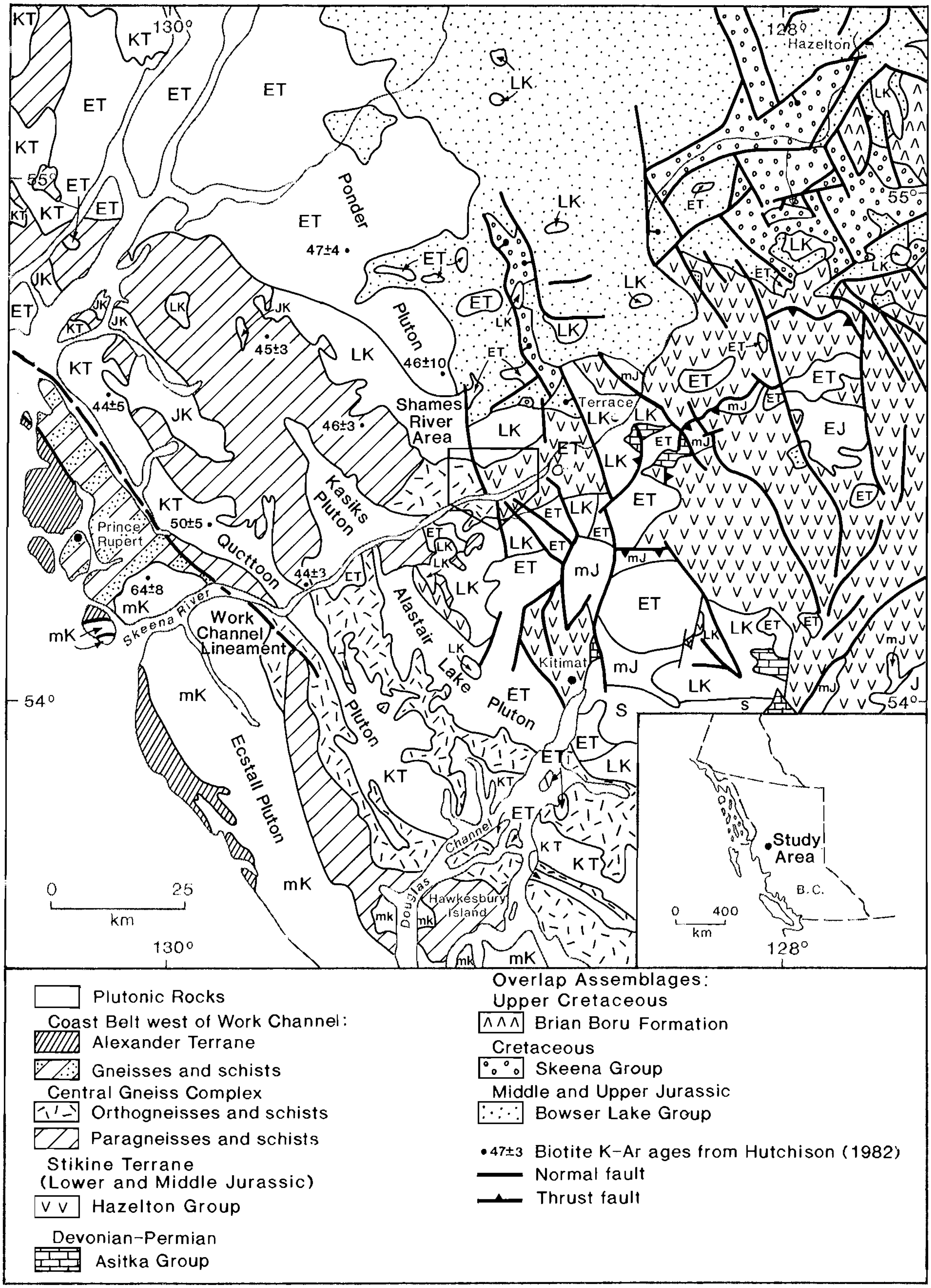


Figure 1. Regional geology and location of study area (modified after Wheeler and McFeely, 1987; Gareau, 1989, and present mapping). Plutonic rock ages: ET, Early Tertiary; KT, Cretaceous-Tertiary; LK, Late Cretaceous; mK, mid-Cretaceous; JK, Jurassic-Cretaceous; mJ, mid-Jurassic; J, Jurassic; EJ, Early Jurassic; S, Silurian.

INTRODUCTION

In the Prince Rupert-Terrace region, the core of the Coast Plutonic Complex is composed of amphibolite- to granulite-facies rocks of the Central Gneiss Complex (CGC). The CGC is separated from kyanite zone rocks to the west by the Work Channel lineament (Fig. 1), along which the CGC was rapidly uplifted during 62-48 Ma (Crawford and Hollister, 1982; Hollister, 1982; Selverstone and Hollister, 1980). Concordant U-Pb zircon dates of 60-57.1 Ma were obtained by Armstrong and Runkle (1979) and Gareau (1989) for Quottoon pluton, emplaced during amphibolite- to granulite-facies metamorphism (Kenah and Hollister, 1982). K-Ar cooling dates of 50-43 Ma obtained from the CGC and associated plutons by Hutchison (1982) and Harrison et al. (1978) are consistent with denudation during the Eocene. Protoliths of the CGC are arc-derived plutonic, volcanic and sedimentary rocks of Paleozoic and Mesozoic age (Hutchison, 1982; Gareau, 1988, 1989; Hill, 1984, 1985; van der Heyden, 1982, 1989; Barker and Arth, 1984). Hill et al. (1985) suggested that the supracrustal rocks represent a highly metamorphosed portion of Stikinia.

The eastern contact of the CGC with low grade volcanic and sedimentary rocks of Stikine Terrane is largely engulfed and obliterated by Late Cretaceous and Eocene plutons, including the Alastair Lake and Ponder bodies (Woodsworth et al., 1985; Hutchison, 1982; Sisson, 1985). East of this boundary, Late Mesozoic to Paleogene structures are north- to east-verging thrust faults involving rocks of the Late Triassic to Early Jurassic Kitselas volcanics (Woodsworth et al., 1985). Tertiary extensional structures, such as block faulting along the Terrace-Kitimat graben, have been recognized by Woodsworth et al. (1985).

The nature of the eastern boundary of the CGC west and south of Terrace is controversial, and has been variously described as a high angle fault (Woodsworth et al., 1985); an east verging thrust fault (Crawford et al., 1987; van der Heyden, 1982) and a low angle denudational fault (van der Heyden, 1989). Mapping at 1:20 000 scale was begun in 1989 in the Shames River area west of Terrace (Fig. 1) to determine the nature of the eastern boundary of the Central Gneiss Complex there. Laboratory studies will document the timing and nature of deformation and metamorphism. A further objective will be to compare and contrast the protoliths of the CGC with rocks in Stikinia.

LITHOLOGIES

The Shames River area is underlain by two diverse terranes separated by the northwest-trending, high angle Shames River fault (SRF, Fig. 2). To the east, granitic and lower amphibolite facies metamorphosed volcanic rocks (unit 1) are thrust northwards over greenschist facies volcanic and sedimentary rocks (units 2-4), forming an inverted metamorphic sequence. West of the SRF, gneissic granitoid rocks (units 5-9) underlie moderately deformed to undeformed granitoid rocks (units 10, 11) along a thick mylonitic zone (Fig. 3).

Deformed metavolcanic and granitoid rocks comprise the imbricate package of unit 1. Textures indicate that the volcanic rocks were originally andesitic to rhyolitic flows and pyroclastic rocks. Most of the volcanic rocks are dark green, dense, and comprosed of fine grained mixtures of hornblende, chlorite, epidote, quartz and feldspar. Feldspar phenocrysts in flows and lithic fragments in pyroclastic rocks are stretched parallel to well developed, predominantly southeast-dipping, in part mylonitic, foliations. Granitoid rocks show brittle to ductile shearing, and are highly chloritized and epidotized quartz diorite and hornblende diorite containing infrequent mylonitic fabrics. The layered and granitoid rocks are rarely folded into north northeast-vergent folds. Contacts between units are brittle shear zones.

Unit 2 consists of medium- to coarse- grained, white and grey, crystalline and argillaceous marble. Bedding has been transposed by at least two periods of tight to isoclinal folding. Pink, ovoid fusulinids(?) up to 4 mm long were observed at one outcrop. Similar rocks to the east are Early Permian (Woodsworth et al., 1985). In places, thin layers of medium and dark green, phyllitic volcanic rocks are present. Argillaceous layers contain chlorite and epidote. This unit is disconformably overlain by andesitic volcanic rocks of unit 3.

Unit 3 consists of dark green-grey andesitic flows and volcaniclastic rocks. Vesicular, porphyritic and amygdaloidal textures are common. In places, subangular fragments interpreted to be either pillows or bombs up to 4 cm across are enclosed in a dark green phyllitic matrix of chlorite-quartz-feldspar-epidote. The fragments are commonly elongate and plunge gently northeast, parallel to actinolite lineations.

Near Delta Creek, unit 3 is conformably overlain by dacitic to rhyolitic flows, tuffs and breccias of unit 4. Siliceous clasts and phenocrysts are elongate and plunge gently northeast. East of Amesbury Creek, pyroclastic textures are well preserved. There, angular breccia fragments from 2 mm to 5 cm across are porphyritic andesite to rhyolite set in a dacitic tuff matrix. Bedding dips 30-45° northeast.

Units 3 and 4 are tentatively correlated with the lithologically similar Lower Jurassic Hazelton Group (Woodsworth et al., 1985; Tipper and Richards, 1976).

West of the SRF, the rocks are higher grade, moderately to highly deformed, plutonic and minor metasedimentary units (units 5-9) intruded by pre- to post-kinematic granite and granodiorite (units 10, 11).

Unit 5 consists of gneissic biotite and hornblende-biotite granodiorite and tonalite sheets that are cut by abundant ductile shears (Fig. 4). Elongate schlieren of fine grained biotite-hornblende-feldspar schists are common parallel to foliations. Gneissic layering is defined by leucosomes of tonalite interlayered with hornblende- and biotite-rich melanosomes. Rusty metasedimentary and amphibolite layers are rare. The layered and deformed plutonic rocks are cut by hypidiomorphic granodiorite or tonalite sheets, felsic veins, and pegmatites.

Unit 6 is medium- to coarse- grained biotite-garnet granite, for which a concordant U-Pb date of 83 Ma was obtained by Woodsworth et al. (1983). Garnet occurs as phenocrysts up to 1 mm across, and molybdenite occurs as rare flakes

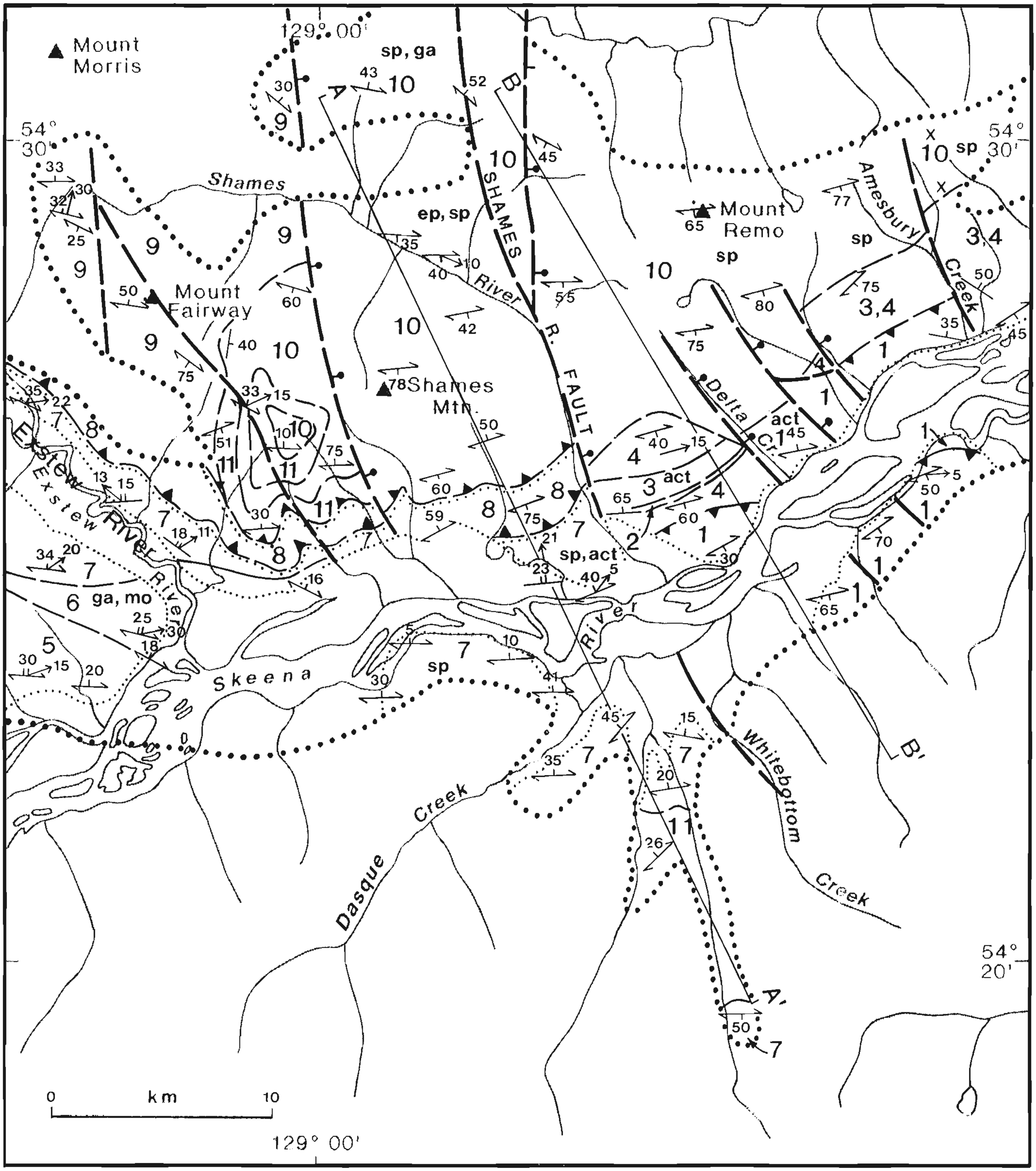
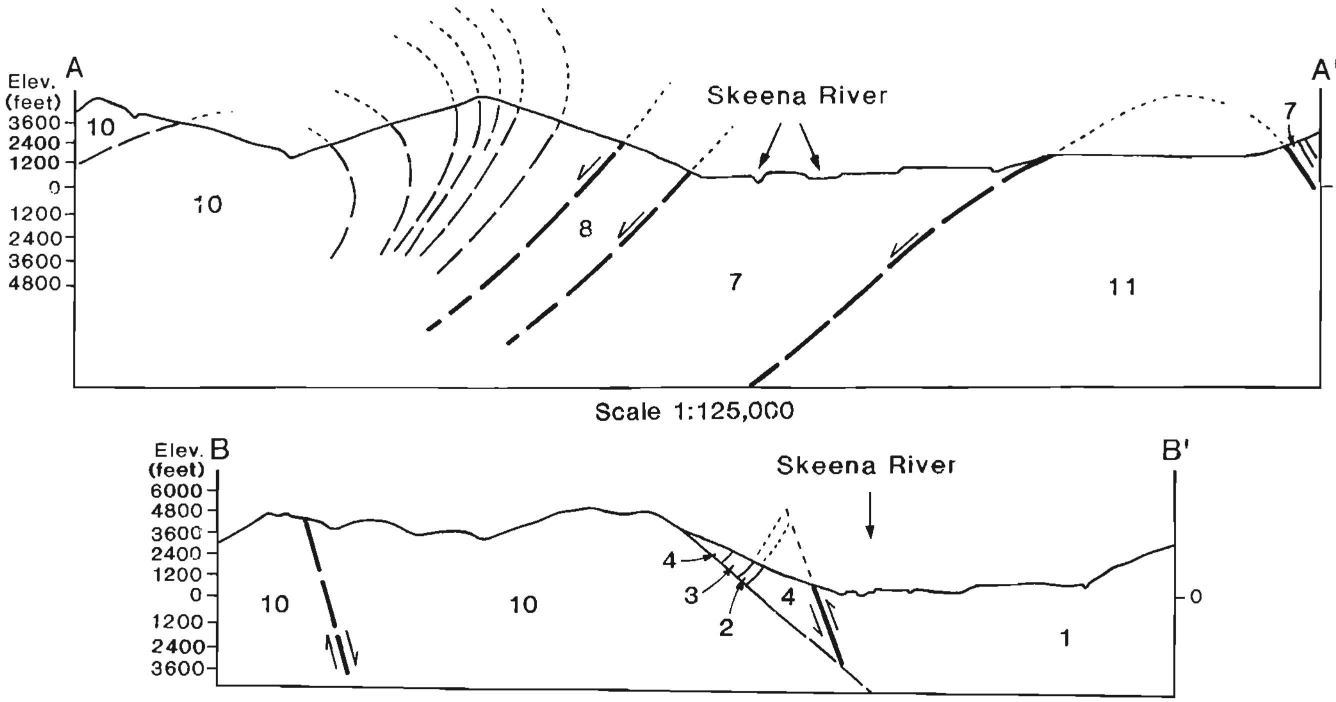


Figure 2. a) Generalized geology of the Shames River area. b) Cross-sections A-A' and B-B'.



LEGEND

Quaternary

Q — Unconsolidated alluvium and slope deposits.

PLUTONIC ROCKS

Undeformed to moderately deformed:

11 — Late- to post-kinematic granite and granodiorite. Hornblende and biotite varieties.

10 — Pre- to syn-kinematic granodiorite with sphene. Hornblende > biotite; rare magmatic(?) epidote west of Shames River fault.

Strongly deformed:

9 — Pre- to syn-kinematic granodiorite. Biotite > hornblende.

8 — Hornblende-biotite-quartz-feldspar schist and gneiss. Minor metasedimentary layers.

7 — Banded mylonitic granodiorite. Biotite > hornblende. Minor metasedimentary rocks.

6 — Granite with biotite and garnet.

5 — Gneissic biotite and hornblende-biotite granodiorite and tonalite sheets.

METAMORPHOSED VOLCANIC, PLUTONIC AND SEDIMENTARY ROCKS

Lower Jurassic Hazelton Group (?):

4 — Dacite to rhyolite flows and pyroclastics.

3 — Andesite flows and pyroclastics.

Lower Permian (?):

2 — Marble with thin argillaceous layers.

1 — Amphibolite facies volcanic-plutonic rocks.

Symbols:

- Foliation (inclined, absent)
- Lineation
- Bedding
- Axial plane and fold axis
- Contact (observed, inferred, gradational)
- Normal fault (barb on upper plate)
- Ductile normal fault
- Thrust fault
- Limit of mapping

Abbreviations:

act	actinolite
ep	epidote
ga	garnet
mo	molybdenite
sp	sphene

Figure 2b

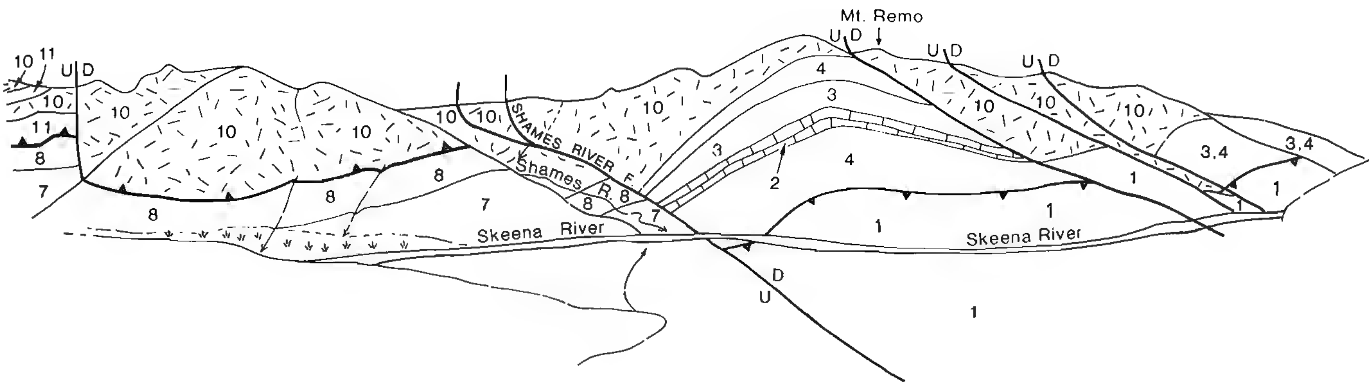


Figure 3. Sketch of Shames River area, looking north, showing Shames River fault separating imbricate package of volcanic-plutonic rocks to the east from deformed granitoid rocks to the west.

Figure 4. Ductilely sheared layers in gneissic granodiorite of unit 5, indicating upper plate to the northeast (right) sense of shear.

up to 3 mm across. Textures grade from hypidiomorphic to strongly foliated. The rocks are cut by abundant ductile normal shears (Fig. 5), many of which are invaded by moderately foliated tonalite, indicating ductile shearing at high temperatures. Toward the upper and lower margins of this unit, grain size decreases and cataclastic textures indicate ductile shearing.

The distinctive rocks of unit 7 are best exposed in road cuts on Highway 16, about 1 km west of Shames River. The unit consists of banded, strongly lineated mylonitized biotite > hornblende granodiorite (Fig. 6) and subordinate metasedimentary rocks. Metasedimentary units are biotite-quartz-feldspar + garnet + epidote schists and gneisses, and garnet-diopside-calcite gneiss.

Biotite granodiorite is medium- to coarse- grained, and may contain hornblende, sphene and pyrite. It is cut by concordant and discordant, dark grey to black, mylonitic bands consisting of broken feldspars up to 4 mm set in a fine grained, granular matrix. In places, mylonitic bands completely engulf sheared granodiorite boudins (Fig. 6). Other dark grey layers cut and are themselves cut by granodiorite, and are interpreted as deformed syn-plutonic mafic dykes.

Figure 5. Steep ductile normal shears indicating sinistral movement in coarse biotite-garnet granite of unit 6. Left side is southwest.

These layers commonly occur as detached slivers of biotite-hornblende-feldspar schist within the granodiorite. Elongate, mafic schlieren may represent metamorphosed, deformed xenoliths. Some of these are altered to epidote-actinolite-chlorite-feldspar-quartz masses. The unit is cut by

Figure 6. Banded mylonitic granodiorite of unit 7 showing darker, finer grained mylonite bands wrapping around lighter, coarse grained granodiorite. Upper plate to the northeast (right) movement is indicated.

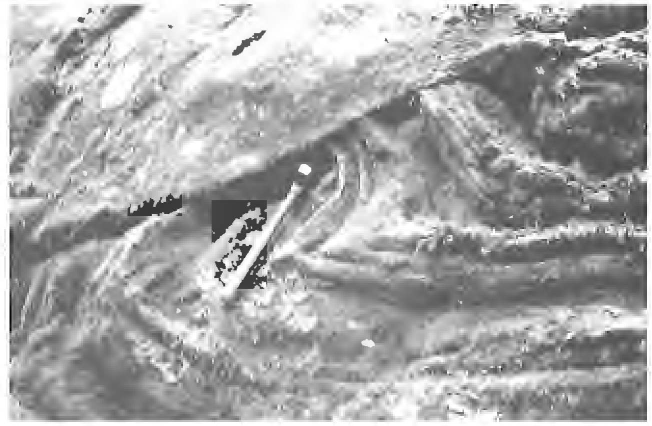


Figure 7. Brittle-ductile shear between granite of unit 11 (above) and folded mylonitic hornblende-biotite schist of unit 8 (below). Upper plate to the northeast (right) movement indicated by fold vergence of the lower plate.

unfoliated to faintly foliated, biotite-muscovite granite dykes and sills, mylonitized hornblende gabbro dykes, and late, unfoliated, biotite lamprophyre dykes trending 330-350°.

Unit 7 grades upwards with decreasing grain size into well foliated biotite-hornblende-quartz-feldspar schist and gneiss of unit 8. Biotite-hornblende-quartz-feldspar schists are in shallow, brittle-fault contact with faintly foliated granite of unit 11 on ridges 1-3 km east of Exstew River (Fig. 7). Tight S-folds in schists underlying granite plunge gently north, and indicate movement of the upper block to the northeast along gently plunging slickensides. This shear sense is consistent with that obtained from shear bands observed below the fault, which may record the last stages of northeast-directed movement of the upper plate.

Unit 9, best exposed on the ridges south of Mt. Morris, is characterized by grey to rusty-beige-weathering, coarse grained, mylonitized biotite granodiorite with or without hornblende and sphene. Screens of gneissic and schistose material are common, many being isoclinally folded about gently northeast- and north-dipping axial surfaces (Fig. 8). The rocks are cut by numerous ductile normal shears, some of which affect unfoliated pegmatite veins.

Unit 10, which forms the ridge tops and peaks both east and west of the Shames River, consists of faintly to unfoliated hornblende > biotite granodiorite with sphene. Minor magmatic(?) epidote is present only in rocks west of the SRF, and may indicate a deeper level of crystallization for those rocks. The unit was previously mapped as belonging to the Ponder pluton (Hutchison, 1982). It commonly has a well developed feldspar and/or hornblende lineation plunging gently northeast. The similarity of lineation of unit 10 to those of underlying deformed units indicates that unit 10 was emplaced during deformation. Mafic enclaves locally form up to 35 % of the rock.

Immediately east of SRF, hornblende-biotite granodiorite of unit 10 intrudes metamorphosed volcanic rocks of unit 4 along a 100-150 m wide agmatite zone containing angular xenoliths up to 4 cm across. The granodiorite is fine grained and foliated parallel to the contact, which dips moderately southeast. Six kilometres northeast of that contact, a 150 m wide mylonite zone, dipping steeply southeast and trending northeast, occurs in the granodiorite. On Shames Mountain, west of SRF, a discrete, 200 m long, 10-15 m wide mylonite zone dipping steeply and variably northwest and southeast occurs within relatively undeformed rocks. There, biotite granite and pegmatite dykes cut mylonite fabrics at high angles but are themselves mylonitized parallel to the fabric of the country rocks. Unit 11, possibly the youngest mappable unit, is best exposed on the ridges east of Exstew River and south of Skeena River. It consists of unfoliated to faintly foliated granite and granodiorite. The rocks are hypidiomorphic, fine- to coarse- grained, and commonly flow-foliated parallel to their intrusive margins. Both hornblende and biotite bearing rocks are present. East of Exstew River, granite occurs as horizontal sheets, containing angular xenoliths of foliated granitoid rocks of units 9 and 10. Fine grained granitic dykes are also found as intrusions in all units mapped. South

Figure 8. Looking west-southwest at S-shaped, isoclinal recumbent folds in granodiorite of unit 9. Mylonitic foliation and later intrusive sheets are folded about gently northwest dipping axial planes. Fold vergence and S-C fabrics indicate upper plate to the southeast (left) sense of shear.

of Skeena River, coarse grained, unfoliated granite (unit 11) is overlain by mylonitic granitoid rocks of unit 7.

Unfoliated lamprophyre dykes containing biotite phenocrysts and basaltic-andesite dykes trend northwest, and dip steeply northeast and southwest. They are fairly common in all units mapped. Hutchison (1982) obtained a whole-rock K-Ar date of 41.1 ± 5.6 Ma from a mafic dyke west of the area. Other lamprophyre dykes in the region are less than about 35 Ma (Harrison et al., 1978; van der Heyden, 1989).

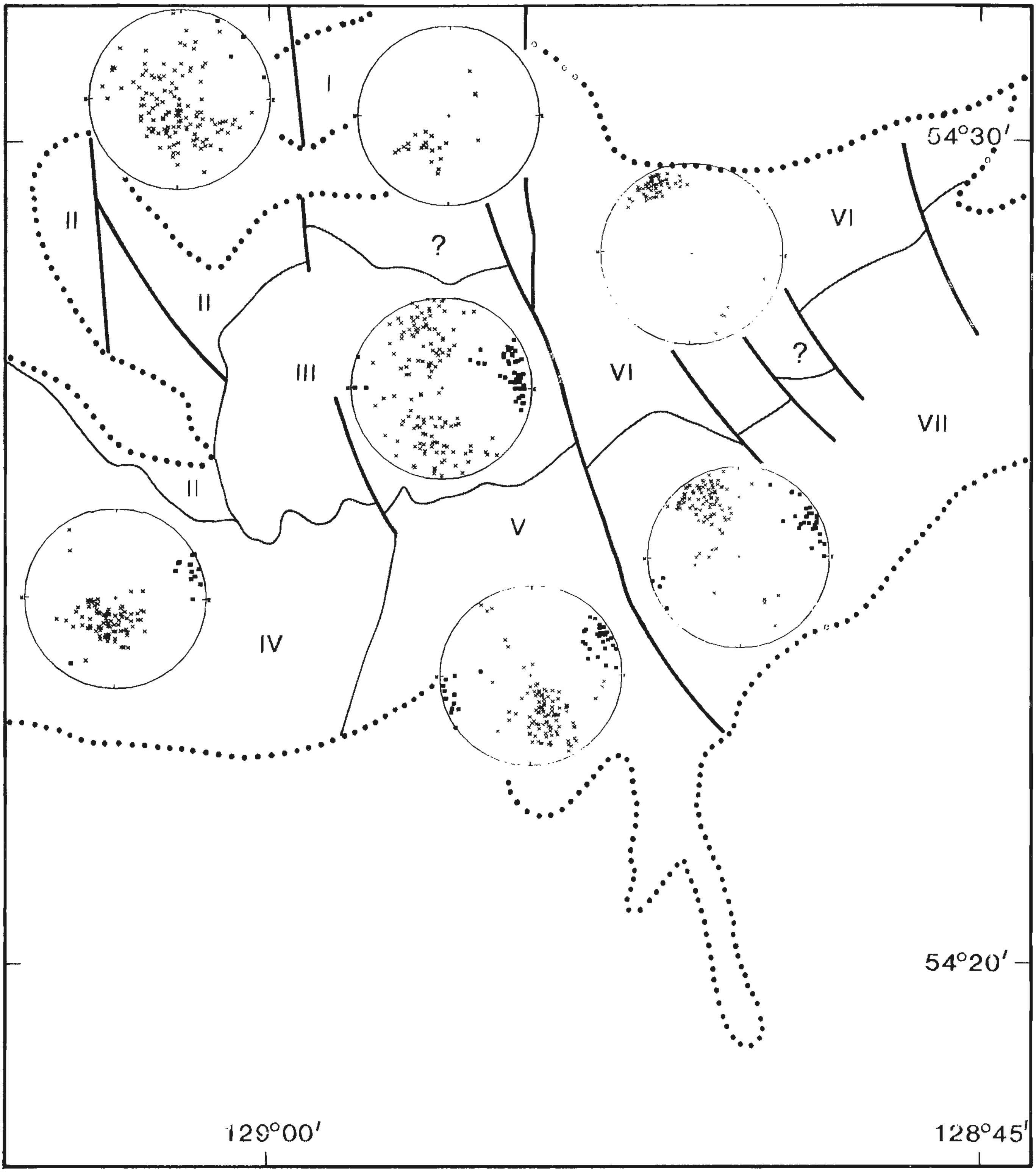


Figure 9. Structural domains of the Shames River area, showing lower hemisphere projections of lineations (squares) and poles to foliation (crosses).

STRUCTURE

The area may be divided into seven structural domains (Fig. 9). Gently northeast- and southwest-plunging mineral and elongation lineations are found throughout the map area. Conspicuous are steeply southeast-dipping foliations east of SRF contrasted with north-dipping fabrics to the west, first noted by Woodsworth et al. (1985).

With few exceptions, ductile shears west of SRF are extensional and dip gently towards the north (Fig. 10). The 83 Ma U-Pb zircon date on ductile sheared garnet-biotite granite west of Exstew River (Woodsworth et al., 1983) and Eocene (46 ± 10 Ma) K-Ar biotite cooling ages for the syn- to post-kinematic Ponder pluton in the north (Hutchison, 1982), brackets ductile deformation between Late Cretaceous and Eocene time.

Upper-plate to the northeast movement along shallow, northeast-plunging lineations is shown by most kinematic indicators east of Exstew River and contrasts with both northeast- and southwest-directed normal movement farther west (Fig. 4, 5). In places, later, southwest-directed ductile shearing cuts earlier mylonite bands and late pegmatite veins. Ductile shears are commonly invaded by moderately foliated, syn-kinematic granite, indicating high temperature shearing. X:Z ratios of cigar-shaped inclusions range from 8:1 to 20:1 in the most deformed units.

At one locality, moderately foliated, late granites of unit 11 are in brittle shear contact with schists of unit 8 (Fig. 7). Because unit 11 clearly cuts mylonitic fabrics elsewhere, this shearing occurred after the main ductile shearing event, during continued deformation.

Conjugate, extensional ductile shears are common in deformed granitoid rocks west of the SRF. In domain III, inter-shear angles are approximately 75-85°, increasing to 90-110° in the underlying domains II and IV. The maximum principal stress orientations (σ_1) inferred from these conjugate ductile shears are vertical; the least principal stress (σ_3) is parallel to lineations plunging gently northeast.

West of the SRF, axial planes are concordant with foliation and dip gently to moderately northwards (Fig. 11). Hinge lines plunge gently northeast and southwest; a few plunge gently north and northwest and show a tendency to be dragged into parallelism with northeast-plunging lineations during ductile shearing. In places, mylonitic foliation is isoclinally folded (Fig. 8). Up to two phases of folding have been recognized in granitoid rocks west of SRF. Rootless, isoclinal hook-shaped folds are common in unit 7. In places up to 50 % shortening has taken place in folded granite dykes.

Hornblende granodiorite of unit 10 west of Shames River forms an asymmetric antiform. Foliations dip gently south in the core, steepening, overturning and becoming more mylonitic southward, where they are dragged into parallelism with mylonitic foliation in unit 8. Poles to foliation describe a north-south trending, steeply west-dipping girdle. Measured fold axes are parallel to lineations, and coincide with the pole to that girdle. Earlier conjugate fractures, inferred to have formed during emplacement of the granodiorite, indicate a north-south trending, subhorizontal orientation for the principal stress. These observations indicate that this body was emplaced as an antiform which was later

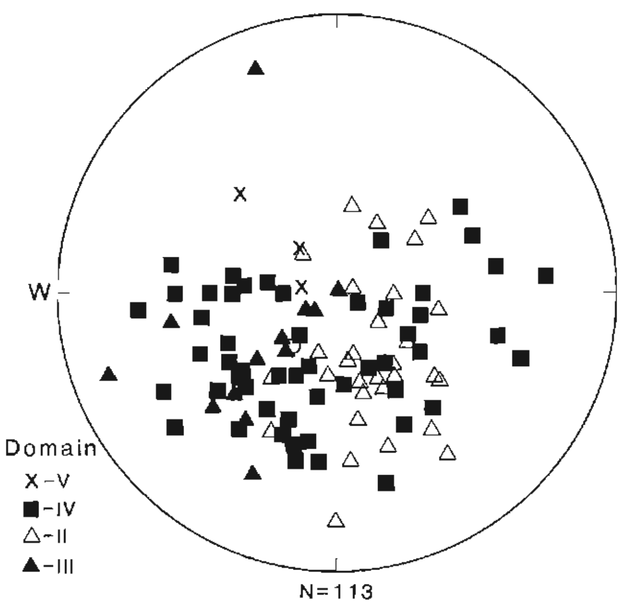


Figure 10. Lower hemisphere plot of poles to ductile shears in deformed granitoid rocks west of Shames River fault.

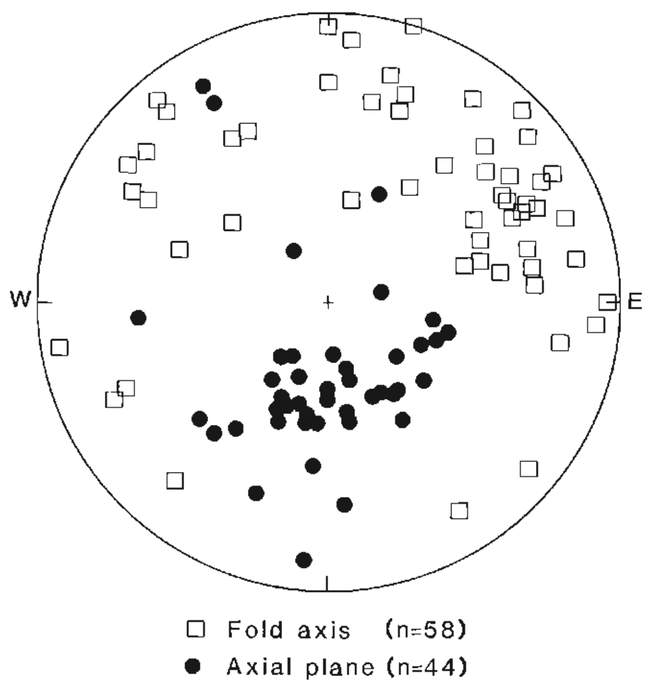


Figure 11. Lower hemisphere plot of fold axes and poles to axial planes west of Shames River fault.

modified by northeast-directed ductile shearing. South-vergent, recumbent isoclinal mesoscopic folds are observed in granodiorites of unit 9 (Fig. 8), which underlies this unit along a moderately east-dipping, ductilely sheared, intrusive zone.

East of the SRF, amphibolite grade metavolcanic and granitic sheets are thrust northeastward over low grade metavolcanic rocks, resulting in an inverted metamorphic sequence. Quartz diorite at one locality is folded into northeast vergent, Z-shaped chevron folds. Volcanic textures in the upper plate of the imbricate zone are not well preserved, whereas in the lower plate, they are moderately well preserved. Contacts between granitoid and volcanic rocks in the imbricate zone are commonly brittlely-sheared and altered. X:Z ratios from stretched, gently northeast plunging clasts range from 5:1 to 9:1. Thrusting affects, and therefore postdates formation of, Early Jurassic and older rocks. This thrusting event predates ductile extension, since the imbricate package is intruded east of the SRF by unit 10, which is deformed by ductile extension affecting 83 Ma granite (Woodsworth et al., 1983), west of the SRF.

Two phases of folding are distinguished in marble of unit 2. Second phase axial planes of upright isoclinal folds dip steeply northwest. Hinge lines plunge steeply southwest or northwest.

Northwest- and north- trending, high-angle normal faults cut all lithologies. Steep, north-northeast trending brittle shear zones are common on ridges north of Shames River. Northwest trending normal faults are commonly parallel to basaltic-andesite and lamprophyre dykes. The Shames River fault trends northwest, and juxtaposed the imbricate igneous-sedimentary package cut by granodiorite against ductilely deformed granitoid rocks of the CGC during northeast-southwest extension. Some rotation may have occurred along the fault, giving rise to the contrasting foliation attitudes across it. The youngest fractures in the area are steep, north- and north-northeast- trending brittle faults and shear zones.

METAMORPHISM

West of Shames River, peak metamorphism reaches amphibolite facies. Mafic enclaves in granitic rocks are generally fresh, and comprise biotite-hornblende-quartz-feldspar. Mylonitized granitoid rocks of unit 7 commonly contain actinolite crystals parallel to pervasive lineation, indicating retrograde, upper greenschist facies deformation and mineral growth.

Volcanic rocks east of the SRF near Shames River preserve an inverted metamorphic sequence. Kinematic indicators show that amphibolite grade volcanic and plutonic rocks are thrust northeastwards over greenschist grade volcanic and minor sedimentary rocks. The upper plate contains hornblende-feldspar-quartz assemblages retrograded to actinolite-chlorite-epidote-quartz-feldspar assemblages. Lower plate volcanic rocks contain actinolite-chlorite-epidote-quartz-feldspar assemblages typical of greenschist facies. Lineated actinolite, commonly gently northeast plunging, is thought to be syn-kinematic with respect to extensional ductile shearing which postdates thrusting.

The rapid decrease of metamorphic grade eastward is consistent with an interpretation in which more brittlely thrusted volcanic rocks now in the hanging wall of the SRF slid eastward during normal faulting above a hot, ductile footwall.

Late- to post-deformational greenschist metamorphism affects all rocks, and is characterized by growth of radiating actinolite rosettes along foliation planes. This suggests that metamorphism continued after ductile deformation, perhaps during static recrystallization and/or intrusion of the high level Ponder pluton in the north (Sisson, 1985).

CONCLUSIONS

The high-angle Shames River normal fault separates ductile deformed granitoid rocks of the CGC from greenschist to amphibolite facies volcanic and sedimentary rocks typical of Stikinia in the east. West of the fault and east of Exstew River, northeast directed extensional ductile shearing was followed by a more brittle phase of extensional faulting with the same shear sense. West of Exstew River, both southwest and northeast directed extensional ductile shearing is observed. East of SRF, earlier, northeast directed thrusting involving Lower Jurassic Hazelton Group and Lower Permian(?) igneous and metasedimentary rocks resulted in an inverted metamorphic sequence.

A late, upper greenschist grade, metamorphic event has affected all the rocks, and is related either to the intrusion of the high level Ponder pluton in the north or to post-kinematic static equilibration. The latest recognizable features in the area are north- trending, high angle brittle faults which cut, and are therefore younger than, biotite lamprophyre dykes.

The eastern boundary of the Central Gneiss Complex in this region records a change in tectonics, from Middle to Late Mesozoic thrusting, to post-Late Cretaceous to Eocene extension, and shows similarities to metamorphic core complexes.

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Eastern margin of the Coast Belt in west-central British Columbia

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Abstract

The eastern Coast Belt near 52°N is subdivided into three north-northwest trending zones. The Atnarko Complex (central zone) is a Late Jurassic(?) metamorphic tectonite complex characterized by steep, mainly north-northwest trending dextral mylonites. It represents part of a Middle-Late Jurassic basement to unconformably overlying Early Cretaceous volcanics (western zone). Late Jurassic(?) volcanics (eastern zone) may also lie unconformably on this basement.

The Atnarko Complex resembles the Late Jurassic Gamsby Complex in the Whitesail Lake area. Both may reflect Late Jurassic, dextral transpression along the east margin of the ancestral Coast Belt. They were covered and intruded in Early Cretaceous time by post-kinematic volcanics and plutons. Bounding fault zones on the west and east, respectively, belong to a regionally extensive set of Paleogene, en echelon, dextral transtensional faults along the east margin of the Coast Belt.

Résumé

On divise la zone côtière orientale, située près du parallèle 52°N, en trois zones de direction nord-nord-ouest. Le complexe d'Atnarko (zone centrale) est un complexe de tectonite métamorphique du Jurassique supérieur (?) caractérisé par des mylonites abruptes, dextres, surtout de direction nord-nord-ouest. Il représente une partie du socle du Jurassique moyen-supérieur reposant en discordance sur des roches volcaniques du Crétacé inférieur (zone ouest). Des roches volcaniques du Jurassique supérieur (?) (zone est) peuvent également reposer en discordance sur ce socle.

Le complex d'Atnarko ressemble au complexe de Gamsby du Jurassique supérieur dans la région du lac Whitesail. Ces deux complexes peuvent refléter une transpression dextre du Jurassique supérieur, le long de la marge est de l'ancienne zone côtière. Au Crétacé inférieur ils étaient couverts de roches volcaniques et de plutons post-cinématiques, intrusifs. Des zones bordées de failles à l'ouest et à l'est appartiennent respectivement à un ensemble régional de failles de transtension, dextres, en échelon, du Paléogène, longeant la marge est de la zone côtière.

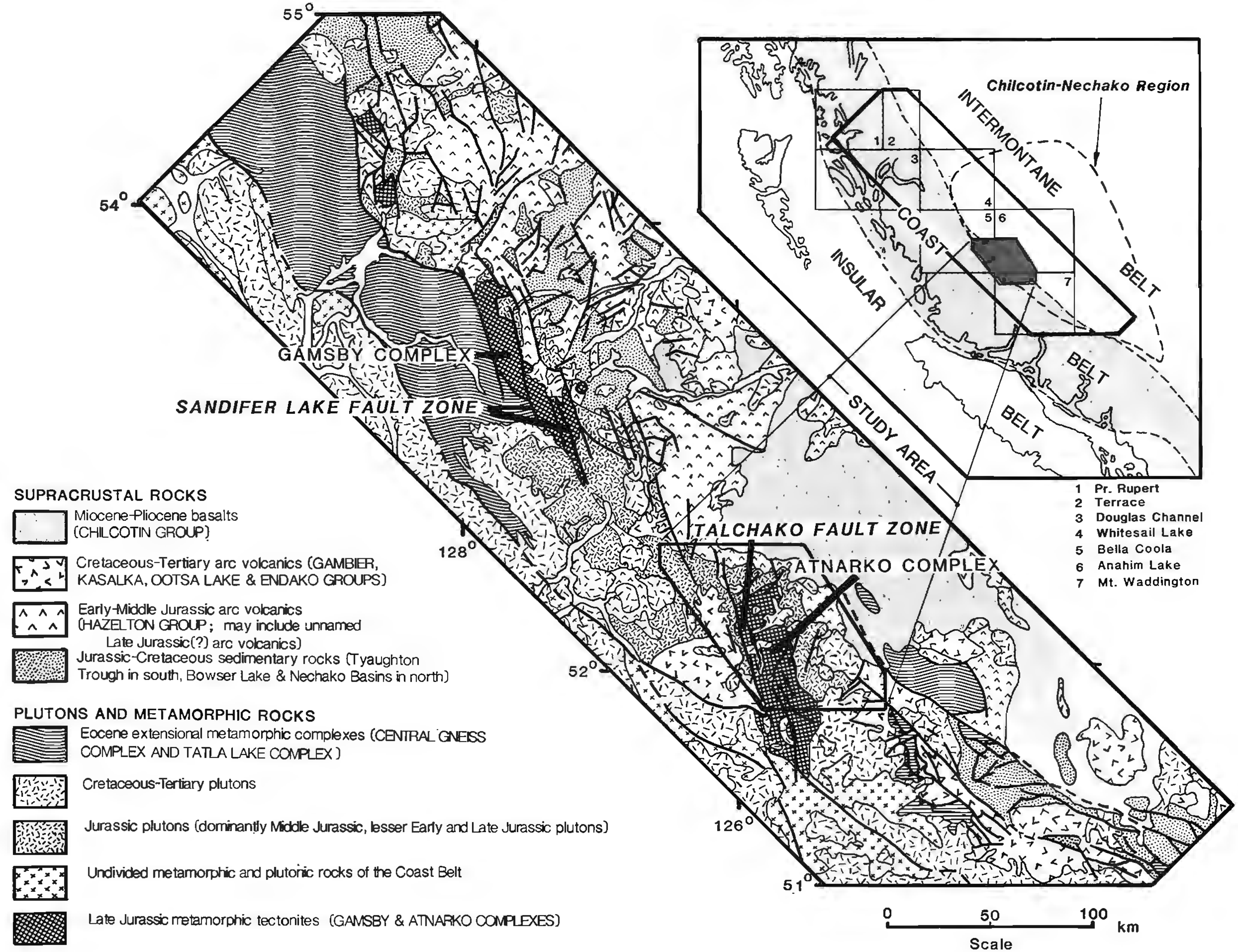

Figure 1. Generalized regional geology of the eastern Coast Belt, 51°-55°N, and location of the study area. Modified after Wheeler and McFeely (1987). Inset shows outline of Chilcotin-Nechako region and other map areas mentioned in text.

INTRODUCTION

This report summarizes field observations and preliminary conclusions of a regional study in the eastern Coast Mountains near 52°N latitude. The study area (Fig. 1) encompasses the southwest part of the Anahim Lake map area (93C), the extreme northwest part of the Mt. Waddington map area (92N), and parts of the eastern Bella Coola map area (93D). It broadly coincides with the geological boundary between the Intermontane and Coast belts and the physiographic boundary between the Interior Plateau and Coast Mountains.

The study is part of a larger effort to unravel the tectonic evolution of the Chilcotin-Nechako region. The success of this effort will depend partly on detailed studies along the western margin, because of generally elevated topography and better exposure than in the interior part of the region, which is extensively covered by glacial drift and Neogene basalt. Important clues to the evolution of the region have already been found along the east margin of the Coast Belt in the Whitesail Lake map area (Woodsworth, 1980; van der Heyden, 1989) and in the central Mt. Waddington area (Rusmore and Woodsworth, 1988, 1989). However, very little is known about the eastern Coast Belt in the Bella Coola, Anahim Lake, and northern Mt. Waddington areas, which were last studied in 1965 (Baer, 1973), 1957 (Tipper, 1969a), and 1976 (Roddick and Tipper, 1985) respectively. Few data exist that can help constrain the tectonic evolution of this composite area. The principal goal of the present study therefore is to provide pertinent geochronometric and structural data for this region.

A secondary aim is to test the validity of proposed models for the tectonic evolution of the Coast Belt (e.g. Monger et al., 1982; van der Heyden, 1989). Structural and geochronometric data from the areas north and south of the present study area (see references above) indicate complex and chronologically discordant tectonic scenarios for different parts of the Coast Belt. These apparently dissonant observations can be accommodated in a regional 'intraterrane' model for the evolution of the Coast Belt (van der Heyden, 1989), but this model is based on limited data from a poorly known, complex mountain belt. A rational choice among existing or future models will only be possible after more data from less well known areas (such as the present study area) are available.

During the 1989 field season several weeks were spent on a general reconnaissance of the study area. Emphasis was on establishing regional tectonic map units and their relations, and on collecting material for geochronometry, petrography and micro-fabric analysis. Several areas were studied in greater detail (Fig. 2, 3):

1) The Migma Mountain/Talchako Glacier area in the northwest corner of the Mt. Waddington map area. Greenschist facies metavolcanic and metaplutonic rocks are exposed within a steeply dipping, north-northwest trending mylonite zone.
2) The area between Knot Lakes and Wilderness Mountain, northwest corner of the Mt. Waddington map area. Roddick and Tipper (1985) interpreted this area as a thrust zone, with metamorphic rocks (similar to those near Migma Mountain) thrust to the east over a pluton of presumed Early Cretaceous age. The Early Cretaceous age of this pluton is debatable, as K-Ar hornblende and biotite dates are markedly discordant.
3) The Ptarmigan Lake and Glacier Mountain areas in the southwest Anahim Lake map area. Rocks in this area, previously mapped as granite gneiss, amphibolite, schist, migmatite and gneissic granodiorite (Tipper, 1969a), are a northerly continuation of the metamorphic rocks in the Knot Lakes area.

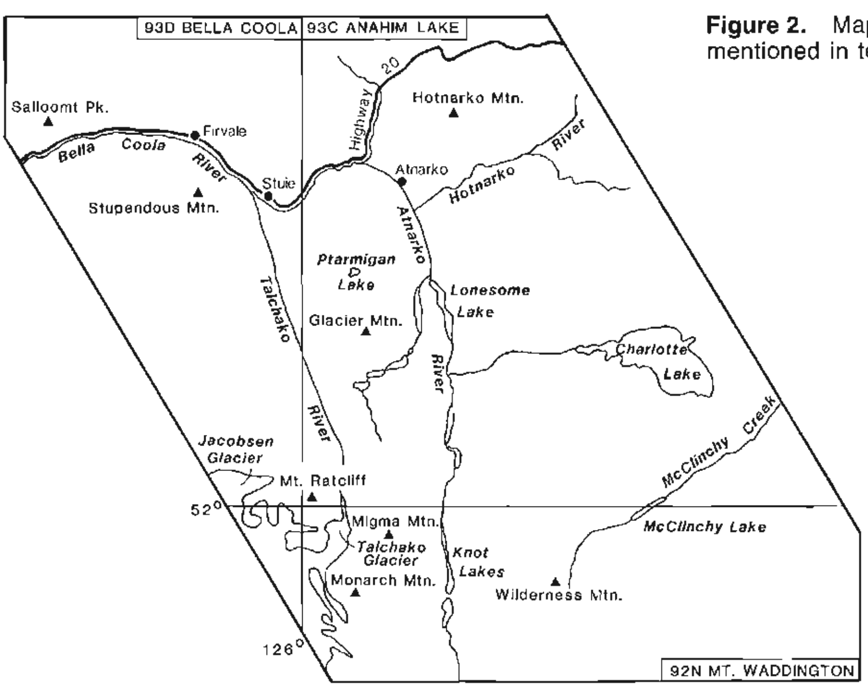

Figure 2. Map of study area showing geographic features mentioned in text.

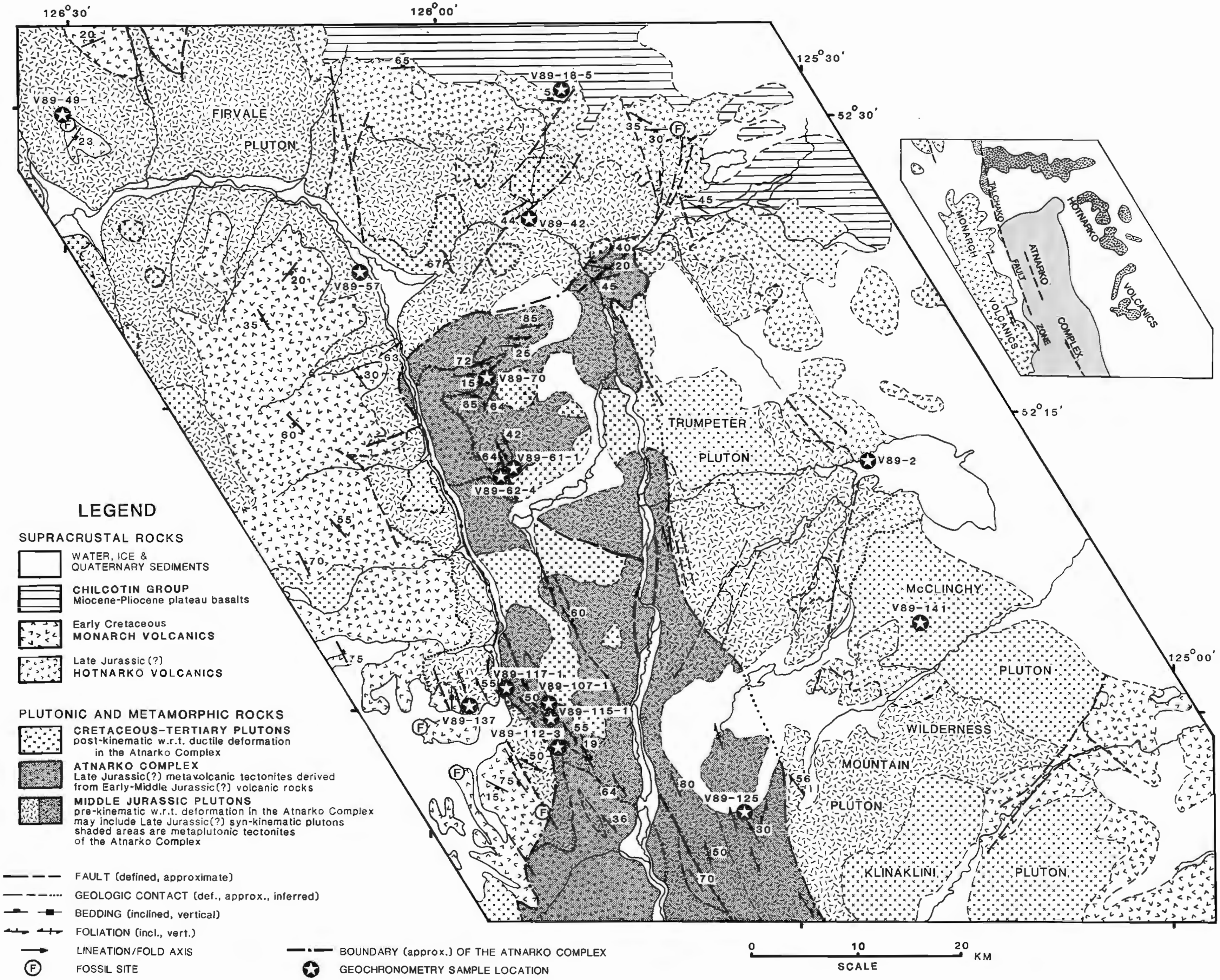

Figure 3. Geological map of the study area. Geology modified after Baer (1973), Tipper (1969a), and Roddick and Tipper (1985). Locations of geochronometry samples in process are indicated. Inset shows subdivision of the area into 3 zones.

Of 35 samples collected for geochronometry, 13 are presently being processed at the Ottawa GSC Geochronology Laboratory for U-Pb dating, 4 for $^{40}Ar/^{39}Ar$ dating (locations are plotted in Fig. 3). Petrography and microfabric analysis of selected samples are in progress, and further field and laboratory studies are planned to augment the tentative conclusions given here.

REGIONAL SETTING

The Chilcotin-Nechako region is a wedge-shaped geographic subdivision of the southwestern Intermontane Belt, bounded on the west by Jurassic to early Tertiary crystalline rocks of the Coast Belt (Fig. 1). Most of this region has traditionally been viewed as part of Stikinia, a large allochthonous terrane of the Canadian Cordillera. Mesozoic supracrustal and plutonic rocks that can be assigned to Stikinia without ambiguity are present in the northern and northwestern part of the area (Whitesail Lake to Anahim Lake map areas). These rocks are dominantly marine arc volcanics and sediments belonging to the Lower and Middle Jurassic Hazelton Group. The stratigraphic cover to Stikinia consists of Middle Jurassic and Lower Cretaceous marine epiclastic sediments of the Bowser Lake and Skeena groups, as well as Upper Cretaceous and Tertiary nonmarine volcanic rocks of the Kasalka, Ootsa Lake, Endako and Chilcotin groups.

During most of Mesozoic and all of Cenozoic time Stikinia was situated somewhere along the western margin of the North American continent, where it was subjected to tectonic events related to large scale interaction between continental and oceanic lithosphere. For instance, the change in tectonic setting from a marine volcanic arc (Hazelton Group) to a successor basin environment (Bowser Lake Group) may be related to the emplacement of allochthonous Stikinia to the western margin of North America (Quesnellia) in Middle Jurassic time, with closure of the intervening Cache Creek ocean. The change from successor basin sedimentation to subaerial, dominantly volcanic conditions (Kasalka, Ootsa Lake and Endako groups), as well as stratigraphic changes associated with an Early Cretaceous marine transgression (Skeena Group), are related to the Late Jurassic to early Tertiary evolution of the Coast Belt. The Coast Belt exposes the roots of successively overprinting, west-facing magmatic arcs and some of their supracrustal framework and cover, modified by large structures related to transpressional and transtensional forces within the new leading edge of the continental crust (van der Heyden, 1989).

The southwestern part of the Chilcotin-Nechako region is underlain mainly by Jurassic and Cretaceous supracrustal rocks of the Tyaughton Trough (Jeletzky and Tipper, 1968; Tipper, 1969b), an important tectono-stratigraphic subdivision of the Intermontane Belt. The Tyaughton Trough may have formed internal to the western margin of North America in response to changing plate vectors following the Late Jurassic genesis of the Coast Belt (van der Heyden, 1989). Areally coincident with the Tyaughton Trough are important compressive and transverse structures, including a Late Cretaceous, northeast-verging fold and thrust belt (Rusmore and Woodsworth, 1988, 1989), and abundant northwest to north-northwest trending strike-slip faults (e.g. Tipper, 1969b; Glover et al., 1988). Between the Anahim Lake and Whitesail Lake areas crystalline rocks of the Coast Belt impinge directly on volcanic and sedimentary strata of Stikinian affinity. There is no record here of a sedimentary basin akin to the Tyaughton Trough, nor do there appear to be any major structures similar to those associated with the Tyaughton Trough. In the Whitesail Lake area, the eastern margin of the Coast Belt is underlain by the Gamsby Complex, a metamorphic tectonite zone representing a remnant of a large early Late Jurassic magmatic and orogenic belt superimposed on Stikinia (van der Heyden, 1982, 1989). Other remnants have been found as far north as the Terrace area and, as discussed below, a large remnant of this belt is believed to underlie part of the present study area.

Much of the western Chilcotin-Nechako region appears to have a crystalline basement. High grade metamorphic rocks in the Tatla Lake area belong to a Eocene extensional core complex that may extend beneath much of the region (Friedman and Armstrong, 1988). A similar situation is postulated to exist farther north, in the Whitesail Lake and Terrace areas, where rocks of the Intermontane Belt and eastern Coast Belt (including the Gamsby Complex) may embody the brittle upper plate above ductilely extended middle to lower crust exposed in the Eocene Central Gneiss Complex (van der Heyden, 1989; Heah, 1990). In addition to the structural basement-cover relations, there may be several regionally extensive crystalline-supracrustal unconformities present along the western margin of the Chilcotin-Nechako region (discussed below).

ROCKS AND STRUCTURES OF THE STUDY AREA

The area can be roughly subdivided into three main, north-northwest trending zones (Fig. 3). The western and eastern zones are underlain by low-grade volcanic and sedimentary rocks with a plutonic basement. The volcanic rocks in the western and eastern zones are here informally named the Monarch volcanics and Hotnarko volcanics, respectively. These two zones merge to the north, enclosing a central zone dominated by greenschist and amphibolite facies metamorphic tectonites. This central zone is here named the Atnarko Complex, after the Atnarko River which drains much of the surrounding area.

Atnarko Complex

The Atnarko Complex is a 20-25 km wide metamorphic and structural culmination composed of deformed and metamorphosed plutons and volcanic rocks, intruded by post-kinematic plutons and locally abundant post-kinematic andesitic dykes. In most places this zone is characterized by steeply dipping, north-northwest trending tectonite fabrics, commonly mylonitic, that display a dextral sense of bulk shear. The Atnarko Complex appears to be dominantly a greenschist facies unit, but the area between Knot Lakes and Wilderness Mountain is in amphibolite facies.

About 85 % of the Atnarko Complex is made up of strongly deformed, commonly mylonitic quartz-diorite and granodiorite (Fig. 4). Local compositional variations include diorite and possibly minor amounts of granite or

Figure 4. Mylonitic quartz diorite intruded by a post-kinematic granite dyke, Atnarko Complex near Ptarmigan Lake. Dyke is 30 cm wide.

quartz monzonite. Primary mafic minerals are strongly chloritized and epidote and plagioclase have been saussuritized. Fortunately, primary hornblende and biotite are locally preserved, providing material that may prove suitable for K-Ar and $^{40}Ar/^{39}Ar$ dating. Contacts with metavolcanic rocks are invariably gradational, with transition zones marked by large metavolcanic screens oriented parallel to local structural trends and by strongly deformed agmatite zones. Field relations in these zones suggest that both pre- and syn-kinematic plutons are present in the deformed suite.

Metavolcanic rocks in the Atnarko Complex consist of strongly strained, greenschist facies, felsic-mafic banded rocks (Fig. 5), commonly interlayered with mylonitic granitoid material. Banding in these rocks may locally represent relict primary layering, but in most places the layering is highly transposed and differing lithologies are structurally interdigitated on all scales. The metavolcanic rocks of the Atnarko Complex are identical in almost all respects to those of the Gamsby Complex in the Whitesail Lake area (van der Heyden, 1982, 1989).

Along its margins the Atnarko Complex grades into variably deformed granitoid rocks. The northern margin of the Atnarko Complex along the Atnarko River is characterized

Figure 5. Greenschist facies, interdigitated felsic-mafic metavolcanics, Atnarko Complex, Migma Mountain area. Note hammer (slightly left of centre) for scale.

by steeply dipping, east-northeast trending, greenschist facies tectonite fabrics and by a south to north ductile-brittle transition within deformed granitoid rocks. The western boundary near Migma Mountain dips moderately to steeply northeast, and is also marked by a ductile-brittle transition within greenschist facies granitoid rocks. Farther north, the western boundary follows the Talchako River valley (Fig. 2), a pronounced north-northwest trending topographic lineament that probably marks the locus of a major fault. Several steep, north-northwest trending brittle faults occur near this lineament, both within the Atnarko Complex and in the volcanic rocks to the west. These faults, which are clearly much younger and unrelated to ductile deformation in the Atnarko Complex, are here collectively referred to as the Talchako Fault Zone (Fig. 3).

The location and nature of the eastern boundary of the Atnarko Complex is obscured by vegetation, glacial drift, and crosscutting plutons, but it may locally coincide with the eastern limit of Tipper's (1969a) map unit 7 (gneissic granodiorite). The proximity of ductilely strained, greenschist facies tectonites along the Hotnarko River to the essentially unmetamorphosed Hotnarko volcanics to the east indicates that the contact there must be either a fault or an unconformity. Farther south, the western margin of the pre- to syn-kinematic Wilderness Mountain pluton (Fig. 3) is a steeply dipping, north-northwest trending, amphibolite facies mylonite zone (Fig. 6) with a pronounced easterly decreasing strain gradient. This zone may represent the eastern structural boundary of the Atnarko Complex; the contact relation of this pluton with the Hotnarko volcanics farther east (west of McClinchy Lake) is unknown.

The Atnarko Complex extends farther south into the Mt. Waddington map area, where Roddick and Tipper (1985) referred to it as the Central Gneiss Complex. Although traditional usage of this name has commonly included all high grade gneisses in the Coast Belt, I propose to reserve it for those areas underlain by gneisses with Eocene hornblende and biotite cooling dates (e.g. the Central Gneiss Complex in the Whitesail Lake, Douglas Channel, Prince Rupert and Terrace areas; see van der Heyden, 1989 for a review). There are presently only four K-Ar dates from the

Figure 6. View looking south at vertical, amphibolite facies mylonites in the western margin of the Wilderness Mountain pluton, Atnarko Complex.

Atnarko Complex (140 ± 11 Ma hornblende and 86.1 ± 3.2 Ma biotite from sample GSC 78-61/62, Wanless et al., 1979; 121 ± 12 Ma hornblende and 65.9 ± 2.6 Ma biotite from sample GSC 80-22/23, Stevens et al., 1982). These dates are from the amphibolite facies tectonite zone along the western margin of the Wilderness Mountain pluton. They clearly indicate that the Atnarko Complex is not a Eocene metamorphic complex, but rather that pluton emplacement, metamorphism and ductile deformation are pre-Early Cretaceous events that were followed by cooling below 550°C in Early Cretaceous time.

The Atnarko Complex bears a remarkable lithological resemblance to the early Late Jurassic Gamsby Complex in the Whitesail Lake area. The Gamsby Complex has similar Early Cretaceous hornblende cooling dates, and is partly derived from Middle Jurassic (Bajocian) granitoid protoliths (van der Heyden, 1989) that are quite similar to the plutons which surround and structurally grade into the Atnarko Complex. These latter plutons are at least as old as Late Jurassic because they are unconformably overlain by the Early Cretaceous Monarch volcanics (see below). Based on these observations it is tempting to conclude that the Atnarko Complex is also an early Late Jurassic metamorphic tectonite zone, superimposed on Middle Jurassic granitoid rocks and their older volcanic country rocks. This conclusion, which would have important regional implications if it can be verified by U-Pb and K-Ar geochronometry, must be considered tentative at the moment.

Monarch volcanics and their granitoid basement

The altered but undeformed plutons bounding the Atnarko Complex to the west are unconformably overlain by low grade Early Cretaceous volcanic and sedimentary rocks. Based on lithological similarities and the presence of Hauterivian-Barremian fossils (Baer, 1973; Tipper, 1969b) these rocks may be correlative with the Gambier and Fire Lake groups of the southern Coast Belt (G.J. Woodsworth, pers. comm., 1989; Wheeler and McFeely, 1987). They are here informally called the Monarch volcanics after Monarch Mountain, the highest peak in the study area, which is entirely underlain by the volcanic succession.

The Monarch volcanics are dominantly pyroclastic, with abundant tuffs and coarse volcanic breccias, but volcanic flows are also common throughout the unit. Sedimentary intercalations are minor, although they are more common near the base of the assemblage. The unit appears quite unaltered except near intrusions and brittle faults, where it becomes difficult to identify. The rocks are various shades of green, mauve and less commonly maroon. Volcanic breccias and lahars locally contain very coarse clasts. Flows are mostly massive and are almost everywhere characterized by plagioclase phenocrysts; pyroxene and hornblende porphyry flows are local varieties. Some flows are amygdaloidal, with vesicles filled with quartz, epidote and possibly zeolites, prehnite, and pumpellyite. The Monarch volcanics are probably andesitic, but lesser amounts of basalt and dacite may be present.

Abundant green and grey andesitic dykes, many with stubby plagioclase phenocrysts, form dyke swarms in the lower parts of the volcanic succession north of Monarch Mountain. These dykes are difficult to distinguish from flows in the unit, but bedding attitudes in nearby argillite layers indicate discordant, crosscutting contact relations. These dykes are inferred to be feeders to volcanic flows higher up in the unit.

The sediments in the unit are mostly rusty weathering black and grey argillites forming layers up to 20 m thick. Thin laminations and graded bedding were observed in a few places, but these sediments are mostly massive. Near Saloomt Peak, the sedimentary base of the unit was examined in some detail. At the contact fine grained arkose coarsens upward over 2 m and is abruptly overlain by about 30 m of clast-supported granitoid cobble and boulder conglomerate. Clasts are derived from directly underlying, severely altered quartz diorite or granodiorite. The conglomerate is interbedded with discontinuous, coarse grained arkose lenses, rusty weathering black argillites containing plant fragments, and rare, thin bedded limestone. Upward the conglomerate gives way to a thick succession of argillites and sandstones, which are in turn overlain by typical Monarch volcanics. A previous Jurassic age assignment for fossils collected from this locality (Baer, 1973) is presently considered unreliable (T. Poulton, pers. comm., 1989). As the sediments and overlying volcanics are indistinguishable from typical Monarch volcanics, they are here considered correlative.

No other exposures of the basal unconformity within the study area were known prior to the present study, although they are probably not uncommon west of the Talchako River. An important new location was found in 1989 south of Mt. Ratcliff. Here, matrix-supported coarse volcanic breccias and minor interbeds of grey argillite contain up to 15% angular to subrounded, altered hornblende quartz diorite clasts (Fig. 7) derived from an immediately underlying pluton. The sedimentary succession seen near Saloomt Peak is not present at this location. The actual contact was buried under a few feet of snow and rubble, but there is no doubt about its unconformable nature.

The hornblende quartz diorite pluton underlying the volcanics at Mt. Ratcliff is important because it grades into the Atnarko Complex west of Migma Mountain. Field relations

Figure 7. Mixed volcanic-sedimentary breccia at base of Lower Cretaceous Monarch volcanics, southeast of Mt. Ratcliff. The breccia contains 15% hornblende quartz diorite clasts derived from a underlying pluton.

suggest that a brittle-ductile transition occurs within this pluton, which must therefore be pre- or syn-kinematic. Abundant post-kinematic andesite dykes in this area intrude the Atnarko Complex, the hornblende quartz diorite, and the Monarch volcanics. Collectively these relations suggest that the Atnarko Complex is part of a crystalline basement to the Monarch volcanics. They lend considerable support to the conjecture that the Atnarko Complex is a Jurassic metamorphic and structural complex.

Variable bedding attitudes indicate that the Monarch volcanics are broadly folded, and the succession is clearly disrupted by later faults, especially along the Talchako Fault Zone, where steep bedding predominates adjacent to discrete shear zones. Despite these disrupting structures the assemblage is still recognizable as a regional unconformable cover on a granitoid basement. Erosional remnants, such as the one near Saloomt Peak, are preserved in the northwest corner of the study area. The presence of a dominantly granitoid basement beneath the Lower Cretaceous volcanics was not noted in earlier studies period Baer (1973) did not distinguish between granitoid rocks and rocks of volcanic origin in a 'greenstone' unit (Baer's unit 7) that underlies extensive areas in the eastern half of the Bella Coola map area. Where examined along the Bella Coola and Talchako rivers, this unit consists entirely of granitoid plutonic material, injected by locally abundant mafic dykes. Baer (1973) showed several other areas underlain by diorite and quartz diorite, but how much of the total unit is granitoid is presently unknown. In any case, the new data argue against the presence of a reentrant in the plutonic eastern boundary of the Coast Belt, such as has appeared on previous maps (e.g. Tipper et al., 1982; Wheeler and McFeely, 1987).

Hotnarko volcanics and their granitoid basement(?)

The eastern part of the study area coincides with the physiographic edge of the Coast Mountains. Isolated ranges surrounded by extensions of Interior Plateau erosion surfaces are underlain by thick piles of structureless volcaniclastic rocks, volcanic flows, and lesser epiclastic sediments, intruded by quartz diorite and granodiorite plutons. Outcrop in this area is generally poor, and large areas above treeline are commonly covered by tundra soils and felsenmeer. Exposures are heavily covered by lichen, highly fractured, and locally faulted. These factors make it virtually impossible to accurately locate and identify the nature of contacts between map units. Near assumed intrusive contacts with plutons the volcanic rocks are commonly baked and intruded by granitoid dykes. Tipper (1969a) has documented a clearly intrusive contact only north of McClinchy Lake. These intrusions are believed to be Late Cretaceous or early Tertiary, based on lithological similarities with plutons of that age in the Whitesail Lake area.

The Hotnarko volcanics are informally named after the Hotnarko River, which drains part of the area underlain by these rocks. They are mainly structureless volcanic flows and breccias, with minor interbedded black shales and volcanic sandstones. They are difficult to distinguish from the Monarch volcanics, with which they share many characteristics. The Hotnarko volcanics appear to unconformably overly granitoid rocks that can be traced, via a brittle-ductile transition, into the Atnarko Complex. Near Hotnarko Mountain the apparent base of the volcanic succession consists of mauve coloured tuff-breccias with rare granitoid clasts. These overly and are locally interbedded with thin chert-pebble conglomerate and immature volcanic sandstone layers that may have a component of granitoid detritus. In addition, nearby float includes matrix-supported volcanic cobble conglomerate containing a few granitoid clasts. These features are all near scattered outcrops of altered hornblende quartz diorite, but the contact itself was not seen. Bedding attitudes suggest that the quartz diorite immediately underlies the clastic rocks at this location.

These observations and the regionally homoclinal appearance of the volcanic succession (the contact with the underlying plutonic rocks generally follows topographic contours) make it tempting to correlate the Hotnarko volcanics with the Monarch volcanics. However, fossils collected by Tipper (1969a) from the basal sedimentary succession on Hotnarko Mountain were tentatively identified as Bathonian to lower Oxfordian (Jeletzky in Baer, 1973). The fossil site was revisited as part of this study, and several samples of volcanic sandstone with *Trigonia* sp., belemnites, and coral fragments were collected. Preliminary examination (T. Poulton, pers. comm., 1989) revealed the *Trigonia* sp. to be a Callovian-Oxfordian form, confirming the previous age assignment, and indicating that the sediments at the base of the Hotnarko volcanics are time-correlative with the Ashman Formation (Tipper and Richards, 1976) of central British Columbia.

The conformably overlying Hotnarko volcanics may thus be as young as Oxfordian. Two samples of rhyolitic volcanics have been collected for U-Pb geochronometry in an effort to confirm this tentative conclusion. If verified, this could have important implications. Thick volcanic accumulations of Callovian age are unknown in British Columbia, but the Oxfordian Netalzul volcanics form scattered piles in the Smithers and Hazelton map areas (Tipper and Richards, 1976). Volcanic rocks of this age may also be present in the Whitesail Lake area (van der Heyden, 1989), but this is not proven. The presence of widespread early Late Jurassic plutons in the Coast Belt (van der Heyden, 1989; R.M. Friedman, pers. comm., 1989) certainly raises the possibility that coeval volcanic strata may be exposed locally along its eastern margin. The tentative conclusion drawn here is that the Hotnarko volcanics may represent one of the few remaining supracrustal remnants of an early Late Jurassic magmatic arc.

The apparent unconformity between the Late Jurassic(?) Hotnarko volcanics and plutonic basement in the Hotnarko River area is in accord with the suspected Middle Jurassic age of these plutons. The plutons are lithologically identical to Middle Jurassic (Bajocian) plutons in the Whitesail Lake area. They are clearly distinguishable from younger plutons by their more mafic character (notably by their hornblende content), higher degree of alteration and fracturing, and especially by the local abundance of agmatite zones and crosscutting mafic dykes. U-Pb geochronometry is in progress to verify the Middle Jurassic age of these plutons.

Structural notes

Two main north-northwest and northeast to east-northeast structural trends can be recognized. At least 3 phases of folding have been identified in the Atnarko Complex, and several generations of ductile and brittle shear zones may be present. Only a relative chronology can be established in individual areas, but the tentative age assignments of the units discussed above can be used to suggest age limits for some of the major structures in the study area.

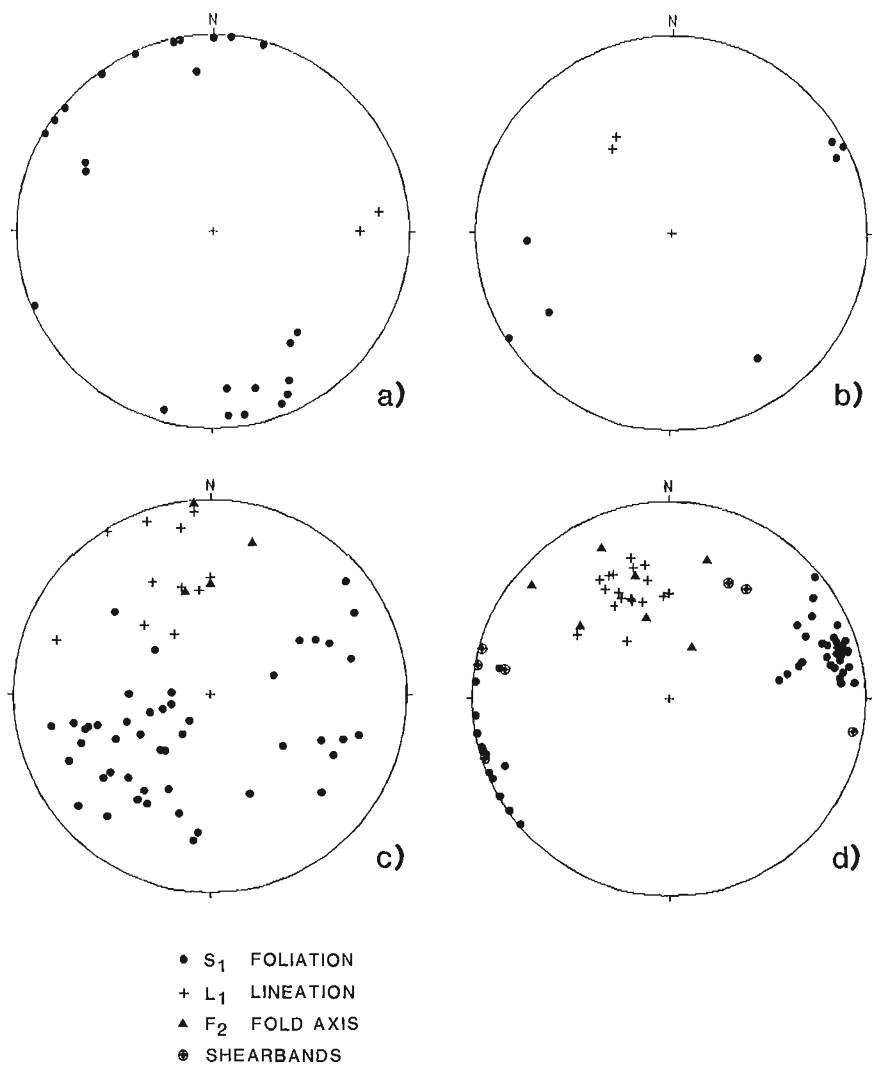

Figure 8. Equal area stereographic projections of structural data from selected parts of the Atnarko Complex; a = Ptarmigan Lake area, b = Glacier Mountain area, c = Migma Mountain area, d = Wilderness Mountain area.

Structural attitudes from selected areas are plotted on stereo diagrams (Fig. 8), and representative attitudes for the area as a whole are plotted in Figure 3. The oldest structures recognized in the area are generally steeply dipping ductile tectonite fabrics in the Atnarko Complex (S_1; Fig. 6). Tight to isoclinal, commonly intrafolial folds (F_1) are present in the interdigitated and banded felsic-mafic metavolcanic and metavolcanic-metaplutonic successions (Fig. 9). F_1 fold hinges are everywhere parallel to a locally well developed stretching lineation (L_1) defined mainly by quartz rodding in metaplutonic rocks. A dextral sense of bulk shear for the north-northwest trending tectonites was determined from asymmetry of pinch-and-swell structures (Fig. 10), discrete offsets along narrow shear bands, and from preliminary investigation of microfabrics in oriented samples (Fig. 11).

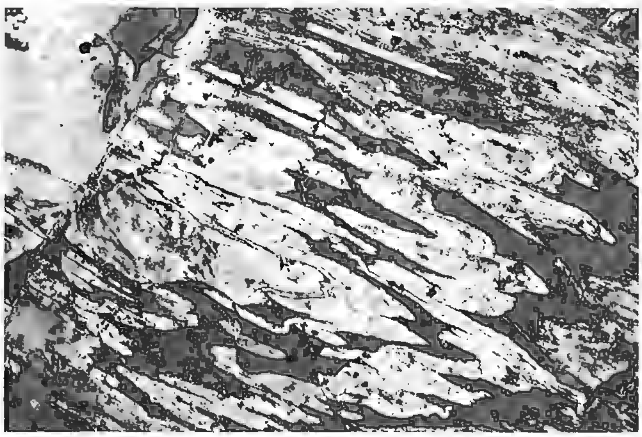

Figure 9. F_1 folds in banded felsic-mafic metavolcanics, Atnarko Complex, Migma Mountain area. Note pencil at top for scale.

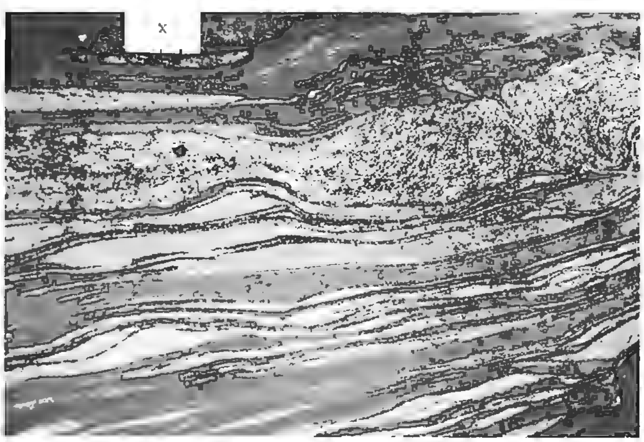

Figure 10. Asymmetric pinch-and-swell structure in mylonitic, amphibolite facies orthogneiss, western margin of the Wilderness Mountain pluton, Atnarko Complex. The shear is dextral; the extension lineation is contained in the plane of the picture.

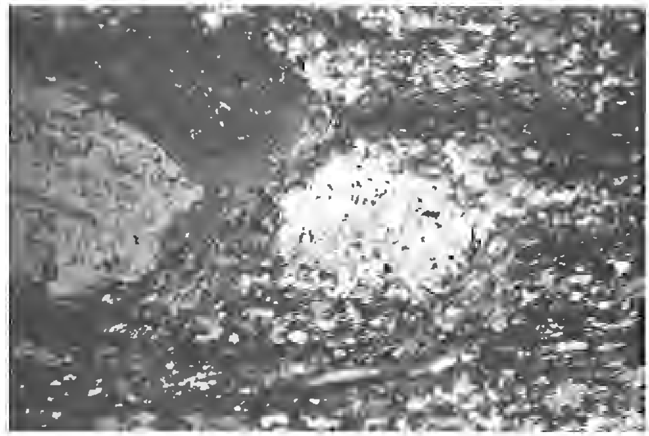

Figure 11. Photomicrograph showing dextral shear bands bounding a recrystallized plagioclase porphyroclast, from a amphibolite facies tectonite, western margin of the Wilderness Mountain pluton, Atnarko Complex.

Figure 12. F_3 buckle folds, superimposed on a F_2 isoclinal fold in mylonitic quartz diorite, Glacier Mountain area, Atnarko Complex.

Along the west side of the Atnarko Complex the F_1 foliation is folded into open to tight, locally isoclinal but more commonly asymmetric and disharmonic folds (F_2), with foldaxes oriented subparallel to L_1. Local crenulations of F_1 foliation and buckles of F_2 axial planes (Fig. 12) are evidence for a third phase of folding (F_3) in the Atnarko Complex. Note that the S_1 foliation and L_1 lineation are present throughout the Atnarko Complex; the younger structures may be only locally developed.

The general north-northwest trend of ductile fabrics, sense of shear, and metamorphic grade, suggest that the Atnarko Complex represents the mid-crustal level of a wide, dextral shear zone trending slightly oblique (clockwise) to the regional trend of the Coast Belt. However, the northeast trending fabrics along the northern edge of the Atnarko Complex are difficult to reconcile with this suggestion. The relationship between these two structural trends is unknown at present.

DISCUSSION AND CONCLUSIONS

Conclusions presented here are preliminary because they are largely based on circumstantial evidence. U-Pb, K-Ar and $^{40}Ar/^{39}Ar$ geochronometry of pertinent samples from the study area is currently in progress.

The striking resemblance of the Atnarko Complex to the Late Jurassic Gamsby Complex, including its gradational relations with adjacent Middle Jurassic(?) granitoid plutons and its tentative Late Jurassic age, invites the proposal that they are related, originally contiguous units. Along most of its eastern margin the Gamsby Complex is characterized by steeply dipping, north-northwest trending mylonites (van der Heyden, 1982, 1989; G.J. Woodsworth, pers. comm., 1989) similar to those seen in the Atnarko Complex. Both may have formed in a large, dextral, transpressional ductile shear zone, along which the Gamsby Complex was offset from the Atnarko Complex in Late Jurassic time. The presence of a Late Jurassic tectonite belt along the east side of the Coast Belt (south of 55°N) has major implications for the Mesozoic evolution of the adjacent Intermontane Belt and for the nature of the Insular-Intermontane superterrane boundary. These implications are discussed in detail elsewhere (van der Heyden, 1989).

Late Jurassic tectonism in the study area is reflected in the unconformity at the base of the Early Cretaceous Monarch volcanics, and perhaps also by an unconformity(?) at the base of the Hotnarko volcanics. These unconformities reveal a regional Middle-Late Jurassic crystalline basement, of which the Atnarko and Gamsby complexes appear to be an integral part. Early Cretaceous volcanics that are lithologically identical to the Monarch volcanics unconformably(?) overly the Middle Jurassic Trapper pluton in the Whitesail Lake area (van der Heyden, 1989). The adjacent Gamsby Complex is intruded by post-kinematic Early Cretaceous plutons. These Early Cretaceous volcanics and plutons are products of a west-facing magmatic arc superimposed on the ancestral Late Jurassic Coast Belt.

The Gamsby Complex is bounded on its east side by the Sandifer Lake Fault Zone (Fig. 1), a family of Eocene-Oligocene north-northwest trending brittle faults that show evidence for dextral and west-side-down normal displacement (van der Heyden, 1989). The north-northwest trending Talchako Fault Zone at the western margin of the Atnarko Complex also shows evidence for dextral and west-side-down normal motion. These similarities in bounding structures suggest that the Sandifer Lake and Talchako Fault Zones may be closely related. They appear to be members of a regionally extensive set of oblique, north-northwest trending, en-echelon faults along the east side of the Coast Belt. These faults reactivated Late Jurassic ductile shear zones in the Gamsby and Atnarko Complexes. Local continuity across both the Sandifer Lake and Talchako Fault zones argues against large horizontal displacements along these younger fault systems. Evidence from the Whitesail Lake area suggests that these faults may have formed in a Paleogene dextral transtensional regime and that they are related to late stages in the development of extensional core complexes within and along the Coast Belt (e.g. the Central Gneiss Complex and the Tatla Lake Complex; van der Heyden, 1989).

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A progress report on the structural and tectonic framework of the southern Coast Belt, British Columbia

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Journeay, J.M., A progress report on the structural and tectonic framework of the southern Coast Belt, British Columbia, in Current Research, Part E, Geological Survey of Canada, Paper 90-1E, p. 183-195, 1990.

Abstract

Assembly of crustal fragments within the southern Coast Belt (CB) is linked to progressive shortening and transcurrent displacements along the continental margin since Early Cretaceous time. Shortening began along the eastern flank of the CB with thin-skinned imbrication and southwest-directed thrust faulting of volcanic arc and flanking oceanic sequences, and culminated in the Late Cretaceous with thick-skinned imbrication and westward overthrusting of the Cascade Metamorphic Core. Northwest-striking transcurrent faults, such as the Harrison Lake Shear Zone are partly coeval with this thrust system, and may have accommodated orogen-parallel components of displacement during this stage in the evolution of the Coast Belt.

Late Tertiary fault structures are represented by a system of northeast-striking dextral transcurrent faults and associated northwest-striking high-angle reverse faults, and may record crustal shortening associated with eastward subduction of oceanic lithosphere in the late Tertiary.

Résumé

Un assemblage de fragments crustaux au sein de la zone côtière sud est lié à un raccourcissement progressif et à des déplacements à décrochement horizontal le long de la marge continentale depuis le Crétacé inférieur. Le raccourcissement a commencé le long du flanc oriental de la zone côtière avec des écailles peu épaisses et des failles chevauchantes de direction sud-ouest d'un arc volcanique et des séries océaniques flanquant celui-ci et s'est terminé au sommet, au Crétacé supérieur par des écailles épaisses et des chevauchements vers l'ouest du noyau métamorphique des Cascades. Des failles à décrochement horizontal, de direction nord-ouest, comme la zone de cisaillement de Harrison Lake, sont en partie contemporaines avec ce système chevauchant et pouvant avoir accommodé des composantes de déplacement parallèles à l'orogénèse au cours de cette phase de l'évolution de la zone côtière.

Des structures faillées du Tertiaire supérieur sont représentées par des failles à décrochement horizontal dextre de direction nord-est et associées à des failles inverses à fort pendage en direction nord-ouest; elles peuvent être l'indice d'un raccourcissement de la croûte associé à une subduction vers l'est de la lithosphère océanique au Tertiaire supérieur.

* Change Gambier Group to Gambier Assemblage
* Correct Spelling q Riniard Froedman, Spelled Friedman not Freidman

INTRODUCTION

Overview

The Coast Belt of southern British Columbia is an accretionary complex of crustal dimensions which straddles the boundary between the Insular and Intermontane superterranes. It has evolved in stages throughout the Mesozoic and Cenozoic, and has been the locus of thin- and thick-skinned imbrication, right-lateral transcurrent faulting, polyepisodic metamorphism and arc-type magmatism since Early Cretaceous time.

Early stages in the development of the accretionary complex include the impingement and amalgamation of terranes along the continental margin in middle Jurassic to Early Cretaceous time, either by collision of previously assembled crustal blocks (Alexandria-Wrangellia-Stikinia megaterrane) along the Cache Creek-Bridge River suture (van der Heyden and Woodsworth, 1988), or by subduction of small oceanic and/or marginal basins and incremental stacking of flanking terranes within the overriding plate of an east-dipping subduction zone (Monger, 1986; Gallagher et al., 1988; Brandon et al., 1988). Major constructional phases in the evolution of the Coast Belt-Cascade orogen, occurred primarily in mid- and Late Cretaceous time and include large-scale imbrication and westward telescoping of the accretionary complex. Brandon and Cowan (1985), Monger (1986) and McGroder (1988) have argued that this constructional phase of the orogen was a consequence of Alpine-style ''collision'' and thrust imbrication above an evolving subduction zone. These models imply significant shortening and tectonic thickening along the continental margin, and provide a link between deformation and metamorphism within foreland and hinterland zones of the accretionary complex. Brown (1987), on the other hand, argued for northward translation of thrust sheets within the accretionary complex, driven by large-scale displacements along a system of orogen-parallel dextral transcurrent faults. The model is based on kinematic analyses of fault zone structures within the northern Cascades, and links this stage in the evolution of the orogen to a mid-Cretaceous change in the angle and rate of convergence between North American and Kula/Farallon plates.

These studies show that large-scale contractional and strike-slip fault systems have played important roles in the evolution of the southern Coast-Cascade orogen but do not address how these fault systems interact in space and time in different parts of the orogen. This study focuses on the geometry, kinematics and timing of major thrust and strike-slip fault systems along the southeast flank of the Coast Belt (Fig. 2) and the extent to which each system has controlled the structural and thermal evolution of terranes in this part of the Canadian Cordillera. Work to date includes 1:50 000 scale mapping, kinematic analysis and geochronometry of fault systems separating the Cascade Metamorphic Core from adjacent rocks of the Coast Plutonic Complex (Journeay and Csontos, 1989; Journeay, 1989).

Previous work and current research

Our understanding of southern Coast Belt architecture (49-51°N) is based largely on 1:250 000 scale mapping and regional syntheses of Crickmay (1930), Roddick (1965), Roddick and Hutchison (1973), Woodsworth (1977), Monger (1986, in press-a,b) and more detailed studies in the vicinity of Harrison Lake by Reamsbottom (1971, 1974), Lowes (1972), Pigage (1976), Bartholomew (1979), Gabites (1985) and Arthur (1986). The results of these studies are the basis for a new compilation of southeastern Coast Belt geology (49°30′-50°30′, Fig. 2).

This compilation incorporates the results of several new research projects that have been undertaken as part of this study (Fig. 2). These include 1:50 000 scale mapping and geochemical studies focussed on the internal stratigraphy, structure and isotopic signature of fluid systems within the Gambier Group (Fire Lake District) by Greg Lynch (GSC), 1:25 000 scale mapping and analysis of contact relationships between the Gambier Group and Cadwallader Terrane in the Lillooet Lake region by Janet Riddell (University of Montana), 1:25 000 scale mapping, structural analysis and thermo-barometric studies of the Coast Belt-Northwest Cascade Fault system (CB-NCS) in the Harrison Lake region by Thornton Tyson (University of Washington), and geochronometry of fault systems and plutonic suites by Richard Friedman (University of British Columbia). Details of these studies are reported separately in this volume.

Tectonic assemblages

The southeast flank of the Coast Belt orogen can be subdivided into three northwest-trending tectonic domains (Fig. 1). These domains are cut by intrusions of the Coast Plutonic Complex, which collectively make up more than 60% of the southern Coast Belt orogen by area.

The westernmost domain is a magmatic arc complex comprising plutonic suites ranging from Middle Jurassic to Miocene. Deformed metavolcanic and related metasedimentary rocks occur as elongate pendants both within and along the margins of these plutonic complexes. Rocks within these pendants range from Triassic to Cretaceous, but cannot be directly tied to either Wrangellia or flanking terranes to the east. Lying east of this magmatic arc domain, along a complex array of both high-angle fault and intrusive contacts, is an imbricate stack of thrust nappes, here referred to as the Coast Belt-Northwest Cascade thrust system (CB-NCS). The foreland belt of this thrust system includes both unmetamorphosed and low grade metamorphic rocks of the Gambier Group and Cadwallader Terrane. The hinterland belt is underlain by metamorphosed equivalents of the Cadwallader Terrane and by Jura-Cretaceous metasedimentary and related plutonic rocks of the Cascade Metamorphic Core. Structurally above, and in fault contact with metamorphic core zone rocks of this central domain, are chaotically deformed oceanic rocks of the Bridge River-Hozameen Terrane and fault-bounded slivers of low grade sedimentary rocks, believed to be part of the dismembered Methow-Tyaughton Basin. The central and eastern domains of the Coast Belt orogen are cut by Tertiary right-lateral strike-slip faults of the Fraser-Straight Creek system. The offset continuations of these domains occur to the south in the central and eastern domains of the northern Cascades.

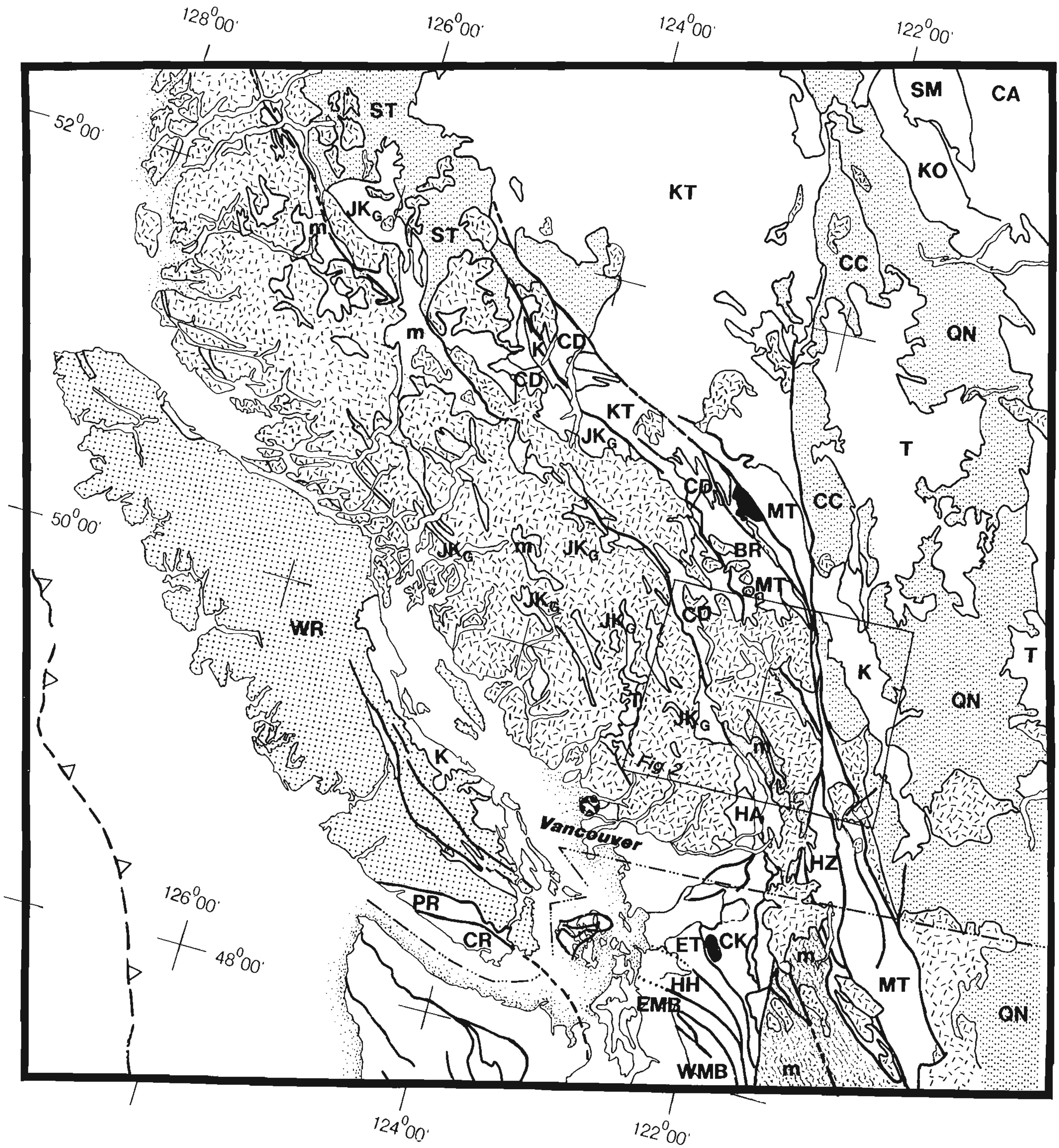

Figure 1. Terrane map of the southern Canadian Cordillera and adjacent Cascades. Terranes include Cassiar (CA), Slide Mountain (SM), Kootenay (KO), Quesnellia (QN), Cache Creek (CC), Stikinia (ST), Methow (MT), Bridge River (BR), Cadwallader (CD), Harrison (HA), Chilliwack (CK), Wrangellia (WR), Pacific Rim (PR), Crescent (CR), Western Mélange Belt (WMB), Eastern Mélange Belt (EMB), Helena Haystack (HH), and Easton terrane (ET). (m) denotes undifferentiated metamorphic rocks of the Cascade and Coast Belt metamorphic core zone. Overlap assemblages include the Gambier Group (JKG), and undifferentiated Cretaceous and Tertiary assemblages (K, KT, T). After Wheeler et al. (1989).

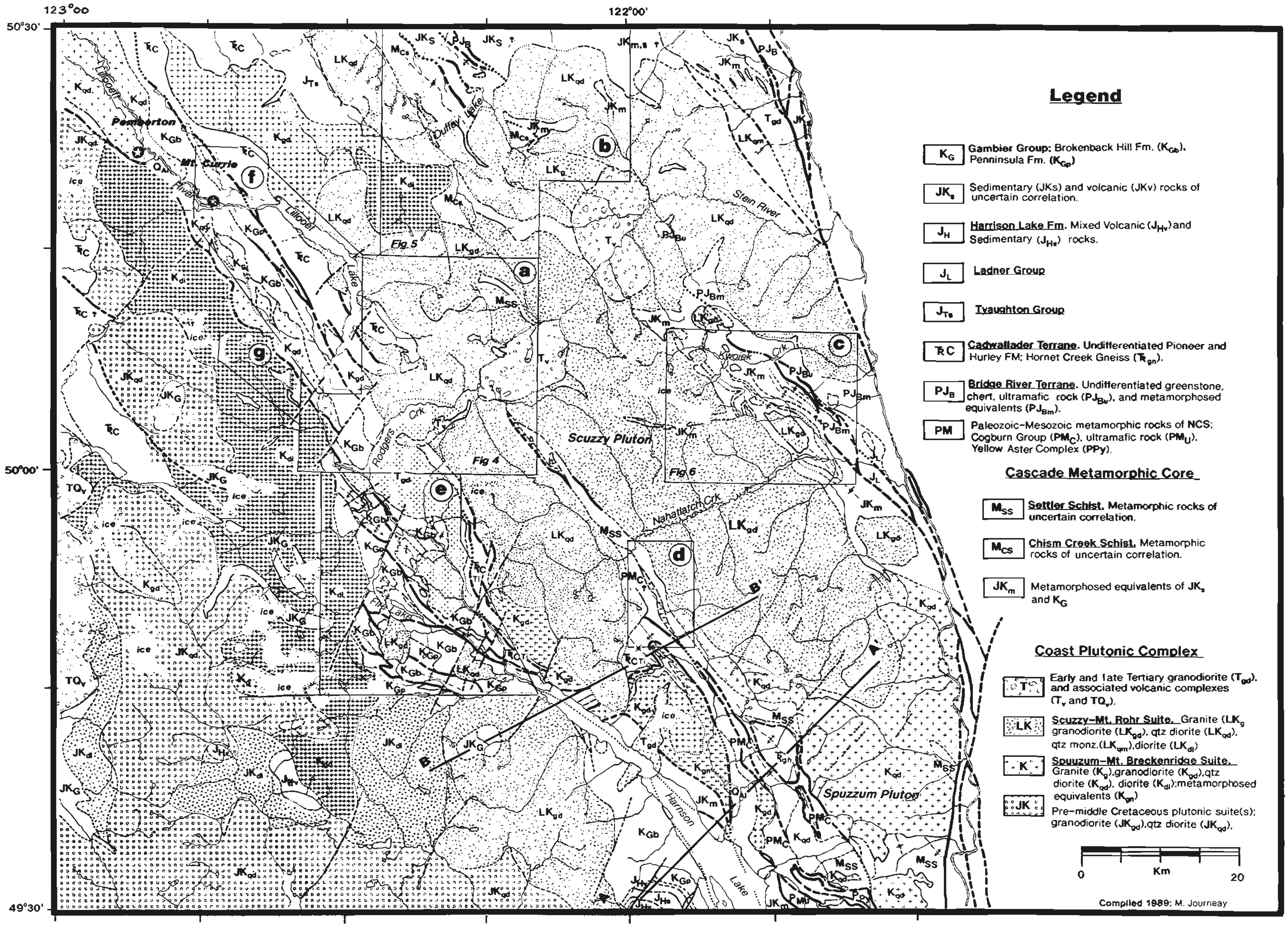

Figure 2. Geological compilation of the southeast Coast Belt. Areas of current research include the mapping results of (a,b,c,) this study, and (d) Tyson (1990); (e) Lynch (1990); (f) Riddell (1990) and (g) Freidman (1990). See text for discussion.

Gambier Group

The Gambier Group, as defined on the current terrane map of the Canadian Cordillera (Wheeler et al., 1989), is an overlap assemblage which unconformably overlies pre-Middle Jurassic terranes and plutonic suites of the southern Coast Belt. Along the southeast flank of the Coast Belt, the Gambier Group includes Upper Jurassic to upper Lower Cretaceous flows, tuffs and associated volcaniclastic conglomerates, sandstones and shales of the Mysterious Creek, Billhook Creek, Penninsula and Brokenback Hill Formations (Crickmay, 1925; Arthur, 1986). These rocks unconformably overlie an older volcanic arc complex comprising Lower to Middle Jurassic flows and related pyroclastic rocks of the Harrison Lake Formation, and are intruded by younger magmatic suites of the Coast Plutonic Complex.

The Fire Lake pendant of the Gambier Group (Roddick, 1965) lies in the foreland of the Coast Belt-Northwest Cascade thrust system (Fig. 2), and is para-autochthonous with respect to equivalent age rocks along the northwest shore of Harrison Lake. Lynch (1990) has refined the internal stratigraphy and structure of Gambier Group rocks within the Fire Lake pendant and has correlated them with autochthonous Penninsula and Brokenback Hill formations to the south.

Fault slivers of both the Penninsula and Brokenback Hill Formations extend northwestward from the Fire Lake pendant along the flank of the Pemberton plutonic suite (Journeay and Csontos, 1989; Riddell, 1990), and are probably part of a continuous belt of Gambier Group rocks that can be traced northward to the headwaters of the Lillooet River. Fault slivers mapped as Slollicum Schist along the northeast shore of Harrison Lake (Monger, 1986) are also considered to be part of the Gambier belt. Several of these fault slivers contain volcanic and related sedimentary sequences that closely resemble those of the Penninsula and Brokenback Hill Formations. Low grade felsic metavolcanic rocks from one of these slivers yield a U-Pb zircon age of 100 Ma (R.R. Parrish, pers. comm., 1989). These data support the view that rocks of the Gambier Assemblage in this region of the Coast Belt may have been part of a continuous intracratonic or back-arc basin that was later shortened and overridden by deeper tectonic levels of the accretionary complex in Mid- to Late Cretaceous time. The amount of shortening in this part of the thrust belt, and hence the initial width and tectonic setting of the Gambier basin are uncertain.

Cadwallader Terrane

The Cadwallader Terrane, as defined by Rusmore (1985), is a an exhumed Upper Triassic volcanic arc complex comprising basaltic flows, tuffaceous sandstones, conglomerates and turbidites and unconformably overlying Lower Jurassic (Upper Norian to Middle Bajocian) sedimentary rocks. The terrane is interpreted to be allochthonous with respect to North America, and to have been amalgamated with adjacent rocks of the Bridge River Terrane in Middle Jurassic time (Rusmore et al., 1988).

Rocks believed to be part of the Cadwallader Terrane have been identified as far west as Lillooet Lake (Woodsworth, 1977), where they are juxtaposed against the Gambier Assemblage along a system of high-angle reverse faults (Journeay and Csontos, 1989; Riddell, 1990). The faults which separate these two terranes can be traced with continuity to the south end of Lillooet Lake, where they are cut by post-kinematic intrusive suites of the Coast Plutonic Complex (Fig. 2, 3). Shear zone fabrics along both brittle and ductile strands of this fault system record a consistent northeast-side-up sense of displacement.

Hanging wall rocks within this part of the Cadwallader Terrane consist of a heterogeneous assemblage of metavolcanic greenstone, interlayered andesite and chert, volcaniclastic sandstone, phyllite and siltstone. Structurally overlying these lower grade assemblages are discontinuous lenses of well foliated amphibolite and metasedimentary rocks that are locally intruded by semiconcordant sheets of gneissic granodiorite. Both suites of rocks record a two-stage history of penetrative deformation. They are locally mylonitic and display discontinuous variations in metamorphic grade that are best explained by fault imbrication along the western margin of the Cadwallader Terrane.

Rock suites believed to be part of this disrupted Cadwallader Terrane (previously mapped as Twin Islands Group by Roddick, 1965), occur in fault slivers along the east flank of the Fire Lake pendant and can be traced into a mélange of high grade thrust nappes exposed along the east side of Harrison Lake (Journeay and Csontos, 1989; Tyson, 1990). Dominant lithologies in these fault slivers include clast-supported volcanic conglomerate (locally containing boulders of deformed quartz diorite), hornblende gneiss, intermediate and felsic metavolcanic flows and lesser volcaniclastic wackes, pelitic and semi-pelitic schists. Thrust sheets at the base of this imbricate stack consist primarily of talc-sericite and pelitic schists, and metavolcanic rocks from which preliminary U-Pb zircon analyses suggest a Middle Jurassic age of crystallization (R. Friedman, pers. comm., 1989). Together, these imbricate thrust nappes define a belt that is contiguous with and believed to be part of the Northwest Cascades System, as defined by Misch (1966), Lowes (1972), Monger (1986) and Brown (1987).

Cascade Metamorphic Core

Tectonic assemblages which have been mapped as part of the Cascade Metamorphic Core (CMC) in the southern Coast Belt include high grade metamorphic rocks of the NWCS (Twin Island Group) (Journeay and Csontos, 1989), polymetamorphic and related plutonic rocks of the Cogburn Group and Settler Schist (Lowes, 1972), and Buchan-type metamorphic rocks exposed along the flanks and within pendants of the Late Cretaceous Scuzzy pluton (Hollister, 1966; Monger, 1986).

The Cogburn Group lies structurally above rocks of the NWCS and is a mélange of thin-bedded schistose quartzite, metachert, semi-pelitic schist and amphibolite. Near the base of the mélange are discontinuous slivers of ultramafic rock and metagabbro, believed to be part of the Yellow Aster Complex (Lowes, 1972; Gabites, 1985; Monger, 1986). On the basis of these observations, it has been suggested that parts of the Cogburn Group melange may be correlative with the western Shuksan suite and Elbow Lake Formation, as defined by Brown (1987) and Gallagher et al. (1988) in Washington.

Rocks of the Settler Schist lie structurally above the Cogburn Group mélange and are both intruded by, and tectonically interleaved with, gneissic rocks of the mid-Cretaceous Spuzzum pluton. Dominant lithologies include thinly laminated schistose quartzite, graphitic and semi-pelitic schists of variable metamorphic grade, quartzofeldspathic gneiss, amphibolite and minor ultramafic rock. Pendants of Settler Schist occur throughout the Scuzzy pluton, and have been found as far north as Duffey Lake. These pendants are on strike with and closely resemble metamorphic rocks of the Chism Creek Schist as defined by Rusmore (1985). Regional correlations proposed for the Settler Schist include Late Triassic(?) rocks of the Chiwawkum Schist (Nason Terrane; Misch, 1966; Lowes, 1972) and/or Jurassic(?) rocks of the Eastern Shuksan suite (Shuksan blueschist and Darrington Phyllite; Monger, 1989).

Coast Plutonic Complex

Intrusive rocks along the southeast flank of the Coast Plutonic Complex can be subdivided into four major magmatic suites on the basis of regional field relationships and available geochronology. Pre-mid-Cretaceous plutons occur mainly west of the Coast Belt-Cascade thrust system and include Middle Jurassic and Late Jurassic-Early Cretaceous rocks of the Cloudburst and Pemberton plutonic suites (Fig. 2).

Middle Cretaceous suites are distributed throughout the southern Coast Belt and are Albian to Cenomanian. Plutons west of the Coast Belt-Cascade thrust system are only locally deformed (Monger, 1990) and include parts of the Squamish-Porteau Cove granodiorite and Pemberton diorite complex. Equivalent plutonic suites in adjacent parts of the Cascade Metamorphic Core are syn- to late-kinematic with respect to major fold and fault structures of the CB-NWC thrust system, and include both foliated and nonfoliated granodiorite of the Mt. Breckenridge and Spuzzum plutonic complexes. Recent geochronometry suggests that these two plutonic complexes may be part of a composite intrusive suite ranging in age from 100-91 Ma (R.R. Parrish and R. Friedman, pers. comm., 1989).

Late Cretaceous intrusive rocks of the Scuzzy and Mt. Rohr plutonic complexes are part of a continuous magmatic belt that appears to be localized along the east flank of the Coast Belt orogen. Plutons of this intrusive suite are post-kinematic with respect to major northwest-trending fault structures, and cut across both foreland and hinterland belts of the CB-NWC thrust system. U-Pb analyses of zircons from the central and eastern parts of the Scuzzy plutonic complex indicate a Santonian-Early Campanian age of crystallization for this part of the intrusive suite (R.R. Parrish and R. Friedman, pers. comm., 1989). K-Ar cooling dates on both hornblende and biotite are 79-70 Ma (Gabites, 1985; R.R. Parrish, pers. comm., 1989) and imply a period of rapid cooling and possible unroofing of the Coast Belt during Campanian and early Maastrichtian time.

Cutting across the Late Cretaceous magmatic arc is a composite suite of late Tertiary plutonic and related volcanic rocks, the Chilliwack-Pemberton suite. Plutonic rocks of Oligocene and Early Miocene age make up the older component of this suite and include both medium- and coarse-grained granodiorite of the Doctors Point pluton (R.R. Parrish, pers. comm., 1989) and related unnamed plugs along the east side of Harrison Lake (Gabites, 1985; Fig. 2). Younger components of this suite include middle Miocene granodiorite of the Rogers Creek pluton and related(?) volcanic rocks of the Pemberton Belt. These rocks are locally deformed by high-angle brittle fault zones, and provide a record of recent tectonic activity in the southern Coast Belt.

Structural framework

Overview

The structural grain and tectonic fabric of the southeast Coast Belt is defined by a mid-Cretaceous to Early Eocene network of northwest-striking faults and related fold structures (Fig. 2). From west to east, these include imbricate thrust nappes and high-angle reverse faults of the Coast Belt-Northwest Cascade System (CB-NCS), ductile and brittle strike-slip faults of the Harrison Lake Shear Zone (HSZ), dip-slip faults of the Duffey Lake-Kwoiek Creek region and imbricate strike-slip and normal faults of the Yalakom-Marshall Creek system(s). These structures are cut by major right-lateral transcurrent faults of the Fraser-Straight Creek system, and by a network of late Tertiary oblique-slip and related high-angle faults that appear to be widely distributed throughout the southern Coast Belt.

Coast Belt-Northwest Cascade System (CB-NCS)

The CB-NCS comprises a set of imbricate thrust nappes and high-angle reverse faults which separate weakly deformed and low grade metamorphic rocks of the central Coast Belt from deeper crustal levels of the adjacent Cascade Metamorphic Core (Fig. 2). The fault system includes structures previously mapped along the east side of Harrison Lake by Crickmay (1930), Lowes (1972), Reamsbottom (1971, 1974) Gabites (1985) and Monger (1986), as well as faults and related fold structures along strike to the north in the Lillooet River valley (Roddick, 1965; Woodsworth, 1977; Journeay and Csontos, 1989). The fault system is believed to extend southward into the Cascades of Washington (Lowes, 1972; Monger, 1986), where it was initially described by Misch (1966).

Imbricate thrust faults and related fold structures involve low grade rocks of the Gambier Group, Cadwallader Terrane and higher grade equivalents within the Cascade Metamorphic Core. Individual fault strands are marked by localized zones of high strain in which mylonitic foliations and associated down-dip stretching lineations are well developed. Kinematic indicators within these fault strands and asymmetric fold structures in adjacent thrust sheets both record an upper plate to the southwest sense of shear (Journeay and Csontos, 1989) and are consistent with a model of thrust imbrication as first proposed by Crickmay (1930).

These structures are in turn folded, and cut by a set of high-angle reverse faults in which kinematic indicators record a northeast-side-up sense of shear. Geometrical relationships between early and late structures (Fig. 3) imply a two-stage history of imbrication (see also Tyson, 1990). It is, however, not clear whether these faults evolved episodically, or as part of a continuous sequence within an evolving accretionary complex.

Isotopic studies of both syn- and post-kinematic plutons (Gabites, 1985; R.R. Parrish and R. Friedman, pers. comm., 1989) indicate that fault strands along the west side of the thrust system were active between 91-86 Ma. Faults which imbricate the Cogburn Group and related ultramafic rocks of the Yellow Aster(?) Complex are cut by small plugs west of the main body of the Spuzzum pluton, which are reported to be as old as 110 Ma (Gabites, 1985). Geometric and kinematic relationships between these two sets of faults are uncertain.

Harrison Lake Shear Zone (HLZ)

Faults which define the Harrison Lake shear zone include a network of ductile mylonite zones along the east shore of Harrison Lake (south of Big Silver Creek), and brittle strike-slip faults exposed along the west and north shores of Harrison Lake (Monger, 1986; Journeay and Csontos, 1989). These structures cut thrust-related fabrics in metavolcanic rocks as young as 100 Ma, and record a consistent right-lateral sense of displacement (Journeay and Csontos, 1989; Talbot, 1989).

Other dextral strike-slip faults which may be part of the Harrison Lake system include the Big Silver Creek fault (Journeay and Csontos, 1989) and localized ductile shear zones within the Breckenridge gneiss complex (Talbot, 1989). Synkinematic muscovite from the Big Silver Creek fault yields a K-Ar date of 93.5 Ma (R.R. Parrish, pers. comm., 1989) and implies that strike-slip faulting may have

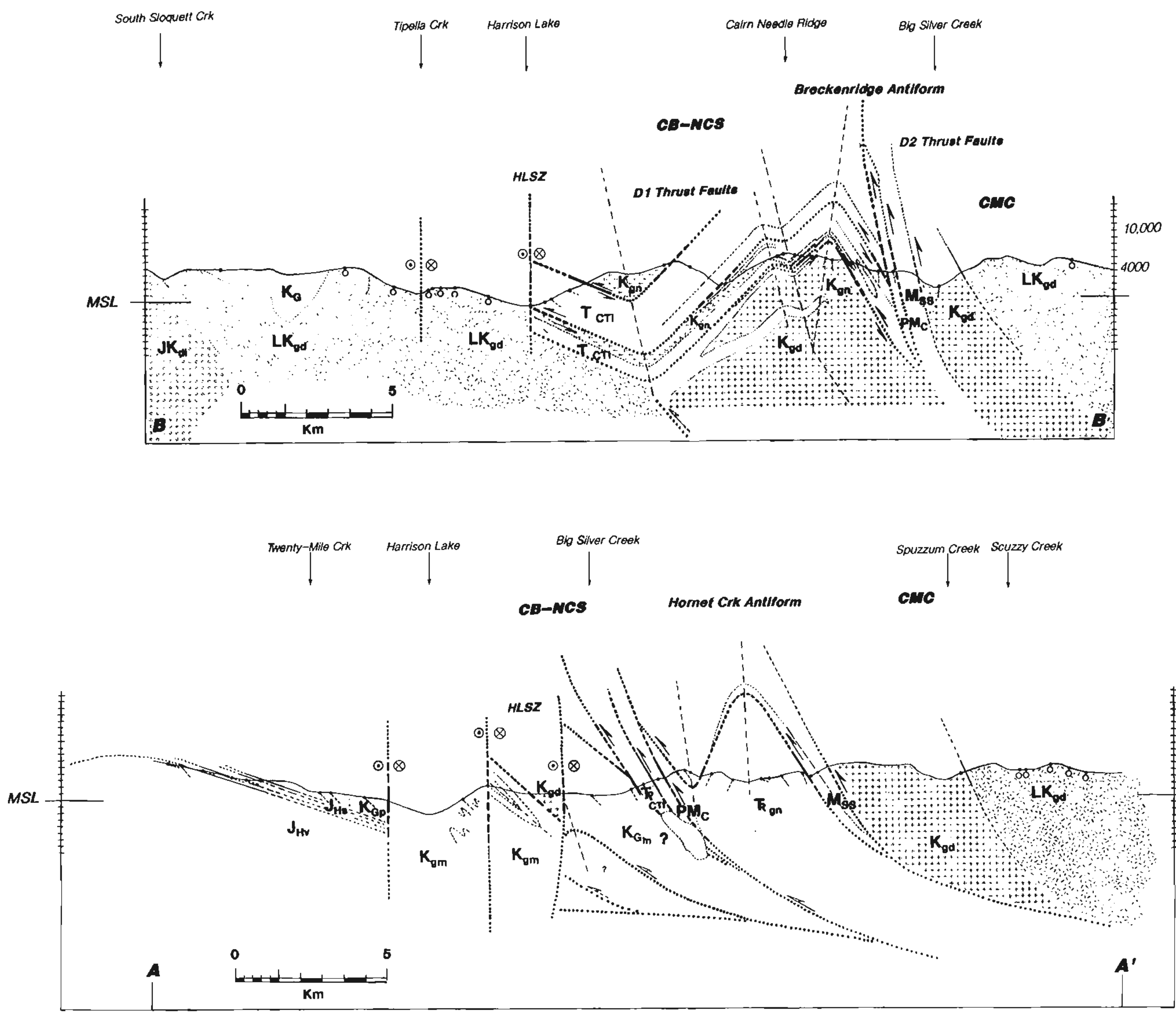

Figure 3. Geological cross-sections of the Coast Belt-Northern Cascade Thrust Belt (CB-NCS). See Figure 2 for legend and location of cross-sections.

been coeval with thrust imbrication, at least within this part of the Coast Belt. Right-lateral ductile shear zones occur throughout the northern Cascade orogen (Brown, 1987), and may also be part of the Harrison Lake Fault System.

Bralorne-Kwoiek Creek System

Fault strands of the Bralorne-Kwoiek Creek system mark the western boundary of the Bridge River-Hozameen Terrane along the east flank of the Cascade Metamorphic Core. The system is well-defined in the Bralorne and Kwoiek Creek areas, but is obscured south of Duffey Lake by undeformed Late Cretaceous plutons of the Scuzzy-Mt. Rohr intrusive suite. Correlation of these fault strands, as implied by Woodsworth (1977) and proposed by Rusmore (1985) and Monger (1986), is based on relative timing arguments and similarities in structural style in both the Bralorne and Kwoiek Creek regions.

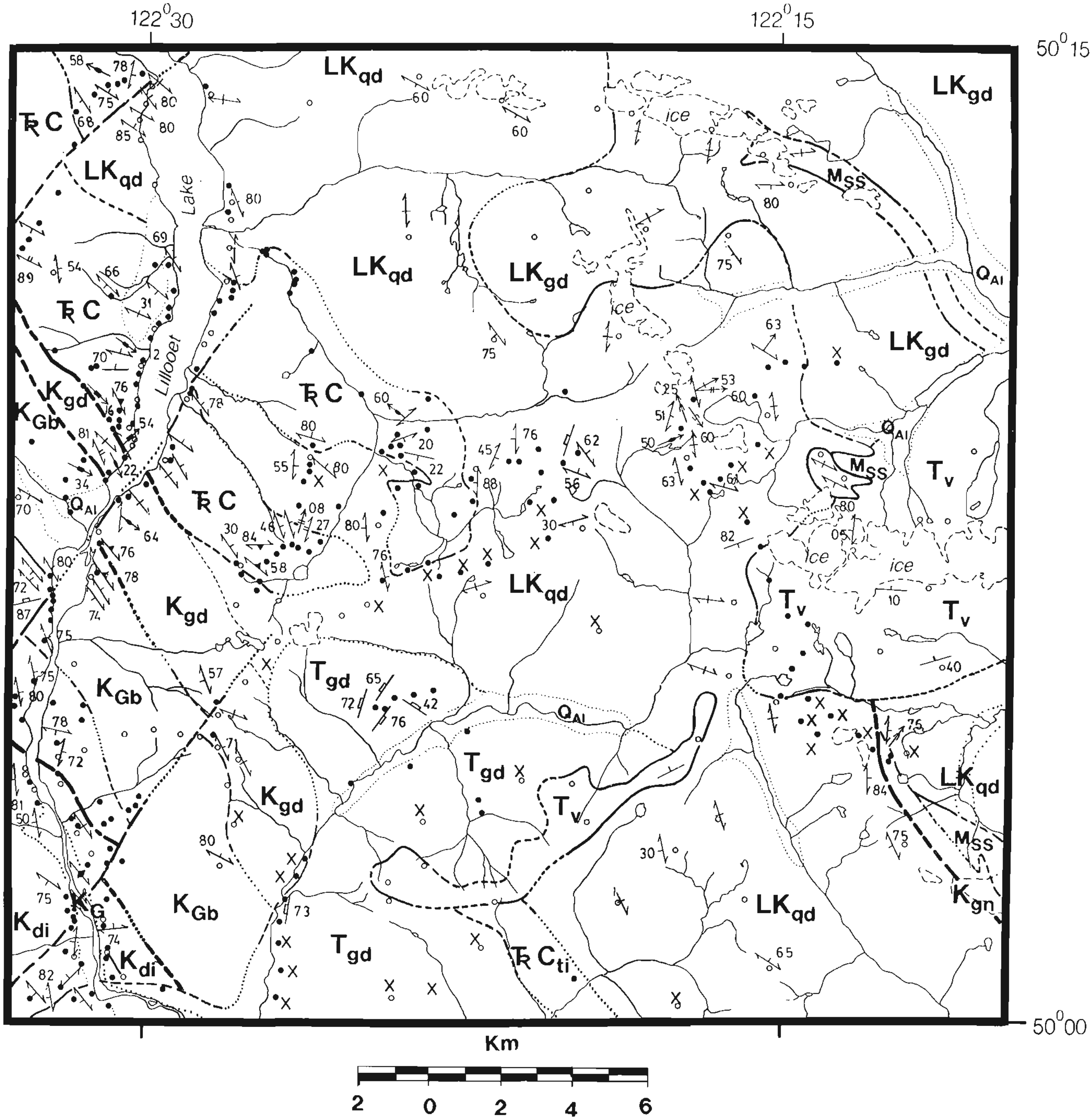

Figure 4. Geological map of the Rodgers Creek region. Open circles are station locations and data of Roddick and Hutchison (1973) and Woodsworth (1977). Filled circles are station locations and data of this study. See Figure 2 for legend and location.

Fieldwork during 1989 focused on the geometry and kinematics of fault structures in the Duffey Lake region, and included reconnaissance mapping of the Kwoiek Creek Fault, as redefined by Monger (in press a,b). Preliminary results of this mapping (Fig. 4, 5) support proposed correlations of the Bralorne-Kwoiek Creek faults, and are consistent with the observations of Rusmore (1985).

Two sets of regional fold structures and related fabrics are recognized in domains bounded by the Duffey Lake-Kwoiek Creek faults, and within footwall rocks to the west (Fig. 4, 5). Both sets of fold structures verge toward the southwest, and are similar in style to those described in the foreland zone of the CB-NCS (Fig. 3). Fault strands in both the Duffey Lake and Kwoiek Creek regions dip steeply to the northeast (50-80°), and cut the overturned limbs of both early and late generation folds. Fault zones are riddled by a network of scaly serpentinite and talc-carbonate schist, and typically contain a melange assemblage of chaotically deformed ribbon chert, greenstone, marble and hydrothermally altered ultramafic rock. They are characterized by localized zones of brittle and ductile deformation (10-50 m wide) in which flattening foliations, noncylindrical folds and down-dip stretching lineations are well developed.

The orientation of reclined fold hinges and stretching lineations within individual fault strands indicates a history of dip-slip displacement. Shear sense indicators observed at

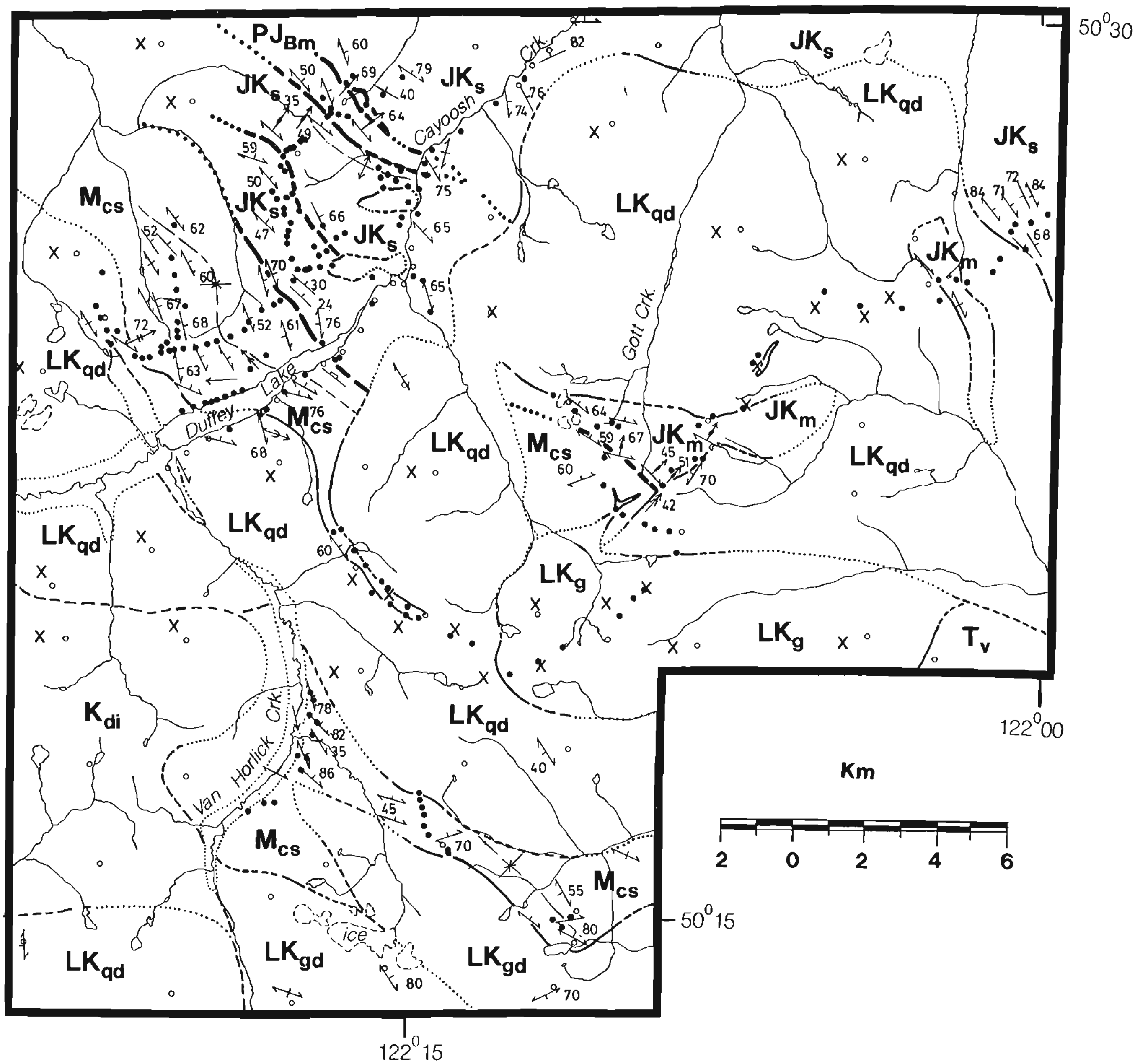

Figure 5. Geological map of Duffey Lake region. Open circles are station locations and data of Roddick and Hutchison (1973) and Woodsworth (1977). Filled circles are station locations and data of this study. See Figure 2 for legend and location.

the mesoscopic scale are inconclusive and too few to be used with confidence. However, all fault zones are sharply defined at their base with hanging wall gradients in both flattening and rotational strain that consistently decrease upsection, away from the basal detachment. Juxtaposition of a highly strained hanging wall against a relatively less deformed footwall implies that bounding faults most likely ramped upsection to the southwest from deeper tectonic levels, perhaps as part of an imbricate thrust system.

Arguments presented by Rusmore (1985) suggest that faults of the Bralorne-Kwoiek Creek System were active between 100 and 85 Ma. The system locally includes fault slivers of the Taylor Creek Group (Albian-Cenomanian) and is cut by post-kinematic plutons that yield concordant U-Pb zircon dates of 84-86 Ma (Rusmore, 1985; R.R. Parrish, pers. comm., 1989). To the south, the Bralorne-Kwoiek Creek fault is cut by northerly-trending strands of the Fraser River strike-slip system and is presumed to have been displaced some 80-110 km southward into the Cascades of northern Washington during in the Early Tertiary.

Late Tertiary Fault Systems

Faults of the southeast Coast Belt that are believed to have been active in late Tertiary time include a system of northeast-striking transcurrent faults, and related(?) high-angle reverse and normal faults. Northeast-striking transcurrent faults are well exposed in the Harrison Lake-Lillooet Lake region, and record components of both right-lateral strike-slip and oblique-slip displacement (Journeay and Csontos, 1989; Lynch, 1990). The largest of these structures (Glacier Creek Fault, Fig. 2) offsets pre-existing faults of the Northwest Cascade System by as much as 4 km (Lynch, 1990), and is cut by Miocene granodiorite of the Rogers Creek pluton (16 Ma-K-Ar; Woodsworth, 1977). The pluton is itself cut by a network of north-and northeast-striking transcurrent faults that record only minor amounts of displacement (Fig. 3).

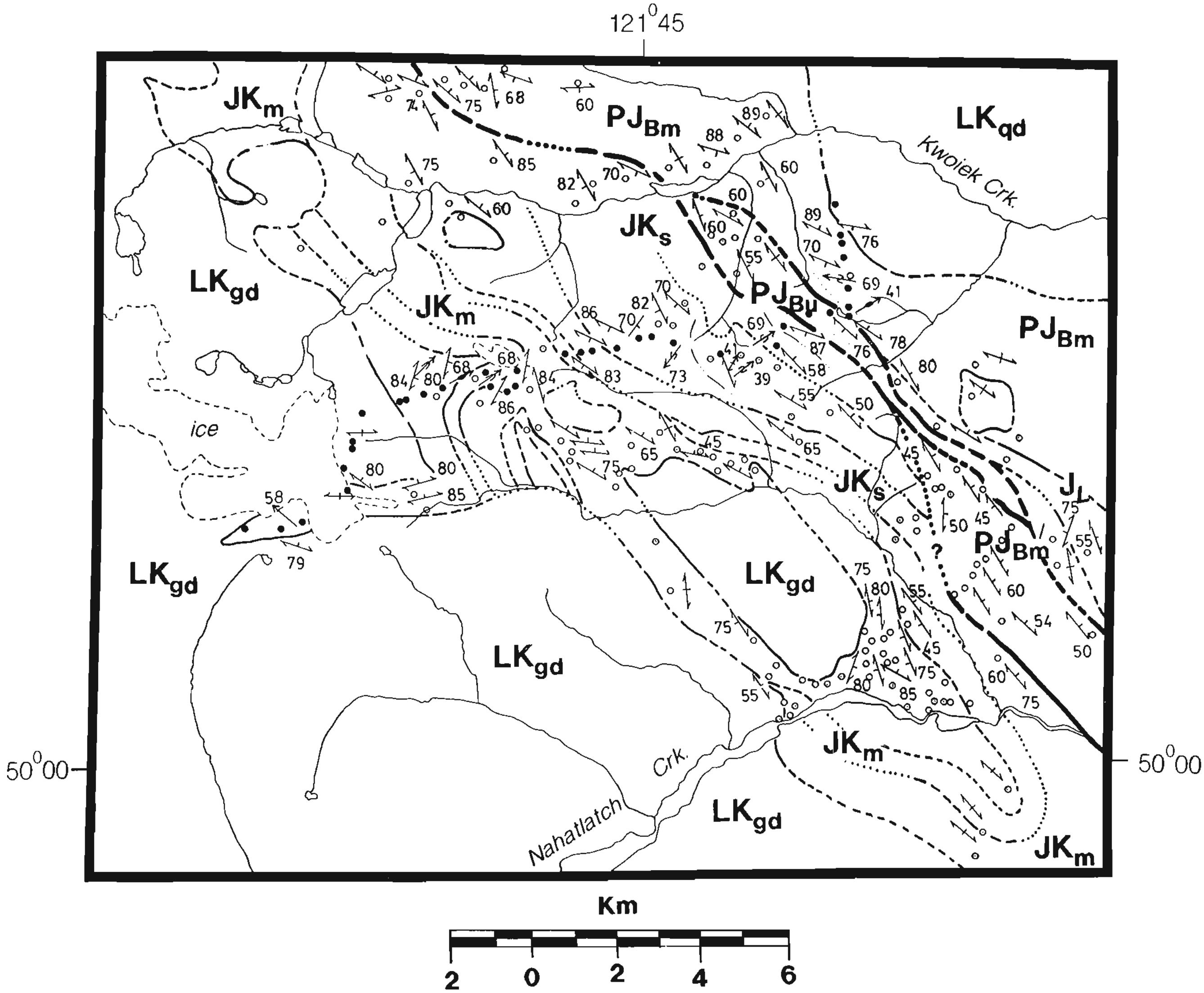

Figure 6. Geological map of Kwoiek Creek region. Open circles are station locations and data of Hollister (1966) and Monger (1986). Filled circles are station locations and data of this study. See Figure 2 for legend and location.

These late Tertiary faults acted as conduits for the emplacement of Miocene intrusive breccias and related alkalic volcanic complexes of the Pemberton magmatic belt, and appear to have been important in localizing associated hydrothermal systems (Journeay and Csontos, 1989). Spatially associated with these structures are northwest-striking reverse faults. The timing and kinematics of both transcurrent and high-angle reverse faults suggest that these structures may be part of a regional system that formed along the continental margin in response to northeast-southwest shortening. If so, these faults may provide a means of evaluating the crustal response to eastward underthrusting of oceanic crust since middle Tertiary time, and recent patterns of uplift and tectonic activity within the southern Coast Belt.

History of metamorphism

Metamorphism along the southeast flank of the Coast Belt can be linked directly to imbrication and eastward telescoping of terranes during mid-Cretaceous evolution of the Northwest Cascade Thrust system (M_1), and to the emplacement of Late Cretaceous plutons during uplift and/or unroofing of the Cascade Metamorphic Core (M_2).

M_1 metamorphism

Early stage M_1 metamorphism is characterized by Barrovian mineral assemblages that range in grade from lower greenschist facies to middle and upper amphibolite facies (Reamsbottom, 1971, 1974; Lowes, 1972; Pigage, 1976; Bartholomew, 1979; Gabites, 1985). These M_1 assemblages are best developed in imbricated fault slivers along the west flank of the Cascade Metamorphic Core, and in isolated pendants of gneissic rock within late- and post-kinematic plutons to the north.

Matrix chlorite, muscovite, biotite and feldspar are aligned with early S_1 flattening foliations and are locally crenulated by younger generation F_2 folds. Porphyroblastic garnet, staurolite, hornblende, kyanite and sillimanite are syn- and late-kinematic with respect to the dominant schistosity, and locally record significant post-metamorphic strain. Inclusions of chlorite, biotite and andalusite within these porphyroblasts occur in fault slivers east of Harrison Lake, and imply an earlier history of crystallization that may either be part of the same prograde event, or a relic of an older history of metamorphism.

M_2 metamorphism

M_2 metamorphism is post-kinematic with respect to structures and associated fabrics of the CB-NCS and Bralorne-Kwoiek Creek fault systems, and is spatially associated with Late Cretaceous plutons of the Scuzzy-Mt. Rohr intrusive suite. M_2 assemblages that occur in thrust slivers along the west flank of the Cascade Metamorphic Core include late-stage overgrowths and porphyroblasts of andalusite, sillimanite, garnet and hornblende. Equivalent phases in graphitic schists along the north and east flanks of the Scuzzy pluton include porphyroblastic garnet, staurolite and andalusite. In the vicinity of Kwoiek Needle, M_2 assemblages record a history of Buchan-type metamorphism in which metamorphic grade decreases away from the main body of the Scuzzy pluton (Hollister, 1966). The sequence and spacing of M_2 assemblage zones (staurolite-garnet-andalusite-kyanite-sillimanite) implies a nearly isobaric field gradient, and suggests that emplacement of the Scuzzy pluton and associated M_2 metamorphism most likely occurred at intermediate crustal levels (about 10-12 km; Hollister, 1966).

Conditions of metamorphism

"Peak" metamorphic conditions were estimated with conventional garnet-biotite and garnet-plagioclase thermobarometers using mineral composition data of Reamsbottom (1974), Pigage (1976) and Bartholomew (1979). Preliminary results of these calculations are summarized in Figure 7.

Assemblages in fault slivers east of Harrison Lake (Reamsbottom, 1974) record pressures of 7.4-9.6 kb and temperatures of 675-750°C. Equivalent rocks of the Yale Creek region (Pigage, 1976) record pressures of 6.9-8.0 kb and temperatures of 600-635°C. These data require crustal loads of about 20-30 km during peak conditions of metamorphism and are consistent with a model of mid-Cretaceous thrust imbrication and associated tectonic burial, as proposed by Lowes (1972) and Journeay (1989).

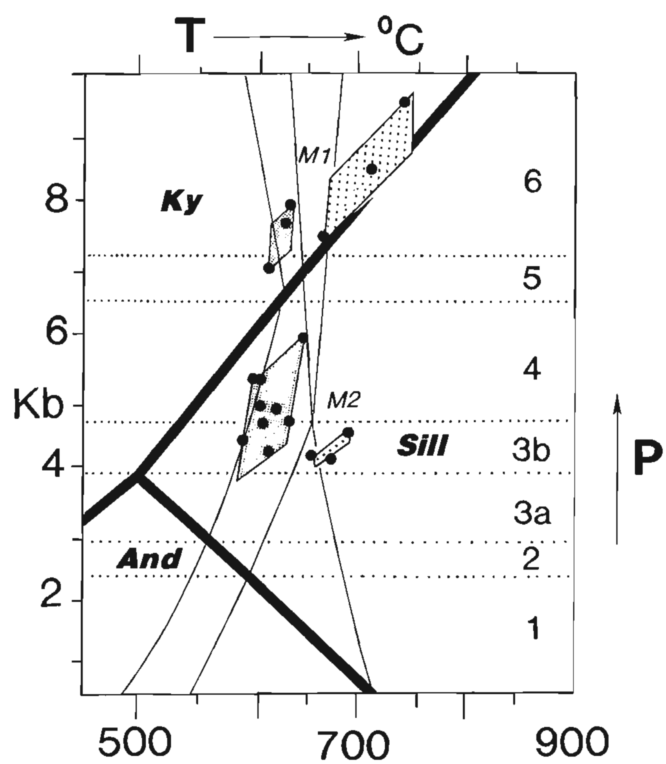

Figure 7. Idealized PT phase diagram for minerals of the model pelitic system and results of thermobarometry. Heavy stipple are results from Yale Creek region, based on mineral composition data of Bartholomew (1979) and Pigage (1976). Light stipple are results from Mt. Breckenridge area, based on mineral composition data of Reamsbottom (1974).

Assemblages that occur adjacent to post-kinematic plutons in the Mt. Breckenridge area (Reamsbottom, 1974), and within sillimanite-bearing rocks of the Yale Creek region (Bartholomew, 1979), record ''peak'' metamorphic conditions of 4.5-6.1 kb at 660-690°C and 3.7-6.0 kb at 575-650°C, respectively. These data are consistent with the results of Hollister (1966), and suggest that the emplacement of Late Cretaceous plutonic suites and associated M_2 metamorphism within this part of the Coast Belt occurred at intermediate crustal levels (15-20 km), presumably during uplift and/or unroofing of the Cascade Metamorphic Core.

DISCUSSION

Similarities in structural style, timing and histories of deformation suggest that the CB-NCS and Bralorne-Kwoiek Creek Fault zones may represent different crustal levels of a regional fold and thrust belt system that evolved as part of a westward prograding accretionary complex in mid- to late Cretaceous time. Early stages in the evolution of this accretionary complex may have included thrust imbrication of terranes and flanking depositional basins situated between the Intermontane and Insular belts (Bridge River, Cadwallader and possibly Tyaughton groups). The imbrication and stacking of these terranes is interpreted to have resulted in significant crustal shortening, tectonic thickening and the onset of high grade Barrovian metamorphism within the core of the Coast Belt-Cascade orogen. Juxtaposition of these high grade tectonic slivers against lower grade rocks of the adjacent foreland is explained by thick-skinned imbrication and southwestward overthrusting of the Cascade Metamorphic Core over flanking syn-orogenic basin sequences of the Gambier Group, and underlying basement rocks of the Insular Terrane. Oversteepening and backfolding of this thrust belt during southwestward ramping along the buttress of the Insular Terrane may account for observed variations in metamorphic grade within the Cascade Metamorphic Core, as well as apparent infrastructure suprastructure transitions across strike of the orogen. Right-lateral strike-slip faults (e.g. HLZ) and equivalent structures of the northern Cascades may have been important in partitioning components of orogen-parallel displacements during this stage in the evolution of the accretionary complex.

Pre-Late Cretaceous thrust faults along the southeast flank of the Coast Belt are cut by the Fraser River-Straight Creek System (Monger, 1986), across which there is an estimated 80-110 km of Paleocene-Early Eocene displacement (Monger, 1986; Kleinspehn, 1985). These estimates are based on a restoration of the Bridge River-Hozameen Terrane and Albian(?) depositional systems of the Tyaughton-Methow basin across the Fraser Fault, and imply correlation of the Bralorne-Kwoiek Creek and Ross Lake Fault zones (Rusmore, 1985). This correlation is tantalizing from a geometrical point of view, but is complicated by the fact that these two fault systems record very different displacement histories and are not the same age. A more reasonable counterpart for the Bralorne-Kwoiek Creek Fault might be the Jack Mountain Thrust (Misch, 1966), a pre-Late Cretaceous low-angle fault which separates the Hozameen Terrane from underlying metamorphosed rocks of the Jack Mountain Phyllite. Direct correlation of the Bralorne-Kwoiek Creek and Jack Mountain Faults is obscured by early Tertiary displacement across the Ross Lake Fault zone, and thus do not constitute a testable hypothesis. The relationships between Northwest Cascade/Bralorne-Kwoiek Creek Fault systems and equivalent age structures within adjacent parts of the of the central Coast Belt are uncertain.

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Geology of the Fire Lake Group, southeast Coast Mountains, British Columbia

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Lynch, J.V.G. Geology of the Fire Lake Group, southeast Coast Mountains, British Columbia; in Current Research, Part E, Geological Survey of Canada, Paper 90-1E, p. 197-204, 1990.

Abstract

The Fire Lake Group includes the Lower Cretaceous Peninsula and Brokenback Hill formations. The Peninsula is a fining upwards sequence, with trough cross-stratified fluvial conglomerate and coarse marine beach deposits at the base, succeeded by arkose and slate. The overlying Brokenback Hill Formation is mainly volcanic. It progresses upwards from feldspar crystal tuff, to andesite flows, breccia, and heterolithic volcanic conglomerate, to volcaniclastic sandstone, and is topped by welded pyroclastic deposits.

Three phases of deformation are recorded. The earliest is characterized by shallow angle south-southeast-directed thrusting, emplacing rocks of the Peninsula Formation onto the Brokenback Hill Formation. The second phase features tight, large amplitude noncylindrical folds, in association with southwest-directed high-angle thrusting. Ductile shearing, and the assemblage of talc-kyanite indicate conditions of high temperature and pressure. Late deformation consists of northeast-striking dextral-normal dip-slip block faulting.

Résumé

Le groupe de Fire Lake comprend les formations de Peninsula et de Brokenback Hill du Crétacé inférieur. La formation de Peninsula est une série dont les éléments présentent des grains de plus en plus fins vers le haut, avec un conglomérat fluviatile de dépression à stratification oblique et des dépôts marins de plage grossiers, à la base suivis d'une arkose et d'une ardoise. La formation sus-jacente de Brokenback Hill est surtout volcanique. Elle passe du bas vers le haut de tufs à cristaux de feldspath, à des coulées, une brèche et un conglomérat volcanique hétérolithique, des grès volcanoclastiques, pour terminer par des dépôts pyroclastiques soudés.

On trouve trois phases de déformation. La plus ancienne est caractérisée par des chevauchements à faible pendage, de direction sud-sud-est amenant des roches de la formation de Peninsula sur la formation de Brokenback Hill. La seconde est caractérisée par des plis serrés, non cylindriques, de grande amplitude, associés à des chevauchements à fort pendage de direction sud-ouest. Un cisaillement ductile et l'assemblage du talc-cyanite indiquent des conditions de haute température et de forte pression. La dernière phrase de déformation est constituée d'un morcellement par failles à déplacement normal, dextres, de direction nord-est.

INTRODUCTION

This work was undertaken to provide detailed data on the Fire Lake Group, which occurs as one of a scattered series of Jurassic-Cretaceous pendants in the southern Coast Mountains (Fig. 1). The nature of the depositional environment for strata in many of these pendants and the correlation between units, are uncertain. The study also gives data on the setting of several known gold occurrences, such as the Money Spinner, Barkoola, and Blue Lead veins, and will help to assess the potential for volcanogenic massive sulphide mineralization.

The Fire Lake Group outcrops about 100 km northeast of Vancouver (Fig. 1) between the Lillooet River and the southeast margin of Garibaldi Provincial Park. It occupies half of the Glacier Lake 1:50 000 map area (92G/16).

Regional mapping by Roddick (1965) delineated the Fire Lake Group. Paleontological work of Jeletzky (1965) established an Early Cretaceous age for fossils collected by Roddick and indicated their equivalence to the Peninsula Formation of Crickmay (1925, 1930) immediately southeast along the western flank of Harrison Lake (Fig. 1). Further regional mapping, structural analysis, and synthesis were presented by Journeay and Csontos (1989), who identified and characterized many of the faults in the region. Details of the mineralization in the group are given by Ray and Coombes (1985).

Jeletzky (1965) also correlated the Early Cretaceous Peninsula and Brokenback Hill formations with the Nooksack Group of northern Washington on the distribution of the bivalve *Buchia*. The Fire Lake Group was correlated with the Gambier group on lithological similarities by Roddick (1965). Collectively these rocks are included in what is termed the Nooksack tectonostratigraphic terrane (Monger and Berg, 1987). These rocks are widely regarded as part of a broad Upper Jurassic-Lower Cretaceous overlap assemblage (Wheeler and McFeely, 1987), linking Wrangellia in the west with Stikinia to the east by latest Early Cretaceous time.

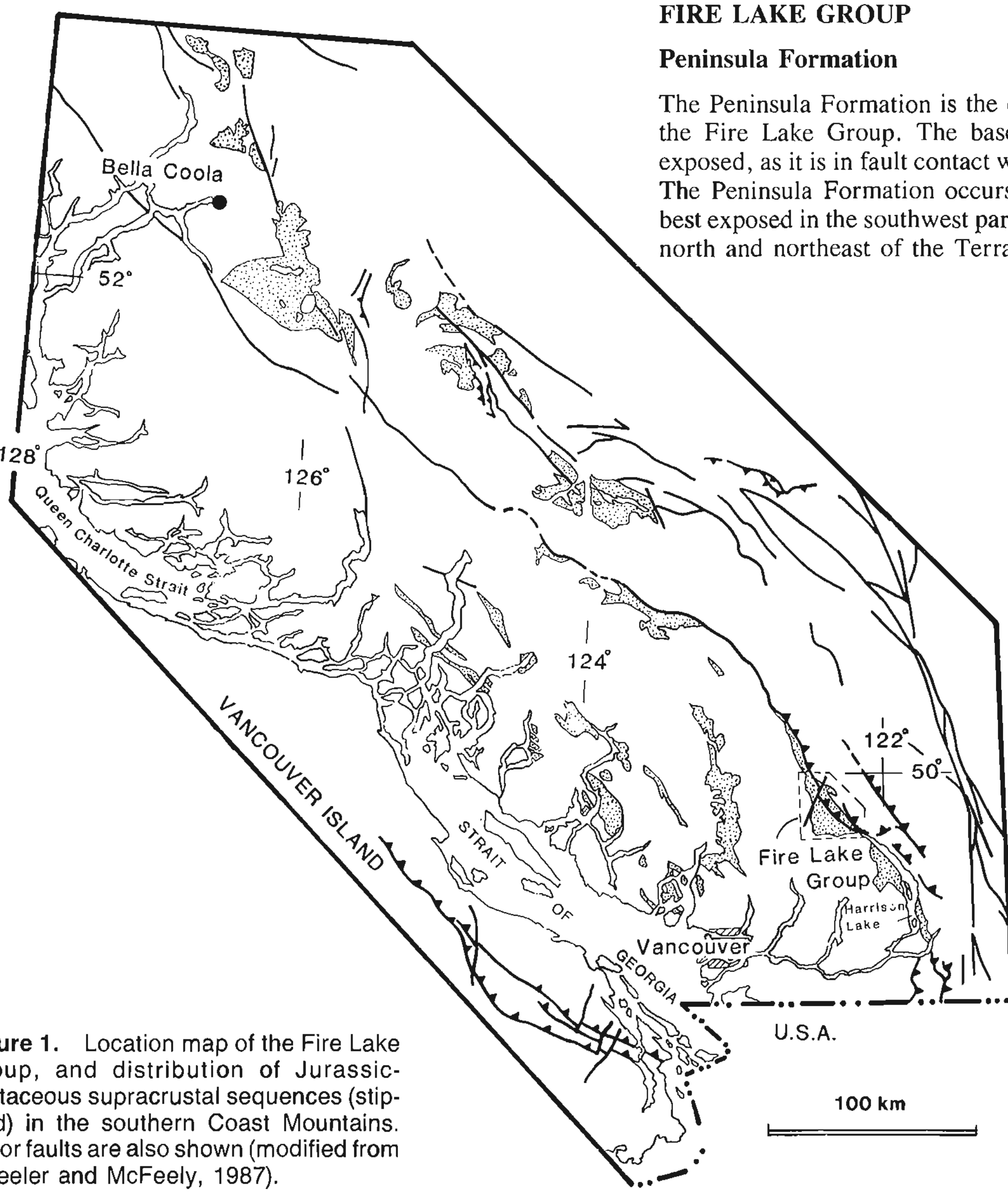


Figure 1. Location map of the Fire Lake Group, and distribution of Jurassic-Cretaceous supracrustal sequences (stippled) in the southern Coast Mountains. Major faults are also shown (modified from Wheeler and McFeely, 1987).

FIRE LAKE GROUP

Peninsula Formation

The Peninsula Formation is the oldest stratigraphic unit in the Fire Lake Group. The base of the formation is not exposed, as it is in fault contact with the surrounding rocks. The Peninsula Formation occurs over a wide area, but is best exposed in the southwest part of the Fire Lake pendant, north and northeast of the Terrarosa Glacier (Fig. 2).

Two members are distinguished. The lower is conglomerate; the upper consists of interbedded arkose and pyritic slate. The conglomerate is about 1200 m thick; clast type and distribution vary. Clasts include andesite, rhyolite, and feldspar porphyry, as well as chert, jasper, or quartzite, siltstone, detrital quartz and feldspar crystals, and a few granite clasts. Beds are generally thick, moderately to well sorted with well rounded pebble to cobble sized fragments, in a clast-supported framework. The stratigraphically lowest parts of the formation contains trough cross-stratified channel gravel and sand cemented by calcite, apparently deposited in a fluvial environment. Coal seams have been found but are small and rare. Conglomerate passes upwards into coarse beach deposits with fauna and flora more typical of the near shore marine environment; these include *Buchia,* pelecypods and belemnites, plant fossils, and tubular, bedding-parallel worm burrows.

The overlying arkose member is approximately 800 m thick. It is typically well bedded, and locally displays planar cross-stratification, hummocky cross-lamination, graded bedding, and soft sediment deformation features such as small normal faults where coarse sands occur above silty layers. Limestone beds or epidotized calc-silicate rock are occasionally found. The Peninsula Formation is marked by an upwards fining cycle; presumably it represents a general transgression.

Further work is required to identify fossils collected from new sites. However, Jeletzky (1965) identified *Buchia uncitoides* and *Buchia pacifica* from the Peninsula Formation of the Fire Lake Group, indicating a Late Berriasian to mid-Valanginian age. Fossils collected in 1979 by G.J. Woodsworth from the southwest part of the area were identified by J.A. Jeletzky as being of probable Hauterivian-Barremian age (G.J. Woodsworth, pers. comm., 1989). The basal member of the formation is apparently not exposed in this area. It typically has a high proportion of granite clasts and extends to the Lower Berriasian (Arthur, 1986).

Brokenback Hill Formation

The Brokenback Hill Formation within the Fire Lake Group is subdivided into four members; these are mostly volcanic and distinct from the sedimentary succession of the Peninsula Formation. The subdivisions and sequence of rocks in the formation are similar to that represented in the stratigraphic section of Arthur (1986), established for the type locality and surrounding area along the northwest flank of Harrison Lake. The correlation is based entirely upon the similarities in lithology, as no fossils have been collected from this formation in the Fire Lake area. Age determinations made to the south based on fossils collected from different parts of the formation, indicate Late Valanginian, Early Hauterivian, and Middle Albian ages (Arthur, 1986).

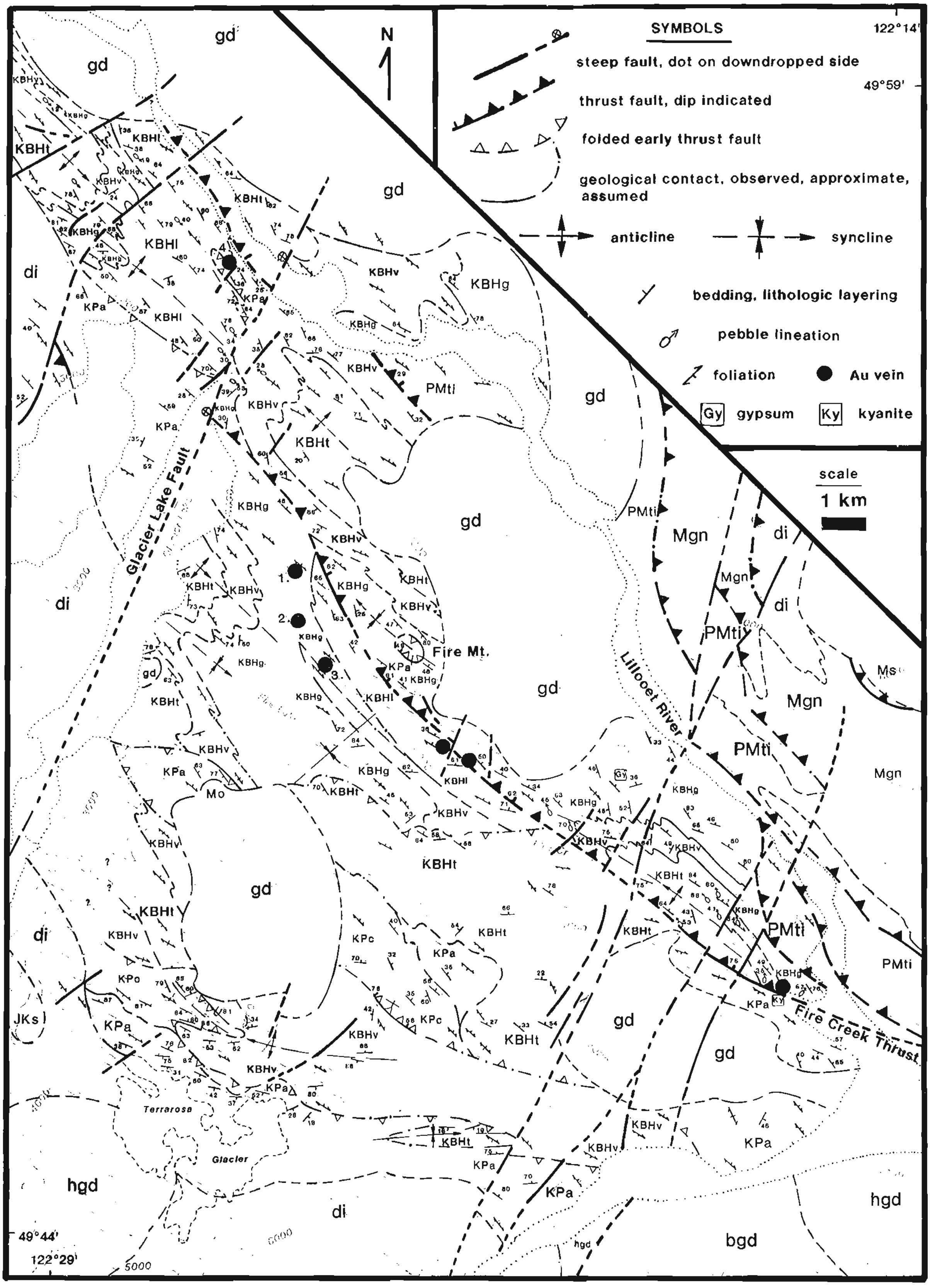
In the south-central portion of the map area (Fig. 2), the Peninsula Formation passes upwards uninterrupted into the Brokenback Hill Formation. The lowest member consists of interbedded feldspar crystal tuff, with slate or phyllite. The tuff contains 70-95 % moderately to well sorted, rounded, medium grained feldspar, in a pelitic matrix. It is generally massive and thin- to thick-bedded, but locally displays graded bedding, flame structures and local mud rip-up clasts. Because of the dominance of feldspar, and near absence of quartz or lithic fragments, the composition precludes an erosional provenance (Dickinson and Suczek, 1979). The rock is interpreted as a crystal tuff deposited and reworked under subaqueous conditions.

Above the feldspar crystal tuff are andesite and intermediate volcanic rocks, rhyolite is rare. These volcanics vary widely in occurrence and texture. Massive flows of andesite with plagioclase and amphibole phenocrysts occur locally, but more commonly the unit occurs as a heterolithic volcanic breccia or conglomerate. The fragmental rocks are poorly sorted, clasts are matrix supported in feldspar crystals, finer volcanic clasts, and abundant mud. Locally these appear as debris flows where 1-2 m sized volcanic clasts are suspended in a finer grained matrix. Monolithic autoclastic breccia of andesite was also found. Breccia clasts, or bombs, with chilled margins and reaction rims are seen in places. The unit is generally metamorphosed to greenschist facies, and this locally obscures primary textures. Secondary chlorite, muscovite, biotite, epidote, albite, calcite, and actinolite are widespread.

The third member is mostly coarse grained volcaniclastic sandstone; locally it is granular or pebbly. It is feldspar-rich but contains green, chloritized lithic volcanic fragments and has a green chloritic groundmass. Quartz is a minor component. The rock may be termed a feldspathic greywacke. Clasts are rounded, but primary sedimentary structures are rare and the member is poorly bedded.

Southeast of Fire Mountain a conspicuous gypsum-bearing unit occurs within the member (Fig. 2). Medium to fine grained gypsum (40-60 %) cement sand in a bed of crumbly and light coloured rock. The bed is 3-5 m thick and can be traced for about 100 m. Disseminated pyrite within it locally makes up 15 % of the rock. Breccia textures are also found, with 10-15 cm clasts of gypsum rock cemented by a second phase of gypsum. Because of its association with the subaqueous volcanic Brokenback Hill Formation, the unit likely represents a portion of an exhalative deposit.

The upper member of the Brokenback Hill Formation is dominated by a complex array of pyroclastic volcanic rocks. The most distinct and commonest rock is lapilli tuff with 1-3 cm angular fragments in a clast-supported framework of aphanitic felsic, intermediate, and varicoloured volcanics. Clasts are commonly welded and flattened in the bedding plane, forming a competent rock. The tuffs are well bedded. Sedimentary rocks occur only at the base where they grade into the underlying member. These volcanics may have been deposited largely under subaerial conditions. Diverse and minor rock types within the member include flow banded rhyolite, massive rhyolite with quartz and feldspar phenocrysts, pumice, andesite, and volcanic breccia.



SYMBOLS
122°14'
49°59'
steep fault, dot on downdropped side
thrust fault, dip indicated
folded early thrust fault
geological contact, observed, approximate, assumed
anticline
syncline
bedding, lithologic layering
pebble lineation
foliation
Au vein
Gy gypsum
Ky kyanite
scale
1 km
N
gd
di
KBHt
KBHl
KBHv
KBHg
KPa
KPc
PMtl
Mgn
Ms
Mo
JKs
hgd
bgd
Glacier Lake Fault
Fire Mt.
Lillooet River
Fire Creek Thrust
Terrarosa
Glacier
49°44'
122°29'
5000
6000

STRUCTURE

Three phases of deformation, including two distinct phases of folding and thrusting, and later block faulting are observed.

Early Shallow Angle Thrusting

The first deformation is characterized by shallow angle and bedding-parallel thrusts which superpose the Peninsula Formation onto the Brokenback Hill Formation. A single major fault is traced (Fig. 2), across which considerable displacement appears to have occurred. Folds are in the hanging wall of the fault. These are tight, and overturned towards the south-southeast (Fig. 3). The folds are moderate in scale and do not affect the map pattern at a 1:50 000 scale. Along the fault surface are many small en echelon shear bands (Fig. 4). These bound sigmoidal foliation with asymmetry consistent with south-southeast directed transport. Sets of conjugate Riedel shears are also abundant. Stretching and boudinage of sandstone beds is observed within the fault zone. Rotation of the boudins also indicates south-southeast transport.

The thrust fault is folded about northwest trending axes. Displacement indicators are folded with the thrust so that where the fault dips steeply it locally appears to have suffered strike-slip. Slip indicators have contrasting dextral and sinistral sense on opposite limbs of the folded thrust.

Steep Angle Thrusting

The second deformation is characterized by tight, large amplitude northwest-trending map scale folds (Fig. 2). Folds are noncylindrical and plunge northwest or southeast. They are generally angular and contain several orders of parasitic folds. A pervasive axial planar cleavage defined by parallel aligned mica penetrates the Fire Lake Group. Deformation apparently occurred during greenschist metamorphism. Clasts from sandstone and conglomerate are commonly flattened and stretched near shear zones.

A large northwest-striking high-angle thrust (the Fire Creek Thrust) juxtaposes lower against upper Brokenback Hill Formation members and is associated with the second deformation. This thrust cuts diagonally through the Fire Lake Group (Fig. 2) along Fire Creek and over the ridge crest of Fire Mountain. This fault is one of a family of Late Cretaceous imbricate thrusts from within the Northwest Cascade System and adjacent Cascade Metamorphic Core (Journeay and Csontos, 1989). They occur to the east of, and bound the Fire Lake Group along its southeastern margin. The Fire Creek Thrust is the most westerly known member of this family.

Fabrics within the Fire Creek Thrust are ductile, and rocks have undergone intense metamorphism. Steeply dipping foliation is bounded by northeast-dipping shear bands producing sigmoidal or "c" and "s" textures, indicating steep angle southwest-directed thrusting (Fig. 5). Thick quartz veins are stretched, boudinaged, detached, and back rotated along the shear bands (Fig. 6). Small-scale folds within the fault zone plunge northwest or southeast and have a noncylindrical geometry. Mineral lineations plunge moderately to steeply northwest to northeast along the folia-

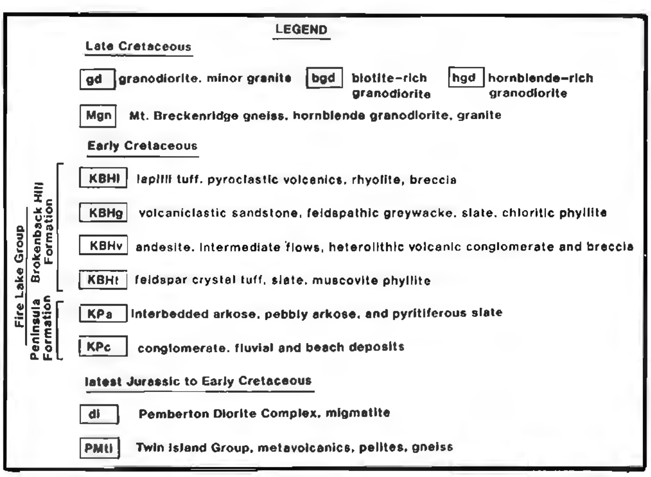


Figure 2. Geological map of the area underlain by the Fire Lake Group. The numbers 1, 2, 3, and 4 refer to the Blue Lead, Barkoola, Money Spinner, and Dandy Au prospects respectively.

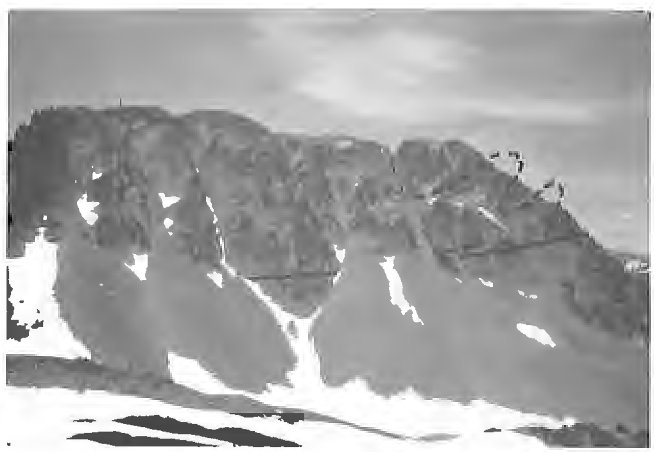


Figure 3. View of Fire Mountain looking east-northeast, showing overturned folds in the hanging wall of early thrust fault. Folds are overturned to the southeast, Peninsula Formation is in the hanging wall, Brokenback Hill Formation is in the footwall.

Figure 4. Shear bands from fault zone photographed in Figure 3, upper unit moved to the right (i.e. south-southeast) during shearing.

tion, and are bent into the shear bands. The lineations are consistent with steep-angle thrusting with a dextral slip component.

Petrographic observations are required for complete documentation of the metamorphic assemblages in the Fire Creek Thrust, however a talc schist with quartz-kyanite segregations is found within the southeast portion of the thrust (Fig. 2). The possible coexistence of talc, kyanite, and quartz in these "whiteschists" suggests that high pressure and temperature conditions prevailed during the thrusting event (Schreyer, 1977). Considerable burial of the Fire Lake Group during the second deformation may be implied. Although the zone of ductile deformation and intense metamorphism is typically only about 100 m wide, microscopic and hand sample scale shear bands which mimic the fault zone, are found in the thrust hanging wall as much as 3 km away from the fault.

Northeast-striking faults

Prominent steep dipping, northeast-striking faults characterize the latest deformation. These are regional in distribution, and extend well beyond the Fire Lake Group (Roddick, 1965; Monger, 1986). The faults are Tertiary; associated structures and fabrics indicate dextral transcurrent motion (Journeay and Csontos, 1989). They localize Tertiary dykes and plutons. Within the Fire Lake Group the Skookumchuck and Sloquet Creek hot springs are controlled by such structures. Surface traces are straight, and they are typically marked by physiographic depressions. The Glacier Lake Fault which runs along the valley of Glacier Lake is a good example (Fig. 2). The faults are rarely exposed. Slickensided surfaces and small-scale offsets in bedding and lithology in a few exposures indicate possibly dextral, normal, and dip-slip motion. Finite map-scale offset along the Glacier Lake Fault, based on displaced lithology, structure, and fold axes, is considered to have been accomplished by dip-slip with equal normal and dextral components, on the order of 4.5 km with downdrop in the northwestern block.

MINERALIZATION

Diverse styles of mineralization are found in the Fire Lake Group. These include syngenetic volcanic-exhalative mineralization, granodiorite-related stockworks and skarn, and high-angle thrust-related mesothermal Au-Cu veins, and late fault-related epithermal mineralization.

Figure 5. Shear bands from the Fire Creek Thrust, view looking northwest. Foliation is asymmetrically bent into the shear bands, indicating steep angle thrust style of faulting with thrust direction towards the southwest.

Figure 6. Exposure of the Fire Creek Thrust looking northwest, demonstrating boudinage and back rotation of quartz veins along shear bands (pack sack at bottom right-hand corner for scale).

The correlation of the Fire Lake Group to the Gambier Group (Roddick, 1965) signalled the potential for volcanogenic massive sulphide mineralization in the Fire Lake Group, as the Britannia Cu-Zn-Pb-Ag-Au orebody is a well known example from the Gambier Group (Payne et al., 1980). A marginal facies to typically zoned volcanogenic massive sulphide deposits, (Kuroko deposit) contains bedded and brecciated gypsum. The discovery of a significant gypsum bed with disseminated pyrite and syngenetic brecciation textures, in the volcanic Brokenback Hill Formation provides evidence for this style of mineralization, and constitutes a good prospect.

Molybdenite is found at the northern tip of the circular granitic body situated in the southwest part of the Fire Lake pendant (Fig. 2). The mineralization occurs in stockwork veinlets, hosted by a garnet-bearing granite. Gossanous zones, silicification, disseminated pyrite, and stockwork veining are present along the margins of some intrusive bodies. Small magnetite skarns were also found at a few localities.

The Fire Creek Thrust is a high-angle, deep seated thrust fault which appears to have influenced the distribution of Au-Cu veins (Fig. 2). In general, steep-angle thrust faults are recognized as an important structural style in controlling Archean lode Au veins, and younger Au vein systems such as in the Motherlode district of California (Sibson, 1989). Gold deposits commonly occur directly within, or adjacent to major steep angle thrusts. In the Fire Lake Group, three significant Au-bearing veins occur in the footwall of the Fire Creek Thrust directly northwest of Fire Mountain: the Money Spinner, Barkoola, and Blue Lead veins (Fig. 2). The veins have received considerable attention since their discovery in the 1890s, including some small scale mining and underground development (Roddick, 1965). Quartz dominates the veins; minor chalcopyrite, bornite, and sporadic native gold are present. Other gangue minerals include muscovite, chlorite, calcite, epidote, and tourmaline. Veins have variable orientations, and structural relationships to the thrust are complicated. The veins formed during active deformation; veins of the Money Spinner claim, form en echelon sigmoidal shears containing composite syntaxial and antitaxial quartz fibre growth. The enveloping surface or shear plane is irregular in outline at the outcrop scale, but appears to be flat lying or moderately dipping with shear sense indicating upper block movement towards the west. Within the Fire Creek Thrust, quartz-pyrite veins dip steeply and parallel the foliation, or occur as shallow dipping tension fractures known as "flats" (Sibson, 1989).

The Dandy prospect north of the Glacier Lake Fault in the footwall of the Fire Creek Thrust (Fig. 2) is distinguished by the presence of sphalerite, galena, as well as gold. The thrust is not exposed here; it follows the bed of the Lillooet River, but a linear zone of coarse breccia runs along the fault in the brittle footwall, (Fig. 8 of Journeay and Csontos, 1989).

The late northeast-striking faults of the region control the emplacement of Tertiary high level felsic plutons and dykes, to which granite-related and epithermal Au mineralization are associated (Journeay and Csontos, 1989). In a few northeast striking faults from the Fire Lake Group, coarse pyrolusite crystals are found in quartz-calcite-pyrite veins.

DISCUSSION

The Fire Lake Group is a sedimentary-volcanic sequence of Early Cretaceous rocks that is thought to have formed in an island arc. The Peninsula Formation is a fining-upwards sequence deposited during volcanic quiescence. Facies progress upwards from fluvial, to beach, to possibly marine shelf. The overlying Brokenback Hill Formation is a complex volcanic sequence dominated by subaqueous autoclastic, and epiclastic rocks of mostly intermediate composition, with welded pyroclastic rocks of likely subaerial origin at the top of the formation.

Three phases of deformation affect the Fire Lake Group. The earliest is characterized by shallow-angle thrusting emplacing the Peninsula Formation onto Brokenback Hill Formation. Kinematic indicators and fold vergence indicate south to southeast thrusting. The deformation records a period of orogen-parallel shortening. The age of this event is bracketed by the Early Cretaceous age of the Fire Lake Group and Late Cretaceous second-phase deformation. It is speculated that thrusting occurred in conjunction with strike-slip faulting along the continental margin, as a step-over thrust or fault tip splay: in the manner of transpressional terranes (Ave Lallemant and Oldow, 1988; Brown, 1987).

The second deformation had a strong effect on the fabric of the rocks. It is characterized by widespread, large-amplitude noncylindrical folds, and northwest-striking steeply dipping cleavage. The Fire Creek Thrust is a part of a family of Late Cretaceous steep angle thrust faults (Journeay and Csontos, 1989) which transported high grade, deep seated rocks of the Northwest Cascade System and Cascade Metamorphic Core onto the supracrustal rocks of the Fire Lake Group. Collectively the faults accommodate considerable arc-normal shortening and uplift, likely forming a hinterland to Late Cretaceous southwest directed thrusting in the southern Coast and Insular belts.

Northeast-striking dextral-normal dip-slip faults of Tertiary age characterize the latest phase of deformation. These may have formed as the result of tension localized at the southwest tip of the Fraser dextral strike-slip fault system, possibly in conjunction with sedimentation in the Georgia Basin.

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Geology of the Cairn Needle area east of Harrison Lake, southwestern British Columbia

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Tyson, T., Geology of the Cairn Needle area east of Harrison Lake, southwestern British Columbia; in Current Research, Part E, Geological Survey of Canada, Paper 90-1E, p.205-211, 1990.

Abstract

The study area records a complex sequence of folding, faulting and igneous intrusion as a response to Cretaceous crustal shortening. A continuous contractional event caused internal deformation within the Northwest Cascades System and the Cascade Metamorphic Core, and the juxtaposition of these two elements. Early isoclinal folds and axial-planar foliation were deformed by the Mt. Breakenridge Anticline, a second generation, northwest-plunging fold. The eastern limb of the anticline is truncated by a steep fault of unknown displacement which in turn is cut by the syn-deformational Spuzzum Granodiorite. Ductile deformation ended prior to intrusion of the unfoliated Scuzzy Pluton. The continuous nature of deformation is recorded by granitic dykes which were intruded throughout the deformational and intrusive history of the area. North-trending, right-lateral, strike-slip faulting postdates intrusion of the Scuzzy Pluton and may be part of a regional Tertiary brittle fault system.

Résumé

On trouve dans la région étudiée une série complex de plis, de failles et d'intrusions ignées qui sont la réponse à un raccourcissement de la croûte au Crétacé. Une contraction continue a été à l'origine d'une déformation interne au sein du système des Cascades nord-ouest et du noyau métamorphique des Cascades, ainsi que de la juxtaposition de ces deux éléments. Des plis isoclinaux et une foliation planaire axiale préliminaire ont été déformés par l'anticlinal du mont Breakenridge, qui est un plissement de seconde génération à plongement nord-ouest. Le flanc est de l'anticlinal est tronqué par une faille abrupte à déplacement inconnu, qui est traversée à son tour par la granodiorite de Spuzzum à déformation contemporaine. Une déformation ductile s'est terminée avant l'intrusion du pluton de Scuzzy non folié. La continuité de la déformation est marquée par des dykes granitiques qui ont pénétré au cours de toute la période de déformation et des intrusions de la région. Des failles à rejet horizontal, dextre de direction nord, se sont produits après l'intrusion de pluton de Scuzzy et peuvent faire partie du système régional du Tertiaire à failles fragiles.

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INTRODUCTION

This study aims to determine the geometry and timing of deformation on the boundary between the northern extension of the Cascade Metamorphic Core (CMC; Monger, 1986) and the Northwest Cascade System (NWCS; Brown, 1987).

It has been proposed that shortening of the crust during Cretaceous time was accommodated by multiple generations of folding and thrust faulting (Journeay and Csontos, 1989). Specifically, an early phase of low-angle, imbricate thrusting is believed to have been responsible for assembly of the NWCS during Early Cretaceous time. This early set of faults coincides with an early phase of folding of compositional layering. The thrusts are folded by a second generation of map-scale folds. This second generation of folds and the early set of imbricate faults are in turn cut by a set of high-angle reverse faults believed to be responsible for the juxtaposition of high-pressure Settler Schist and Spuzzum Granodiorite of the CMC with lower grade Twin Islands Group and Breakenridge Gneiss of the NWCS (Journeay and Csontos, 1989).

In the Cairn Needle area, east of Harrison Lake (Fig. 1) early low-angle faults are apparently truncated by a late set of high-angle faults. This relationship provides a good opportunity to examine the geometry and relative timing of faulting in relation to folding, metamorphism and intrusion.

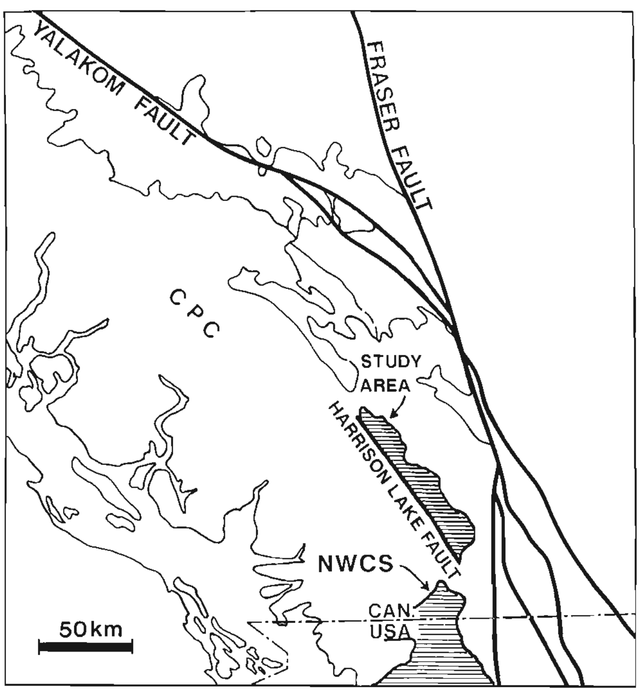

Figure 1. Map of southwest British Columbia showing the location of the study area and the locations of some of the major faults and tectonic elements, including the Coast Plutonic Complex (CPC) and the Northwest Cascade System (NWCS).

STRATIGRAPHIC UNITS

Twin Islands Group

Metamorphic rocks of the Twin Islands Group (Roddick, 1965) are exposed in the southwest of the map area (Fig. 2). The unit is bounded on the west by an intrusive contact with the Spuzzum Granodiorite, on the southeast by a concordant contact with the Breakenridge Gneiss and on the northeast by a fault contact with the Hunger Creek group. The Twin Islands Group is divided into two units. The structurally lower unit is dominated by banded amphibolites, commonly garnet bearing. The higher unit is predominantly semi-pelitic quartz+biotite$\pm$garnet schist. Also present in the upper unit are talc-sericite schists, meta-quartz-sandstones, hornblende-rich schists and amphibolite. The contact separating the lower and upper units is gradational and is interpreted to be depositional.

Hunger Creek group

Metamorphic rocks of the Hunger Creek group (informal name) are bounded on the northwest by an intrusive contact with the Spuzzum Granodiorite and on the southwest by a fault contact with the Twin Islands Group. In the eastern half of the area the Hunger Creek group is intruded by a generally concordant, but locally crosscutting, sheet of Spuzzum Granodiorite (Fig. 2). Large blocks (up to 20 m^3) of the Hunger Creek group are common as inclusions within the sheet of granodionite. The group is bounded on the east by a discordant, intrusive contact with the Scuzzy Pluton. The dominant lithology is a semi-pelitic quartz-biotite-garnet-hornblende-sillimanite schist. Minor marble, calc-silicate, pebble conglomerate, quartzite and amphibolite are present. One marble bed, 5-10 m thick, and a pebble conglomerate bed, 1-4 m thick, form distinctive markers.

The Hunger Creek group was mapped as Settler Schist by Lowes (1972) and Monger (1986). However, the rocks observed in the field area do not match the description of Settler Schist, which is described as being dominated by alumina-rich pelites. The unit that best correlates with the Hunger Creek group is the Twin Islands Group. This correlation, however, is uncertain; thus the informal term Hunger Creek group.

Deformational style

The deformation within the Twin Islands and the Hunger Creek groups is penetrative and characterized by a schistosity defined by the preferred orientation of elongate minerals and flattened quartz grains. Stretching and mineral lineations, discussed below, are common throughout both units.

Metamorphism

Metamorphic grade is middle amphibolite facies for both the Hunger Creek group and Twin Islands Group. Common index minerals are garnet, biotite, sillimanite and hornblende. The timing of metamorphic mineral growth predates or is synchronous with foliation development. Locally, growth of hornblende and garnet porphyroblasts postdates fabric development.

PLUTONIC ROCKS

Breakenridge Gneiss

The Breakenridge Gneiss is a polydeformed, recrystallized, leucocratic gneiss of igneous origin (Reamsbottom, 1971; Monger, 1986). The unit appears to predate the onset of deformation in the area. A U-Pb date about 20 km to the south indicates a crystallization age of 96 Ma (R.R. Parrish, pers. comm., 1989).

Spuzzum Granodiorite

The Spuzzum Granodiorite is a foliated to massive hornblende-biotite granodiorite. Foliation is defined by the preferred orientation of elongate mafic minerals. Equant quartz and plagioclase grains are not flattened, suggesting that the fabric is magmatic. The timing of intrusion of the pluton and development of the fabric is interpreted to be syndeformational. This interpretation is supported by the widespread presence of the magmatic foliation, which is concordant with the regional foliation, and by the generally concordant, but locally crosscutting, contacts with the metasedimentary rocks. The crystallization age for the Spuzzum is about 110-104 Ma (Gabites, 1985; R.L. Armstrong, pers. comm., 1989), although a recent, preliminary U-Pb analysis suggests a 95 Ma age of crystallization (Hettinga, 1989).

Recent fieldwork indicates that the Spuzzum Granodiorite extends at least 6 km to the west of the western edge of the map area.

Scuzzy Pluton

The Scuzzy Pluton is a massive hornblende-biotite granite that crosscuts the foliation in the Hunger Creek group and has a crystallization age of 86 Ma (R.R. Parrish, pers. comm., 1989).

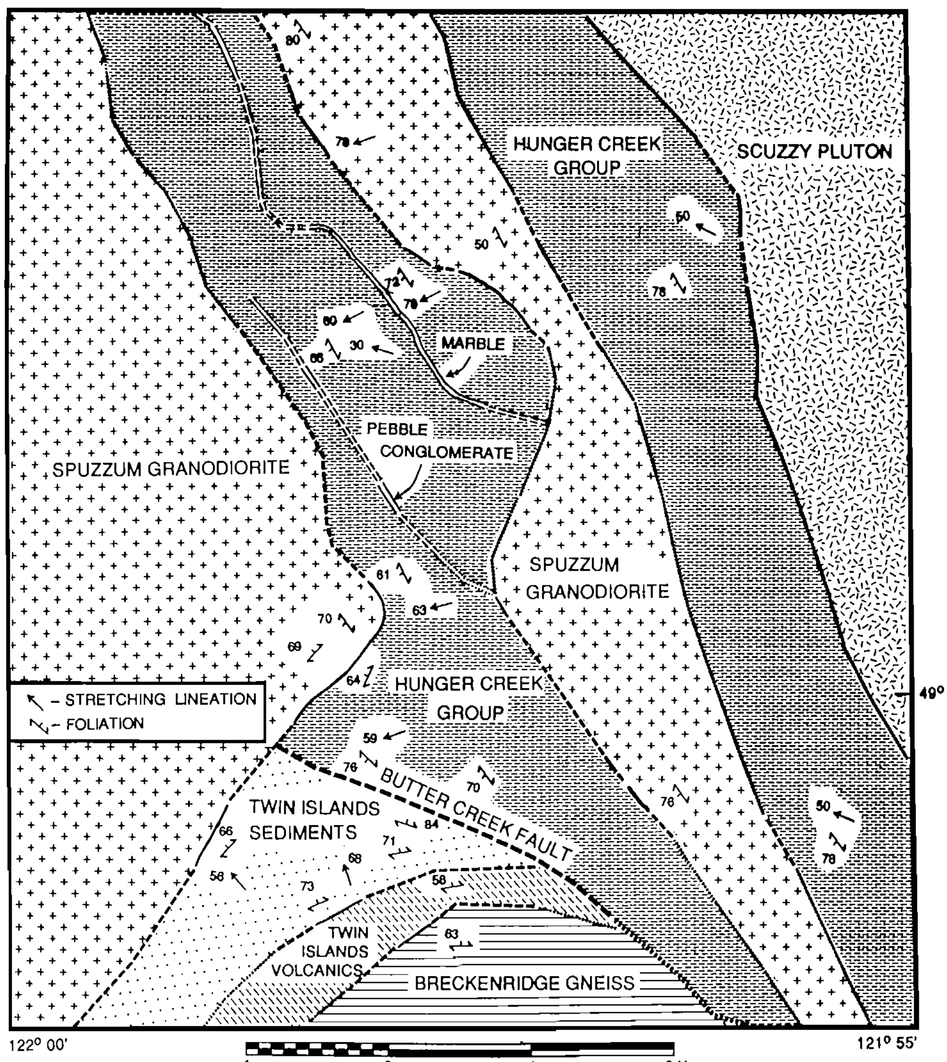

Figure 2. Geological map of the eastern half of the study area. The western half consists entirely of Spuzzum Granodiorite which extends at least to 122°05'W.

Granitic dykes

Granitic dykes are present throughout the map area as intrusive sheets, subparallel to the foliation. They are particularly abundant in the Hunger Creek group, constituting 20-70 % of the outcrop volume. The dykes form spectacular folds and pull-aparts and typically highlight small-scale structures (Fig. 3-5). The timing of intrusion is generally synchronous with deformation, but the dykes also occur as pre- and post-deformation intrusions.

Figure 3. Ptygmatic folding of a granitic dyke in semipelites of the Hunger Creek group. The view is to the southwest and the ice axe head is 20 cm long.

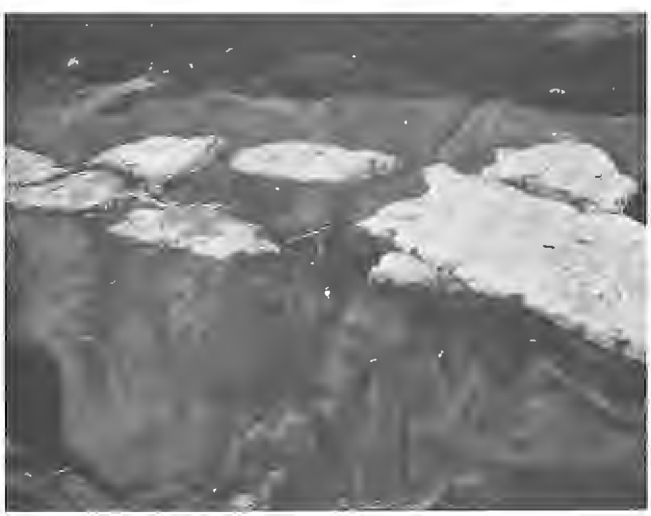

Figure 4. Pull-aparts of a syndeformational granitic dyke within the Spuzzum Granodiorite. View is to the southwest.

MAP-SCALE STRUCTURAL ELEMENTS

Breakenridge Anticline

The foliation in the Twin Islands Group wraps around the eastern limb of the Breakenridge Anticline. The anticline is a map-scale, F2 fold that plunges northwest and was mapped in detail to the south by Reamsbottom (1971). The timing of this fold is bracketed by the early development of the foliation and stretching lineations and later motion on the Butter Creek fault.

Butter Creek fault

The Butter Creek fault juxtaposes Hunger Creek and Twin Islands rocks. The fault trace is unexposed and, therefore, the nature of the fault is poorly understood. It is apparently quite steep, as foliations on both sides of the fault approach vertical close to the fault. The sense of motion and amount of displacement are unconstrained. Previous work suggested that the fault is one of a set of southwest-vergent reverse faults (Journeay and Csontos, 1989). There is no apparent metamorphic discontinuity across the fault, suggesting that the displacement may not be very large or that metamorphism postdated motion on the fault.

The relative timing of motion along the fault is well constrained. The Butter Creek fault truncates Twin Islands Group stratigraphy, requiring that displacement postdated the development of the Breakenridge Anticline (Fig. 2).

Figure 5. Granitic dykes in the Hunger Creek group. The dykes have been extended and then shortened by folding and rotation of the pull-aparts. View is to the northwest.

Displacement predated intrusion of the Spuzzum Granodiorite since the northward trace of the fault is truncated by the Spuzzum.

Brittle faults

There is evidence for discrete north-south to northeast-southwest trending brittle deformation throughout the area. The deformation is generally expressed as narrow (1-2 m) zones of steeply-dipping spaced cleavage. There is one brittle fault within the Spuzzum Granodiorite, about 4 km west of the western edge of the map area (Fig. 6). Dextral, strike-slip motion on the fault is suggested by the offset of a volcanic dyke displaced by less than 50 m. Brittle deformation postdates intrusion of the Scuzzy and is believed to be part of a regional Tertiary brittle fault system (Journeay and Csontos, 1989).

SMALL-SCALE STRUCTURES

Two generations of folds are present. The first generation is rarely observed and is characterized by tight to isoclinal, steeply-plunging folds of compositional layering with an axial-planar foliation (Fig. 7, 8). The development of the F1 folds and the regional foliation in the metasedimentary rocks was probably synchronous.

The second generation of folds is characterized by tight folds of foliation with axial hinges plunging moderately northwest (Fig. 8). The geometry and timing of the F2 folds strongly suggests that they are part of the same deformational event which produced the Breakenridge Anticline and other map-scale F2 folds observed to the southwest by Reamsbottom (1971).

Stretching and mineral lineations are common in both metasedimentary rock packages. Stretched quartz grains commonly have aspect ratios greater than 10:1 (Fig. 10). The lineations are primarily down-dip with a minor oblique component (Fig. 7) and are widely distributed, not localized. No kinematic indicators were found in the metasedimentary rocks. Previous work (Journeay and Csontos, 1989) suggests that these lineations are the result of motion on low-angle thrust faults.

Figure 6. Fault breccia within the Spuzzum Granodiorite with fragments of the Spuzzum and a volcanic dyke.

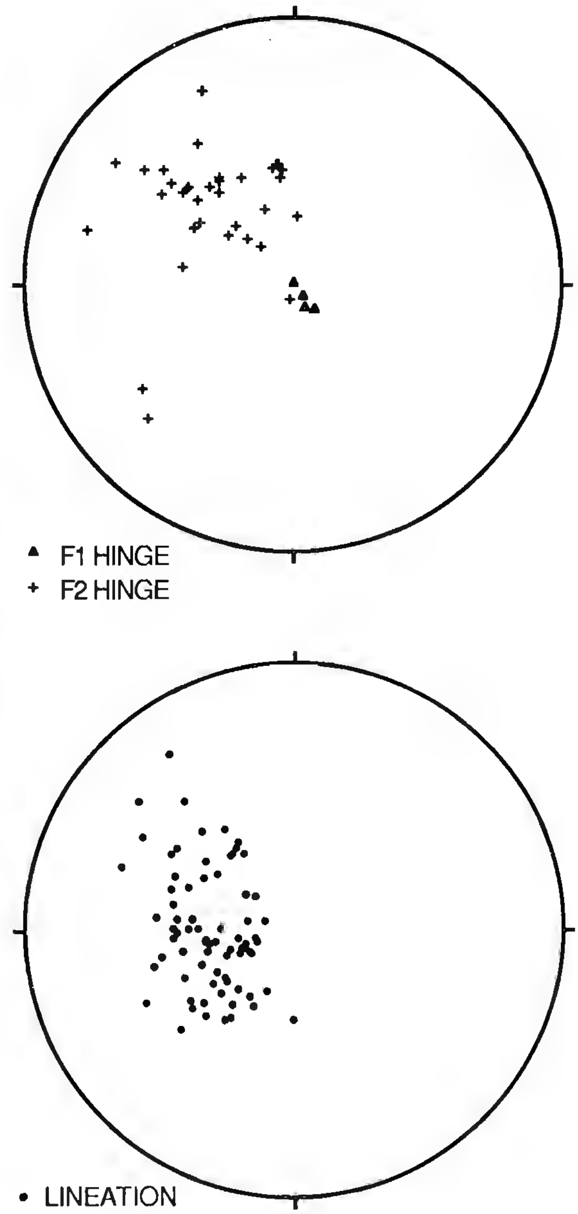

Figure 7. Equal-area stereonets with plots of fold hinges and lineations, both stretching and mineral.

DISCUSSION AND CONCLUSIONS

The relative timing of deformational events in the field area is incompletely understood, but certain critical observations allow a partial chronological reconstruction. The earliest event is the development of isoclinal F1 folds with an associated axial-planar foliation. The development of the foliation in the metasedimentary rocks may have been contemporaneous with F1 folding. Layer-parallel shearing, as indicated by stretching lineations, may have been associated with the development of the foliation. This timing relationship, however, is poorly constrained, and lineation formation may have postdated F1 folding and fabric formation. F2 folding resulted in the formation of the Breakenridge Anticline and associated parasitic folds. F2 folding deforms the foliation and also appears to fold the lineations. The Breakenridge Anticline is truncated by the high-angle Butter Creek fault which juxtaposes Hunger Creek group against Twin Islands Group. Intrusion of the syndeformational Spuzzum Granodiorite postdates motion on the Butter Creek fault. Folding and extension of late granitic dykes within the Spuzzum indicate that deformation outlasted intrusion, though not necessarily crystallization, of the granodiorite. Ductile deformation had ended prior to intrusion of the unfoliated Scuzzy Pluton. The final event recorded in the field area is north- to northeast- trending brittle deformation.

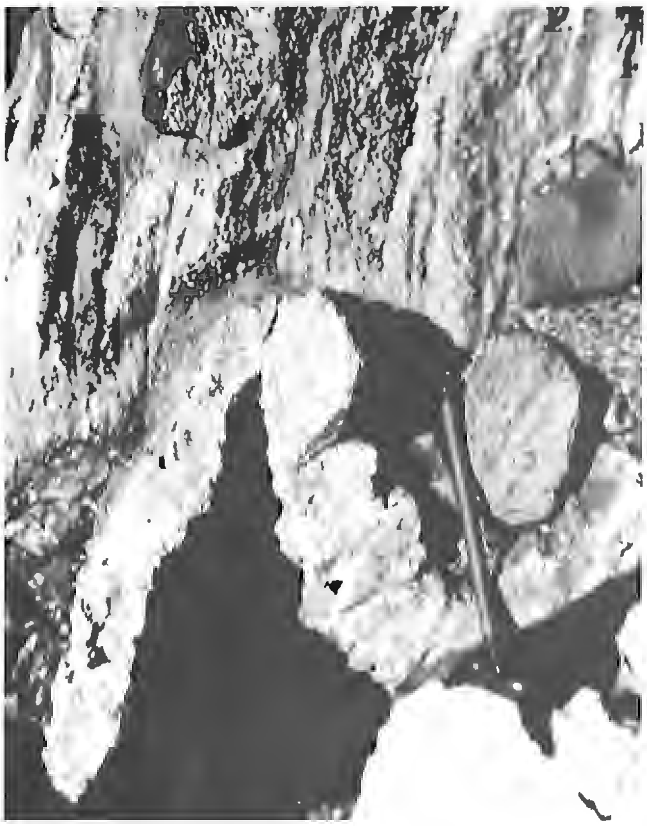

Figure 8. F1 fold within the marble marker in the Hunger Creek group. Photo taken looking down the hinge-line, which plunges steeply to the southeast.

Figure 9. F2 folds of foliation. Fold is highlighted by folded granitic dykes which are subparallel to the foliation. View is to the northwest and the distance across the photo is approximately 60 m.

Most metamorphic mineral growth predated or was synchronous with F1 folding and the development of the foliation. Late porphyroblasts of garnet and hornblende are syn- to post-fabric development. The grade of metamorphism is middle amphibolite facies as indicated by the presence of biotite, sillimanite, hornblende and garnet.

It is critical to note that the style of deformation is not one of punctuated episodes of folding, faulting and intrusion. Instead, the deformation is interpreted as a single, continuous event from the time of F1 folding and fabric development through the intrusion of the Spuzzum Granodiorite. This interpretation is based primarily on the continu-

Figure 10. Stretching lineations in semipelites of the Hunger Creek group. View is to the northeast.

ous deformation of granitic dykes that were intruded throughout the deformational and intrusive history of the area.

The absolute timing of deformation and metamorphism is poorly constrained. Previous work suggested that ductile deformation and metamorphism are bracketed by the involvement of the 96 Ma Breakenridge Gneiss in the deformation and by the undeformed 85 Ma Scuzzy Pluton. The problem with this scenario is that the Spuzzum Granodiorite, 110-104 Ma, postdates development of F2 folds. The Spuzzum is recognized as a composite pluton, however, and there are multiple slices of the Breakenridge Gneiss in the NWCS. Therefore, it is reasonable to suggest that either the Spuzzum in the study area is a young part of the composite pluton or that the Breakenridge Gneiss contains rocks of different ages of crystallization. U-Pb samples have been collected from the Spuzzum pluton, Breakenridge Gneiss, and syndeformational granitic dykes, to attempt to resolve the timing problem.

Preliminary results from this study agree well with the two generations of shortening proposed by Journeay and Csontos (1989). Differences, however, have been observed. First, during the early generation of deformation shortening was accommodated by folding and by motion along foliation planes distributed throughout the metasedimentary sequence. In contrast, Journeay and Csontos (1989) identified discrete thrust faults. The implications of this observation are unclear. The lack of first generation thrust faults may simply be due to the style of deformation in the area or there may be more serious problems with the proposed sequence of events. The second difference between their model and relationships observed in this study is with the direction of motion in the early stage of shortening. They postulated southwest-vergent reverse displacements. Figure 2 shows that in the Hunger Creek group, the foliation and the lineations dip to the southwest. If the lineations are a product of shortening, then this suggests northeast-directed motion. The bulk of evidence from the region consistently indicates southwest-directed motion. The simplest way to resolve this apparent problem is to assume that the Hunger Creek group has been rotated by motion along the Butter Creek Fault.

ACKNOWLEDGMENTS

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Mapping in the northeastern part of the Pemberton Dioritic Complex, Pemberton map area, British Columbia.

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Friedman, R.M. Mapping in the northeastern part of the Pemberton Dioritic Complex, Pemberton map area, British Columbia; in Current Research, Part E, Geological Survey of Canada, Paper 90-1E, p.213-217, 1990.

Abstract

Three northwest- to north-northwest- striking intrusions and a volcaniclastic pendant out crop in the study area (92J). The southeast tonalite body intrudes regional subgreenschist grade volcaniclastic rocks. Several steeply northeast dipping shear zones, 5-300 m wide, contain mineral lineations that plunge about 25° southeast on mylonitic foliation surfaces that roughly parallel shear zone margins. Kinematic indicators indicate dextral slip with a component of normal displacement.

The tonalite intrusion is deformed in a shear zone which is cut by late phase arms of the same body. Heat from the cooling body may have facilitated ductile deformation in adjacent low metamorphic grade volcaniclastic rocks during shear zone activity. The age of motion, unknown at present, is important because the dextral Harrison Lake Shear Zone lies along strike to the southeast.

Résumé

Trois intrusions de direction nord-ouest à nord-nord-ouest et une apophyse volcanoclastique affleurent dans la région à l'étude (92 J). Le massif de tonalité sud-est pénètre des roches volcanoclastiques du faciès des sous-schistes verts. Plusieurs zones de cisaillement à fort pendage nord-est, de 5 m à 300 m de large, renferment des linéations minérales qui plongent à environ 25° vers le sud-est sur des surfaces de foliation mylonitiques qui sont plus ou moins parallèles aux marges de la zone de cisaillement. Des indicateurs cinématiques indique un rejet dextre avec une composante de déplacement normal.

L'intrusion tonalitique est déformée dans une zone de cisaillement qui est traversée par des apophyses de la dernière phase du même massif. La chaleur fournie par le refroidissement du massif peut avoir facilité une déformation ductile dans des roches volcanoclastiques adjacentes à faible métamorphisme au cours de l'activité de la zone de cisaillement. L'âge du mouvement, inconnu jusqu'à présent, est important car la zone de cisaillement dextre de Harrison Lake repose dans la direction générale vers le sud-est.

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INTRODUCTION

The Pemberton Dioritic Complex (PDC), in the central part of the southern Coast Belt (Fig. 1), is a heterogeneous assemblage of rocks composed of hornblende-rich diorite, quartz diorite and amphibolite with lesser granodiorite and stratified rocks (Roddick, 1965; Woodsworth, 1977; Price et al., 1985). It is bounded by a presumed Early Cretaceous pendant on the east and plutonic rocks of the Coast Belt on the west (Fig. 1). Journeay and Csontos (1989) recognized several mappable intrusions in the northeastern area and Monger (1990) has begun work in the southern part of the complex.

This project, which constitutes part of a regional U-Pb and K-Ar geochronometry study of the southern Coast Belt (Friedman and Armstrong, 1989; Friedman, 1989), focuses on the geology and geochronometry of part of the northern PDC. The study area is located in the Whistler (92J/2) 1:50 000 map area (Fig. 2). Mapped rock units, field relationships, and structures are described in this report. The results of geochronometric studies will be reported in a subsequent report.

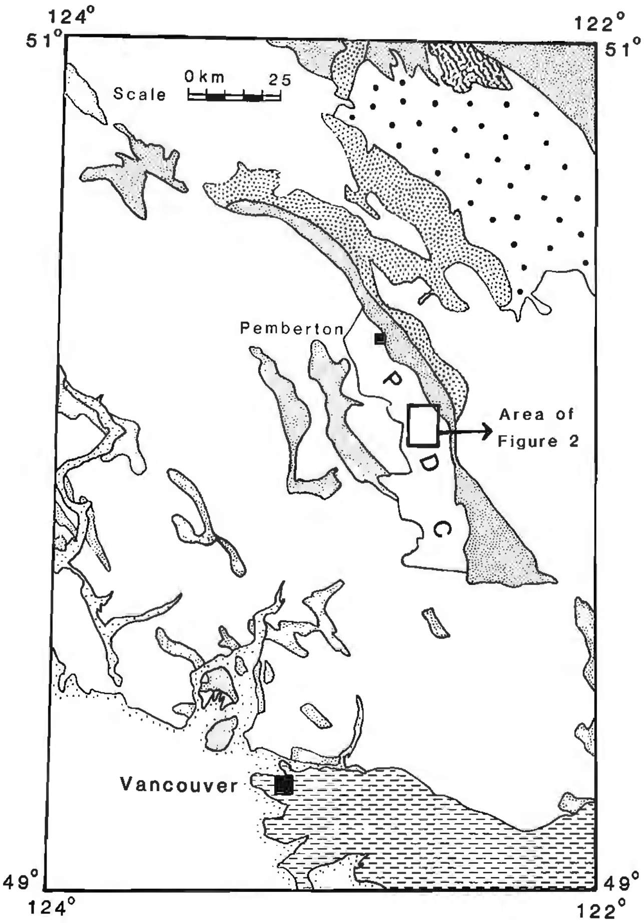

Figure 1. Map of southwestern British Columbia showing the location of the study area and major rock units. PDC: Pemberton Dioritic Complex; other unpatterned areas: granitic rocks; fine stipple: Jurassic and Cretaceous strata, mostly Gambier and Bowen Island groups in the west and Fire Lake groups east of the PDC, Harrison Lake section in the southeast, and rocks of the Tyaughton Basin in the northeast; medium stipple: Cadwallader Group; coarse stipple: Bridge River Group; mottled pattern: Shulaps ultramafite; dashed pattern: Latest Cretaceous to Recent sedimentary rocks and sediments. Sources are Roddick (1965), Roddick et al. (1979), and J.M. Journeay (pers. comm., 1989).

ROCK UNITS AND CONTACT RELATIONSHIPS

Plutonic rocks

Three elongate northwest to north-northwest striking intrusions are shown on Figure 2. The southwestern two were mapped as part of this study and are described below. The position of the northeastern quartz dioritic intrusion on Figure 2 is based on data from a compilation map of the region (J.M. Journeay, pers. comm., 1989).

The central dioritic intrusion is composed of medium grained hornblende diorite and quartz diorite. It is intruded by dykes of intermediate composition which are up to a metre wide. Small clots and stringers of epidote are also common. These dioritic rocks are largely undeformed or weakly foliated, however centimetre-scale ductile shear zones have been observed, and are spatially associated with dykletes (Fig. 3). The southern part of the dioritic intrusion, south of volcaniclastic strata, is based on unpublished mapping by J.M. Journeay.

The southwestern intrusion is made up of medium grained biotite-hornblende tonalite, a relatively felsic rock type in the PDC. Quartz commonly forms 25-35 per cent of this rock, with K-feldspar making up 0-10 per cent. This intrusion is dominantly nonfoliated or weakly foliated but locally is deformed by metre-scale, southeasterly striking, steeply dipping shear zones where mylonitic rocks have been developed. The largest of these shear zones, which is at least 300 m wide, overlaps the contact with volcaniclastic rocks of a small pendant (Fig. 2). Late phase arms of garnet aplite and, less commonly, pegmatite range from a few centimetres to several metres in width and intrude volcaniclastic rocks along the tonalite-pendant contact.

As with the dioritic rocks described above, this tonalite body has been intruded by dykes of intermediate composition. Based on mutual crosscutting relationships between dykes and tonalitic country rock seen at several localities, these dykes are thought to be at least in part coeval with late-stage crystallization of the tonalite body (Fig. 4). At one locality a train of intermediate-composition xenoliths in weakly foliated tonalite is interpreted as a disrupted dyke.

The diorite-tonalite contact varies along strike. Directly south of Carter Glacier a steeply dipping ductile shear zone 5-15 m wide marks the transition from tonalite to diorite. The shear zone is almost completely contained in the tonalitic body. Farther southeast the diorite-tonalite contact is a zone about 100 m wide consisting of a fine grained phase of quartz diorite (shown as a dashed, approximate contact on Fig. 2). The southeastern continuation of this contact is not exposed or is inaccessible.

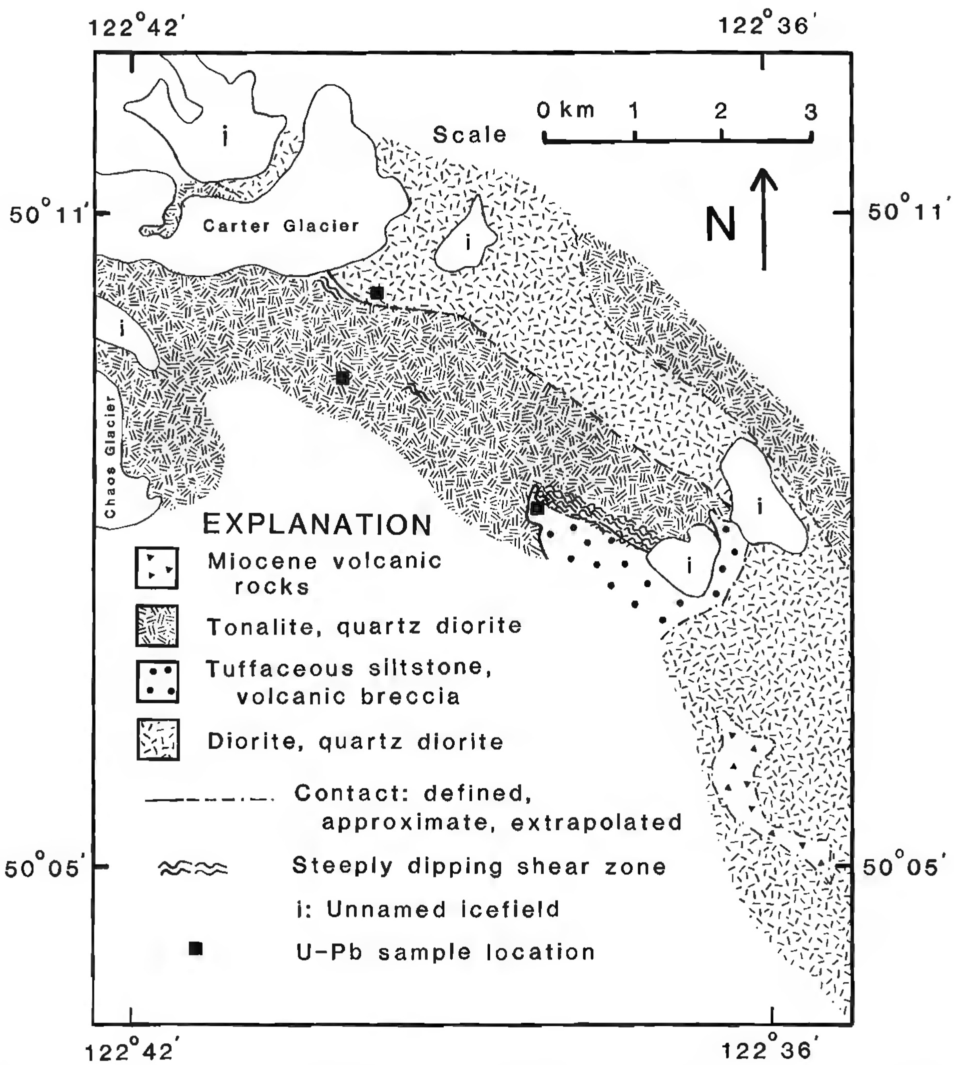

Figure 2. Geological map of study area in the northern part of the PDC. See Figure 1 for regional setting.

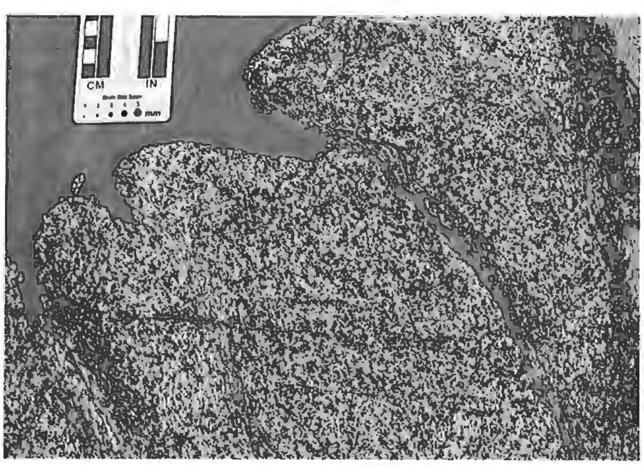

Figure 3. Hornblende diorite with intermediate composition dyke in upper left hand corner. Note minor shear zones associated with three dykletes.

The contact described above yields no information about the relative ages of the diorite and tonalite bodies. Several indirect lines of evidence indicate that tonalite intrudes diorite. Rare centimetre-scale xenoliths of diorite have been observed near the contact. In addition, dykes interpreted to be coeval with late crystallization have been seen only in the tonalite. This argument hinges on an unproven age correlation between lithologically similar dykes in both intrusions. Finally, an outcrop scale body of tonalitic rock that intrudes diorite on a pinnacle near the contact is interpreted as an arm of the main tonalite body. U-Pb zircon dating of these intrusions will support or refute these interpretations.

Volcaniclastic rocks

A pendant at least 500 m thick, made up of mostly volcaniclastic rocks, is exposed in the eastern part of the study area (Fig. 2). The rock types seen are tuffaceous siltstone, volcanic breccia (Fig. 5), rusty-weathering argillite and

Figure 4. Mutually crosscutting relationship between dyke and tonalitic country rock.

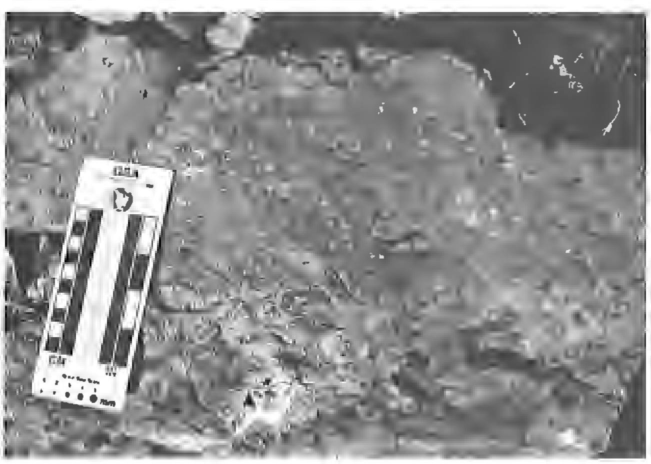

Figure 5. Volcanic breccia in pendant. Note grading of clasts.

lesser calcareous siltstone and marble. Bedding is preserved away from the tonalite contact where rocks do not exceed subgreenschist regional metamorphic grade. Near the contact rocks have been hornfelsed and locally mylonitized (Fig. 6). No fossils have been found and consequently these

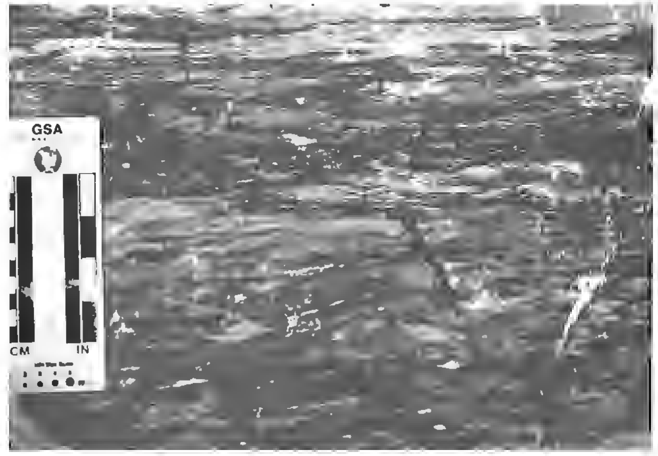

Figure 6. Mylonitic metasedimentary rocks derived from volcaniclastic siltstone.

strata can only be tentatively correlated with Lower Cretaceous rocks of the Fire Lake Group to the east (Journeay and Csontos, 1989; Roddick, 1965).

The pendant-tonalite contact is unequivocally intrusive in nature. Late-phase arms and main-phase tonalitic rocks intrude and engulf volcaniclastic rocks. A segment of this contact is deformed in a ductile shear zone (Fig. 2). Mylonitic fabrics are well developed in both tonalitic and volcaniclastic rocks (Fig. 6). Some late-phase apophyses cut deformation fabrics, indicating that the timing of intrusion and shear zone activity are broadly synchronous.

The southern, pendant-diorite contact (Fig. 2) has been interpolated between volcaniclastic rocks visited in this study (or observed in cliff faces) and rocks described as diorites in previous GSC fieldwork. The nature of this contact is unknown because it was not directly visited. It should be noted that about 700 m away from where dioritic rocks out crop, there is no evidence of contact aureole development.

STRUCTURE

The overall structural grain of intrusions in the study area strikes northwesterly to north-northwesterly and dips steeply. Plutonic rocks are undeformed to weakly foliated except within a few discrete northwesterly striking, steeply northeasterly dipping shear zones where mylonitic fabrics have been developed. As discussed below, fabrics in shear zones provide evidence for dominantly dextral strike-slip offset.

The structural roof of the tonalitic intrusion is exposed at the pendant contact. Subgreenschist regional metamorphic grade volcaniclastic rocks away from the contact aureole indicate that this part of the tonalite was emplaced at high crustal levels.

Shear zones

Several ductile shear zones have been encountered in the mapped area. Three of the more significant ones are plotted

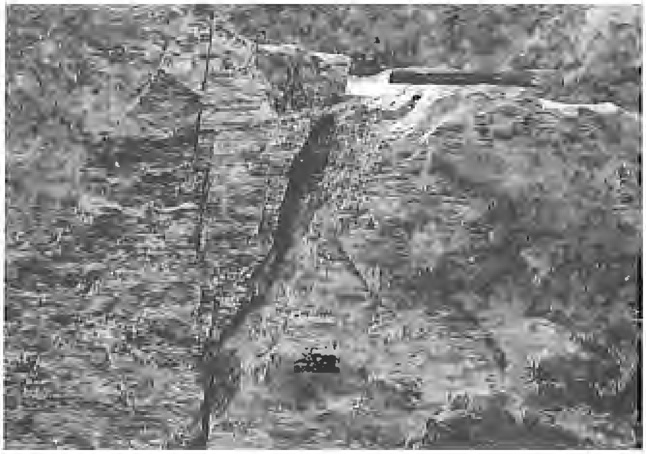

Figure 7. Shallowly plunging mineral elongation lineations on mylonitic foliation surface, central shear zone on Figure 2.

in Figure 2. The boundaries of these zones and the internal mylonitic foliation strike at 125-145° and dip towards the northeast at 70-85°. Mineral elongation lineations (inferred to indicate stretching directions), defined by quartz, plagioclase, and hornblende plunge 10-40° towards 115-140° (Fig. 7). Plunges of about 25° are most common.

Preliminary analysis of kinematic indicators in the field and on slabs cut parallel to lineation and perpendicular to foliation, implies a dextral shear couple parallel to the mineral elongation lineation. Assuming shear zone motion was parallel to mineral elongation lineations, slip was dextral (hanging wall towards the southwest) transcurrent with a component of normal motion. The useful indicators in these rocks are shear bands and asymmetric pressure shadows around porphyroclasts.

The relative timing of motion along the pendant-tonalite shear zone is well constrained. Late-phase aplitic apophyses of the tonalite body cut mylonitic fabric imposed on main phase intrusive rocks, indicating that deformation was coeval with crystallization of the tonalite intrusion. A source of heat from the cooling tonalite body may explain how ductile mylonitic fabric could be developed in subgreenschist regional metamorphic grade volcaniclastic rocks. The observation of very localized tonalitic mylonitic breccia (still in place) indicates that shear zone motion continued as the rocks cooled.

SAMPLING FOR GEOCHRONOMETRY

U-Pb samples weighing about 40 kg each were collected from the central diorite intrusion, tonalite body, and a dacitic sill that intrudes volcaniclastic rocks (Fig. 2). Based on field relationships described above, the magmatic age of the tonalitic body should approximate the timing of motion along the pendant-tonalite shear zone. As a check, the dacitic sill was sampled. It is deformed within the shear zone and is cut by late phase aplites. In addition, similar sills intrude main phase tonalitic in the vicinity. These relationships indicate that the sill should be only slightly younger than main phase tonalite, and should also date the shear zone. On a regional scale the age of activity within this shear zone is important because it is along strike from the dextral, steeply dipping Harrison Lake Shear Zone (Journeay and Csontos, 1989).

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Geology west of Lillooet Lake, near Pemberton, southwestern British Columbia

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Riddell, J.M. Geology west of Lillooet Lake, near Pemberton, southwestern British Columbia; in Current Research, Part E, Geological Survey of Canada, Paper 90-1E, p.219-225, 1990.

Abstract

West of Lillooet Lake, a steep north-northwest-striking thrust fault separates Triassic volcanic and related sedimentary rocks to the east from Cretaceous volcanic and sedimentary rocks of the Fire Lake Group to the west. Movement sense on the fault is east-side-up.

Correlation of rocks to the west of the fault with the Fire Lake Group is based on stratigraphic and fossil similarities with rocks of the Harrison Lake area. Rocks to the east of the fault have been mapped as Triassic Cadwallader Group. It has not been determined whether this correlation is valid.

Résumé

À l'ouest du lac Lillooet, une faille inverse à fort pendage, de direction nord-nord-ouest, sépare des roches volcaniques et des roches sédimentaires apparentées du Trias à l'est de roches volcaniques et sédimentaires du Crétacé, du groupe de Fire Lake, à l'ouest. Le bloc localisé à l'est de la faille a été élevé.

La corrélation des roches à l'ouest de la faille avec le groupe de Fire Lake est fondée sur des similarités de stratigraphie et de fossiles avec des roches de la région du lac Harrison. Les roches situées à l'est de la faille ont été cartographiées comme faisant partie du groupe triasique de Cadwallader. La validité de cette corrélation n'a pas été confirmée.

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INTRODUCTION

Field work conducted during the summer of 1989 covered a map area of approximately 200 km^2 west of Lillooet Lake near Pemberton, southwestern British Columbia.

The area has been mapped as Triassic Cadwallader Group by previous workers, Cairnes (1925), Roddick and Hutchison (1973), and Woodsworth (1977). Mapping by Journeay and Csontos (1989) indicated that the map area is divided in two parts by a north-northwest striking structure, probably a thrust fault, with rocks of the Cretaceous Fire Lake Group lying to the west, and Triassic rocks, possibly of the Cadwallader Group lying to the east. The purpose of this project is to map this area in detail, to study the nature of the contact between these two rock units, and to establish local and regional correlations with equivalent stratigraphic sections in the Coast Belt. This year's mapping has confirmed the existence of a major through-going north-northwest striking thrust fault, and supports correlation of rocks to the west of it with the Fire Lake Group. Analyses of microfossil and radiometric samples collected east of the fault will help to determine whether these rocks are correctly correlated with the Cadwallader Group.

TRIASSIC STRATIGRAPHY

Rocks to the east of the major thrust fault (Main Fault) are assumed to be Triassic based on lithological similarities with rocks in the Tenquille Lake area which contain Norian microfossils (Woodsworth, 1977). No rocks from the map area of this study have been dated.

Four major lithological units have been identified within this Triassic section. At this stage of the project, contacts between these units are not well mapped and an internal stratigraphy has not been developed. The following tentative stratigraphy is suggested and will be tested during the 1990 field season.

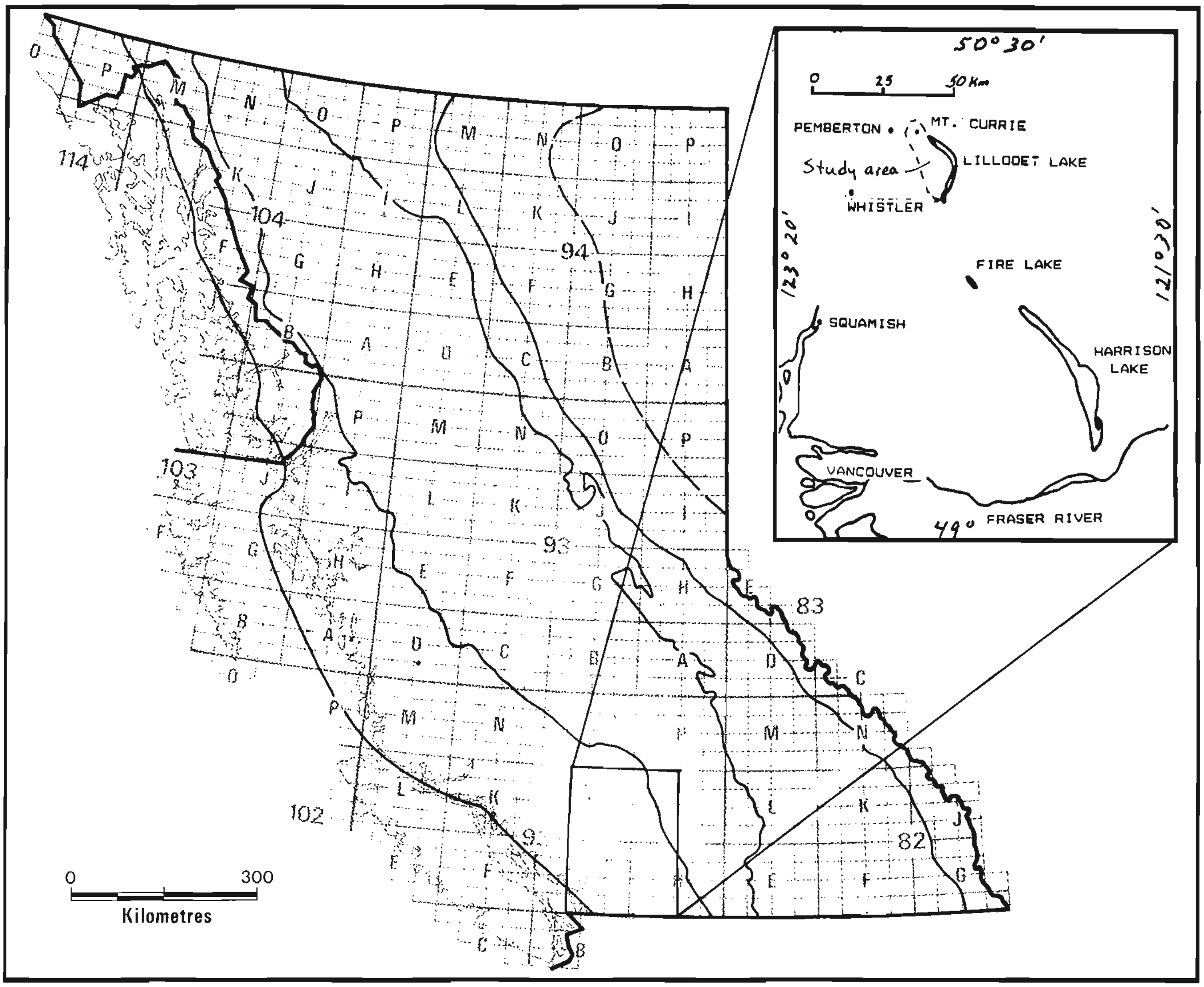

Figure 1. Location map of study area.

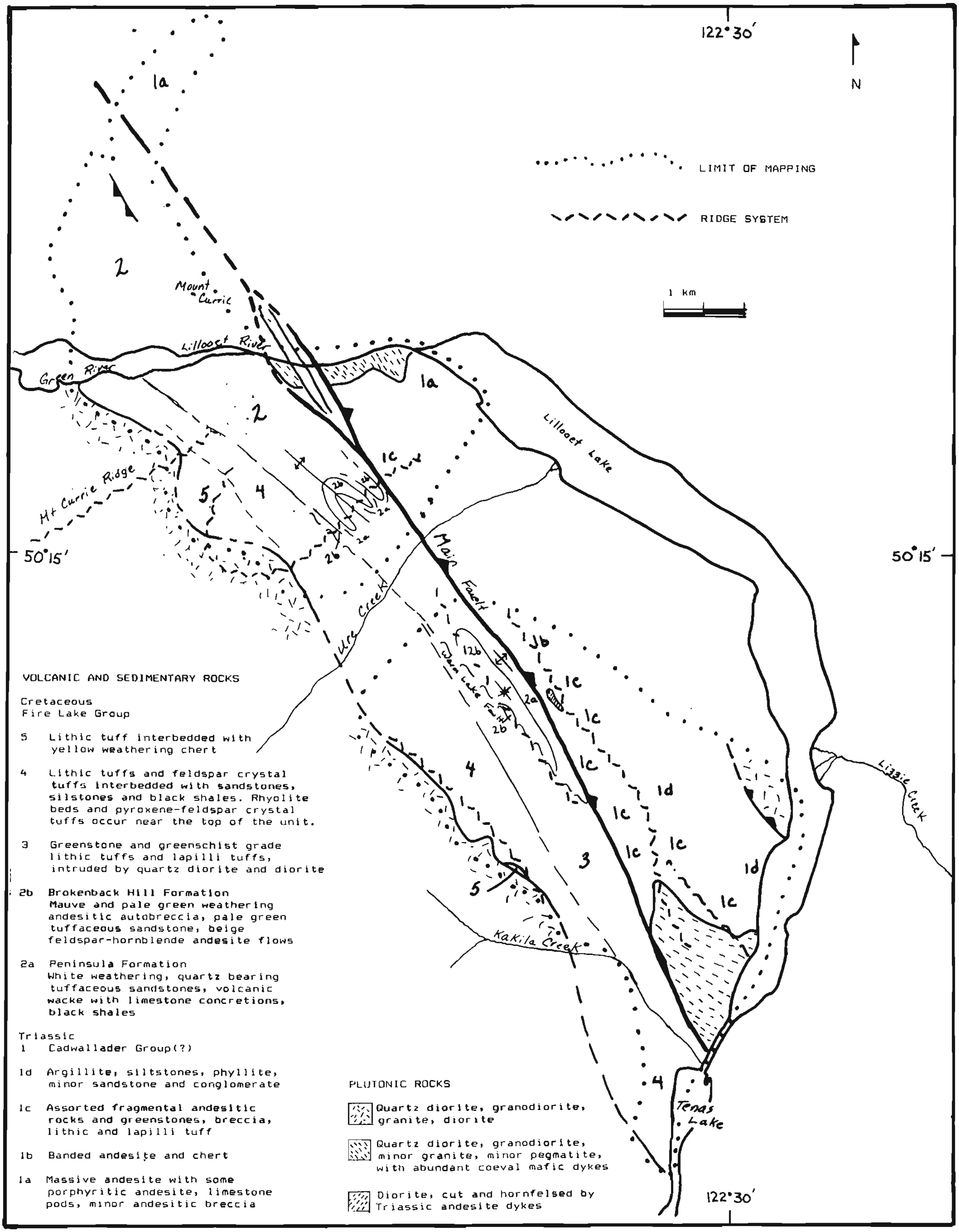

Figure 2. Geological map of the Lillooet Lake region.

Unit 1a (oldest)-massive andesite

Unit 1a is dominated by massive andesite and greenstone, and contains porphyritic feldspar andesite, limestone pods, and minor andesite breccia. Basalt is rare. The massive andesite is speckled with rounded epidote clots and contains abundant epidote veins. Disseminated pyrite is ubiquitous. No pillow structures were seen. This unit is well exposed on the Lill property at the north end of Lillooet Lake, and along the Darcy road near Spetch, about 7 km north of Mount Currie. This unit probably represents the base of the Triassic section.

Unit 1b-banded andesite and chert

A thickness of several hundred metres of banded andesite and chert is exposed at the north end of the easternmost ridge south of Ure Creek. The bedding strikes north-northwest and dips steeply to moderately to the west. The cherts are green, white or black, and form beds 5-6 cm thick. Interbedded andesite is medium to dark green and massive, with beds that are centimetres to metres thick. Porphyritic feldspar andesite is not seen in this section.

Unit 1c-andesitic fragmental rocks

Unit 1c comprises a variety of andesitic fragmental rocks, and greenstone. The fragmental rocks are poorly sorted and show a wide variation in matrix type, clast type and clast size.

Feldspar crystal tuffs and lithic-lapilli tuffs with feldspar crystal tuff matrix are the dominant rock types at the south end of the easternmost ridge south of Ure Creek. Clasts are mainly white weathering felsic volcanic rock with lesser amounts of andesite. Green chert clasts are rare. Clasts make up anywhere from 0-80 % of the rock. Clasts average 3-4 cm in diameter.

The centre of the ridge is partly underlain by a hornblende and feldspar phyric andesitic autobreccia; clasts average 5 or 6 cm in diameter and have deep purple reaction rinds. Distinct from this is another fragmental unit with aphanitic dark green andesitic matrix and a variety of clast types, including andesitic and felsic volcanic clasts, rare green or black chert, diorite, and basalt. Clast size in this breccia is on average 3-4 cm.

All of the andesitic fragmental rocks contain interbedded andesitic flows. Felsic flows are present but not as common.

Unit 1d-sedimentary rocks

Sedimentary rocks are exposed on the west shore of Lillooet Lake on the logging road that extends part way up the lake from the south. Black argillite and phyllite dominate the section. Purple and green siltstones, volcanic sandstones, and minor chert and limestone are also present.

CRETACEOUS STRATIGRAPHY

Unit 2a-peninsula Formation

The lowermost unit seen west of the Main Fault is well bedded and graded, and comprises interbedded, white weathering, feldspar-rich tuffaceous sandstone, siltstone, volcanic wacke and black shale. The section is topped by a pale green weathering tuffaceous sandstone that is conformable with the overlying breccia unit. Trough crossbeds are observed at this contact. Fossil and lithological similarities support correlation of these rocks with the Peninsula Formation as described by Arthur (1986). This unit has been mapped in the Fire Lake area by Lynch (1990).

This unit has several distinguishing characteristics. The white weathering tuffaceous sandstones are quartz-bearing and usually have a slightly calcareous matrix. Shaly beds are millimetres to centimetres thick and are a minor component of the section. North of Ure Creek, these shaly beds are tens of metres thick and make up a major percentage of the section. Plant fossils are locally found in thin shaly beds in the white weathering tuffaceous sandstone. Shaly rip-up clasts are common. Rounded pebbles of feldspar phyric volcanic rocks, coaly fragments, and white to grey chert are found in pebble bands in the volcanic wacke. Brown weathering dark grey massive limestone concretions 10-30 cm across are also found in the volcanic wacke. These concretions contain belemnites at one locality.

Figure 3. Shale rip-up clasts in white-weathering quartz-bearing tuffaceous sandstones of unit 2A (Penninsula Formation).

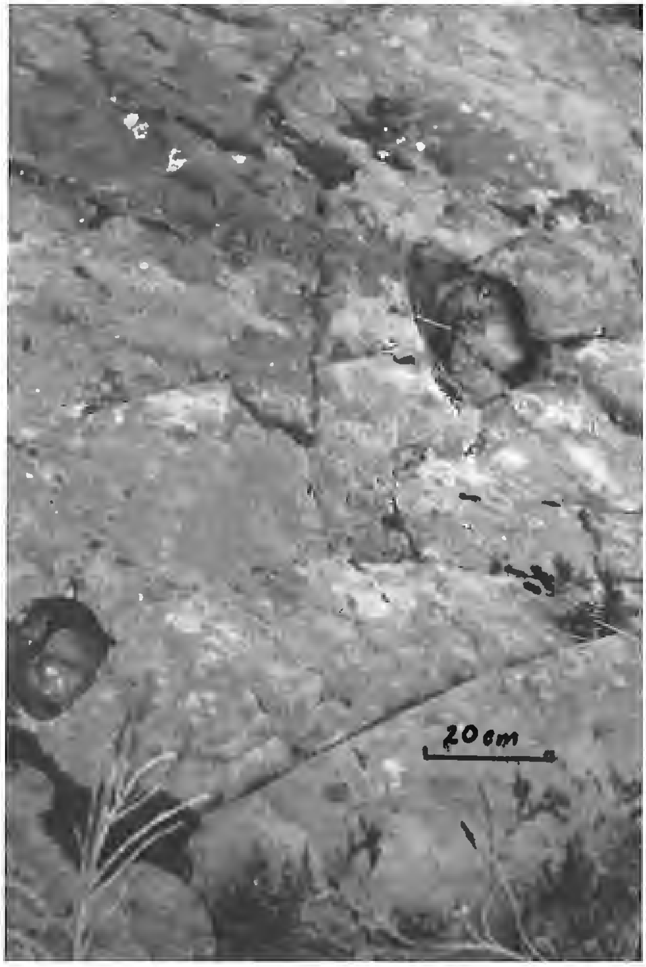

Figure 4. Limestone concretions in volcanic wacke of unit 2a (Penninsula Formation).

Figure 5. Mauve and pale green andesitic autobreccia of Unit 2b (Brokenback Hill Formation).

Unit 2b-mauve and green andesitic breccia unit Brokenback Hill Formation

An andesitic unit dominated by coarse autobreccia conformably overlies the lowermost unit. This unit also contains pale green feldspar crystal tuffs and beige weathering feldspar and hornblende phyric andesitic flows. It is correlated with the Brokenback Hill Formation as described by Arthur (1986) based on lithological similarities.

The breccia unit is marked by distinctive pale pastel green and mauve weathering colours. Breccia clasts are usually 3-6 cm across, but are locally much larger. Clasts over a metre across occur in this unit south of Ure Creek. Clasts can be rounded with reaction rims, or angular with distinct edges. Angular and rounded clasts are seen together in some outcrops. Clasts are feldspar phyric and contain hornblende phenocrysts. Jasper clasts are seen locally in finer grained andesitic breccias (i.e. where clasts average 0.5 cm). Jasper is also seen in the intersticies between closely packed clast blocks. Clast to matrix ratios range from 80/20 to 30/70. The matrix is compositionally equivalent to the clasts, although hornblende phenocrysts are not as abundant in the matrix. This unit is very resistant and forms ridges and benches.

Unit 3-greenstone unit

The Warm Lake fault (informal) separates units 2a and 2b from the overlying units 3 through 5.

The greenstone unit is exposed west of the Main Fault in the south end of the map area north of Kakila Creek. It is cut by a fault within the Fire Lake Group rocks and is not present north of Ure Creek. Unit 3 is composed primarily of greenstone and contains lapilli and lithic tuff units which are metamorphosed to greenschist facies adjacent to quartz diorite and aplite dykes.

Unit 4-interbedded lithic tuffs and sedimentary rocks

Unit 4 is well exposed west of Tenas Lake, in the saddle between the two westernmost ridges south of Ure Creek, and west of a deep saddle north of Ure Creek. The unit contains several lithic tuff units that are 20-30 m thick, and that contain feldspar phyric volcanic and green siliceous clasts in a feldspar-rich matrix. These tuffs are interbedded with black shales, siltstones, tuffaceous sandstones and minor conglomerates. Near the top of the section, thin frothy white rhyolite bands are visible. These are less than a metre thick. Pyroxene-feldspar crystal tuffs are also common near the top of the section. The bands can be several metres thick and contain good euhedral pyroxene crystals that can be up to 1 cm across. Black chert is present at the top of this section on the westernmost ridge. Thick quartz veins striking parallel to foliation are common in the sedimentary rocks of this unit.

Unit 5-interbedded lithic tuff and yellow weathering chert

This unit is well exposed near the south end of the westernmost ridge in the map area. It was found only in float and rubbly outcrop on the ridge directly south of Mount Currie Ridge. The lithic tuff is similar to that found in unit 4, but contains angular purple and brown chert fragments. The yellow weathering chert is pale grey on the fresh surface, laminated and locally convoluted.

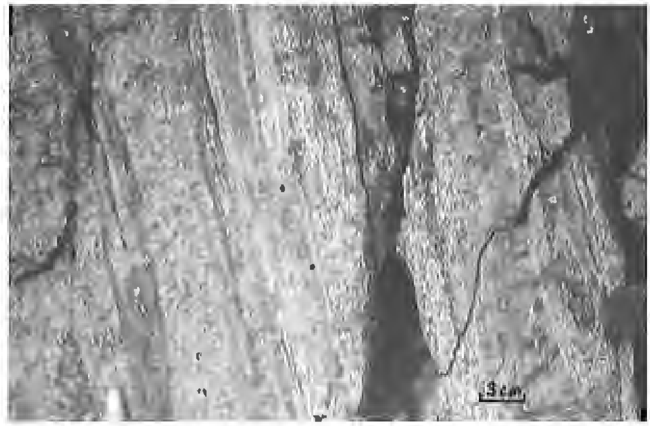

Figure 6. Mylonite zone in quartz diorite along west shore of Lillooet Lake.

PLUTONIC ROCKS

The oldest plutonic rocks in the area are diorites which outcrop at the top of the easternmost ridge south of Ure Creek. The diorites are cut and hornfelsed by andesite dykes. Clasts of the diorite are found in the andesite breccia.

Quartz diorite, granodiorite and granite, and coeval mafic dykes are exposed near the north end of Lillooet Lake along the south shore of the Lillooet River, and at the south end of Lillooet Lake. The age of these rocks is not known.

Strongly foliated quartz diorite is exposed on the west shore of Lillooet Lake, across the Lake from the Lizzie Creek alluvial fan. Near the contact with Triassic sedimentary rocks to the south, the granitic rocks are mylonitized and contain northeast side up kinematic indicators. Mafic phases are locally metamorphosed to amphibolite grade. The sedimentary rocks south of the contact are unmetamorphosed. A northeast side up thrust fault relationship is inferred.

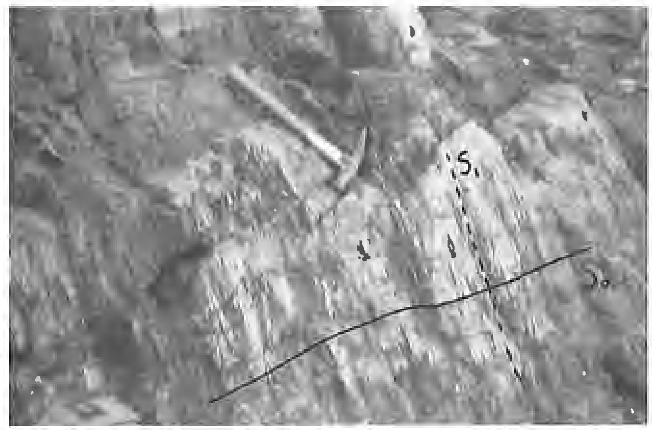

Figure 7. Well-developed bedding/cleavage relationships in hinge zone of fold cored by volcanic wacke of unit 2a (Penninsula Formation) north of Ure Creek.

STRUCTURE

The dominant structural feature in the map area is a steep north-northwest striking thrust fault (Main Fault) that separates Triassic rocks to the east from Cretaceous Fire Lake Group rocks to the west. The fault zone is well exposed along a logging road that runs along the south shore of the Lillooet River. Kinematic indicators at this locality consistently record an east side up movement sense. The fault zone is marked by intense brittle deformation across a width of almost 2 km at this location, and contains zones of sericite and talc schists. The fault zone is not as wide in the rest of the map area; strongly deformed but recognizable Triassic and Cretaceous stratigraphy can be found tens of metres from one another elsewhere.

The vast majority of outcrops in the map area exhibit moderate to intense steeply dipping north-northwest-striking foliation fabrics.

Structures on the east side of the Main Fault are not well understood. In layered rocks of the Triassic section, bedding and foliation orientations are parallel indicating that at least some of the section has undergone isoclinal folding.

West of the Main Fault, the Fire Lake Group rocks display well developed bedding/cleavage relationships. North of Ure Creek, rocks are deformed into broad open upright folds with near horizontal fold axes which parallel the strike direction of the Main Fault. The southwest-vergence of these folds is consistent with a history of east-side-up thrust faulting, as indicated in the north road fault zone exposure.

Deformation prior to movement along the Main Fault is apparent in the Fire Lake rocks. A north-northwest striking fault, the Warm Lake fault (informal), cuts out an unknown amount of section. Pre-existing folding is apparent south of Ure Creek where the overprint of the Main Fault folding event has produced a more complex interference pattern.

Late stage north-south striking extension cracks are common throughout the map area, forming 1-10 m wide chasms and rubble pits.

ACKNOWLEDGMENTS

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Research Grant #RG89-24 from the British Columbia Geological Survey Branch. I am most grateful to Murray Journeay for suggesting the project and for providing encouragement and excellent leadership during the field season. Field assistants Robin Shropshire, Laurie Welsh, Shelley Higman and David Bilenduke worked hard and provided moral support. Greg Lynch provided an introduction to the Fire Lake Group stratigraphy. I am grateful to John and Patricia Goats of Pemberton Helicopters for safe and dependable transportation and expediting.

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The Bridge River Assemblage in the Meager Mountain volcanic complex, southwestern British Columbia

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Stasiuk, M.V. and Russell, J.K. The Bridge River Assemblage in the Meager Mountain volcanic complex, southwestern British Columbia; in Current Research, Part E, Geological Survey of Canada, Paper 90-1E, p. 227-233, 1990.

Abstract

The Meager Mountain volcanic complex erupted 2350 BP to produce the Bridge River Assemblage. This assemblage comprises at least three primary volcanic lithologies representing different eruption styles. The oldest stratigraphic unit is a pyroclastic airfall produced by five discrete phases of eruption, each beginning with phreatomagmatic activity and progressing to magmatic pyroclastic eruptions. The second unit is a pyroclastic block and ashflow deposit which has entrained large, charred logs and pumice blocks and outcrops up to 7 km from the vent area. The third and youngest unit is represented by dacite lavas that form steep bluffs in the present-day Lillooet valley. Although the origin of these welded volcanic breccias remains somewhat enigmatic, they are interpreted to represent hot breccias intimately associated with the production of dacite lavas at the vent. These primary volcanic deposits are partly covered by a volcanic debris flow.

Résumé

Le complexe volcanique du mont Meager est entré en éruption il y a 2350 années BP et a donné l'assemblage de Bridge River qui comprend au moins trois lithologies volcaniques primaires représentant différents styles d'éruption. L'unité stratigraphique la plus ancienne est constituée d'ignimbrites produites par cinq phases distinctes d'éruption, chacune commençant par une activité phréatomagmatique pour passer progressivement à des éruptions magmatiques pyroclastiques. La seconde unité est constituée d'un bloc pyroclastique et d'un dépôt de coulées de cendres qui a entraîné de grands tronçons de bois carbonisé et des blocs de pierre ponce et qui affleure jusqu'à 7 km de la cheminée. La troisième unité, la plus jeune, est représentée par des laves dacitiques qui forment des escarpements abrupts dans la vallée actuelle de Lillooet. Bien que l'origine de ces brèches volcaniques soudées reste assez égnimatique, on pense que ces brèches seraient des brèches volcaniques intimement liées à la production de laves dacitiques dans la cheminée. Ces dépôts volcaniques primaires sont recouverts en partie par une coulée de débris volcaniques.

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INTRODUCTION

The Meager Mountain volcanic complex is situated at the northern end of the of the Garibaldi Volcanic Belt (Anderson, 1975; Read, 1977). Past geological studies of the Meager Mountain edifice have been stimulated by three important lines of research.

Firstly, the volcano is an important member of the Garibaldi (Cascade) Volcanic Belt (Fig. 1, 2) in terms of the edifice's youth and the level of exposure. Consequently, the volcano and its eruptive products have been studied to provide information about the evolution of the Garibaldi Belt and the origins of Quaternary volcanism in southwestern B.C. (e.g. Green et al., 1988). Furthermore Meager Mountain sits in an area of complex tectonic overlap. At the northern end of the Garibaldi Belt, the convergent plate boundary of western North America (Fig. 1) intersects a prominent northeast trending strike-slip boundary (Green et al., 1988) as well as possibly intersecting the Fraser fault. One manifestation of this tectonic complexity is the contemporaneous calc-alkaline and alkaline volcanism initially documented at Salal Creek (Lawrence et al., 1984) and subsequently reported at Meager Mountain (Stasiuk and Russell, 1989).

Secondly, the presence of hot springs at Meager Mountain, marked the region as having geothermal energy potential. Geological projects designed to assess this potential have supported basic research on the geology of this volcanic edifice (Read, 1972, 1977; Mathews and Souther, 1987). For example, B.C. Hydro's exploration program for geothermal energy in the Meager Creek area led to a 1:20 000 scale geological map of the Meager Mountain volcanic complex (Read, 1977).

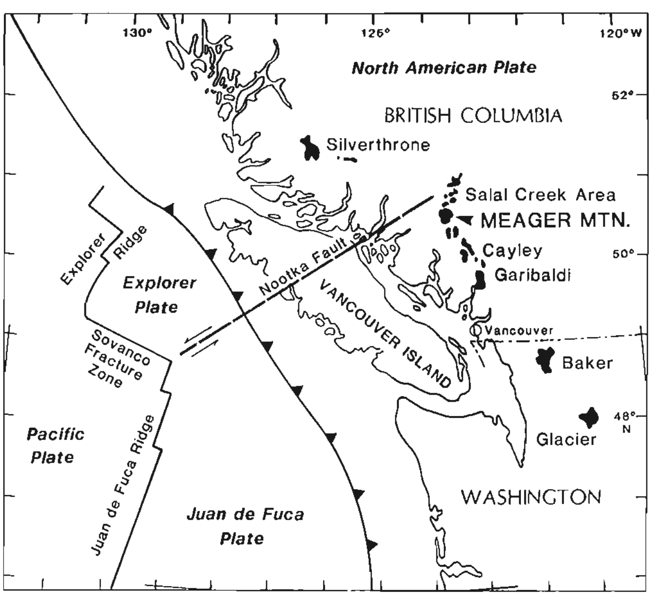

Figure 1. Location of Meager Mountain volcanic complex with respect to major tectonic elements of southwestern British Columbia (after Lawrence et al., 1984).

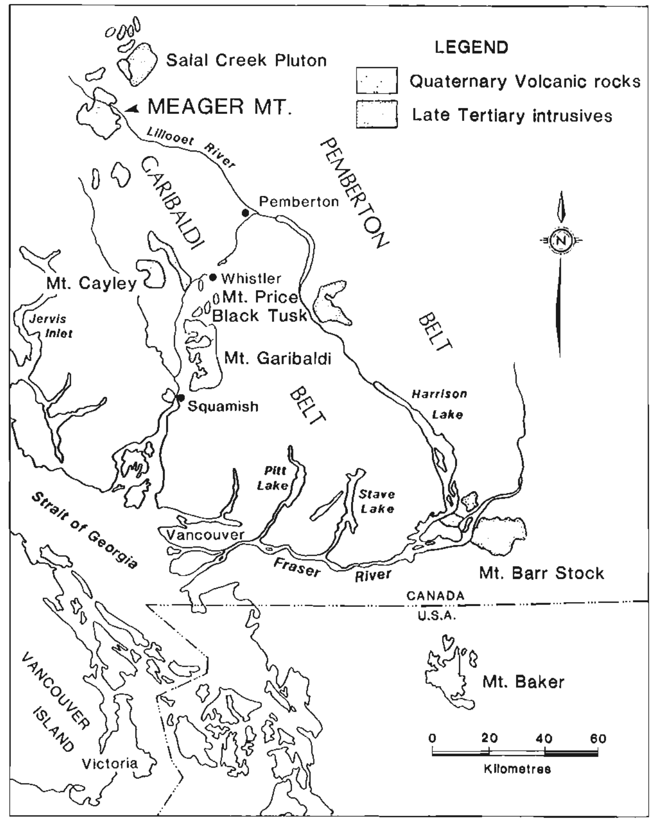

Figure 2. Distribution of Quaternary and Tertiary volcanic rocks in southwestern British Columbia (after Mathews and Souther, 1987).

Finally, the youngest Meager Mountain volcanic rocks (Bridge River Assemblage) represent the youngest volcanic eruptions within the Garibaldi Belt (2350 BP). The fact that these Recent rocks represent, in part, violent pyroclastic eruptions and that the volcano is proximal to Vancouver (150 km north) has also stimulated research into the volcanic hazard potential of the edifice. The diversity of Recent volcanic rock types found in the Bridge River Assemblage provides a useful means to study the near-historic eruptive history of Meager Mountain.

This paper summarizes results of the senior author's 1989 field season in the Meager Mountain area. The main objective of the paper is to discuss the volcanic stratigraphy of the Bridge River Assemblage rocks and thereby elucidate the eruption dynamics and style of this youngest period of volcanism. Consequently, the trace element data are reported as an appendix to the paper.

THE BRIDGE RIVER ASSEMBLAGE

The youngest volcanic rocks exposed at Meager Mountain are felsic pyroclastic rocks and lava flow(s) which are dated by Green et al. (1988) at 2350 BP. These late felsic volcanic rocks comprise the Bridge River Assemblage (Read, 1977). The original map units have been described previously by Nasmith et al. (1967), Read (1972, 1977) and elaborated on by Stasiuk and Russell (1989). The stratigraphy of this group of rocks and the contact relationships between the Bridge River lithologies are well-exposed in the Lillooet River valley (Fig. 3). Despite the previous work, there remain unresolved problems in both stratigraphic correlation between volcanic units and the interpretation of the origins of the near-historic rocks. This paper gives new field observations on the Bridge River Assemblage which we feel clarify several of its enigmatic features.

As a lithological package, the Bridge River Assemblage represents three different styles of volcanic eruption. The first activity is characterized by felsic block, lapilli and ash fallout deposits (e.g. the Bridge River ash of Nasmith et al., 1967). The second eruptive phase produced a pyroclastic flow and was followed by eruption of felsic lava flow(s). Differences in correlation and interpretation mainly concern the third and final primary volcanic lithology. Below we discuss each lithological unit in detail.

Pyroclastic pumice and ash fallout deposits

The observations of Read (1972, 1977) and Stasiuk and Russell (1989) indicate that the first phase of eruption consisted of a single eruption which deposited a blanket of well-sorted, angular, rhyodacitic pumice and ash. The proximal and distal deposits define a plume which trends northeast from the proposed vent area and has a proximal plume axis of about 63° Azimuth (Nasmith et al., 1967). The proximal facies of the pumice fall deposit was examined on the ridge northeast of the vent and slightly south of the plume axis, and low in the Lillooet valley near the southern edge of the plume. The exposure in the banks of the Lillooet River contains a complete section, is 1-2 m thick, and exhibits neither internal bedding nor grading features. The higher elevation exposures, although incompletely preserved, were closer to the plume axis and exhibit distinct bedding and grading. Whereas the presence of the bedding and grading in fallout deposits can be used to constrain the dynamics of the eruption, the incomplete nature of the sections at high elevations precludes any reliable interpretation.

During the 1989 field season, a previously undocumented exposure of the fallout deposits in the Lillooet River valley shows Bridge River pumice beds deposited upon a steep paleoslope of basement bedrock covered by soil. This exposure appears to represent a complete section of the fallout deposits developed close to the plume axis. The deposits have internal stratigraphy and graded bedding and thus record information on the dynamics of the pyroclastic eruption. The fallout deposits comprise approximately 6 m of loosely consolidated to unconsolidated pumice and ash and are overlain by about 4 m of pyroclastic flow deposit and up to 32 m of dense, indurated, vitroclastic breccia.

At this locality the pumice fall deposits can be subdivided into at least five distinct units of varying thickness. These units, from bottom to top, have vertical thicknesses of about 4 m, 0.35 m, 0.45 m, 0.30 m and 1.20 m respectively. Each unit contains reverse graded bedding, which may relate to variations in the height of the eruptive plume (Fisher and Schmincke, 1984). On average, the basal zone of each unit is 10-20 cm thick and comprises lapilli less than 1 cm in diameter. The basal zones coarsen upwards to clasts about 5 cm across, except in the uppermost subunit. The top subunit coarsens upwards to within 30-40 cm of the top, where the pyroclastic material rapidly fines to coarse sand size, coarsens to about 1 cm diameter, fines again to sand and coarsens to 1 cm once again. These last two graded beds are lithologically distinct in that they are dominated by lithic and crystal clasts with as little as 10 % felsic pumice clasts.

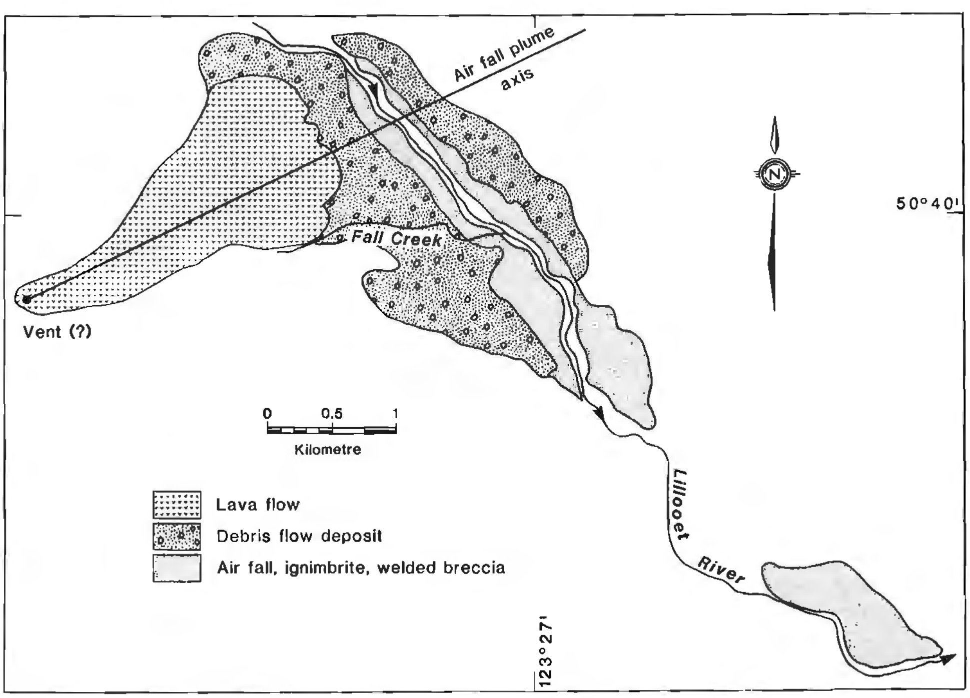

Figure 3. Detailed geological map of the Meager Mountain volcanic complex showing the distribution of the Bridge River Assemblage (after Read, 1977).

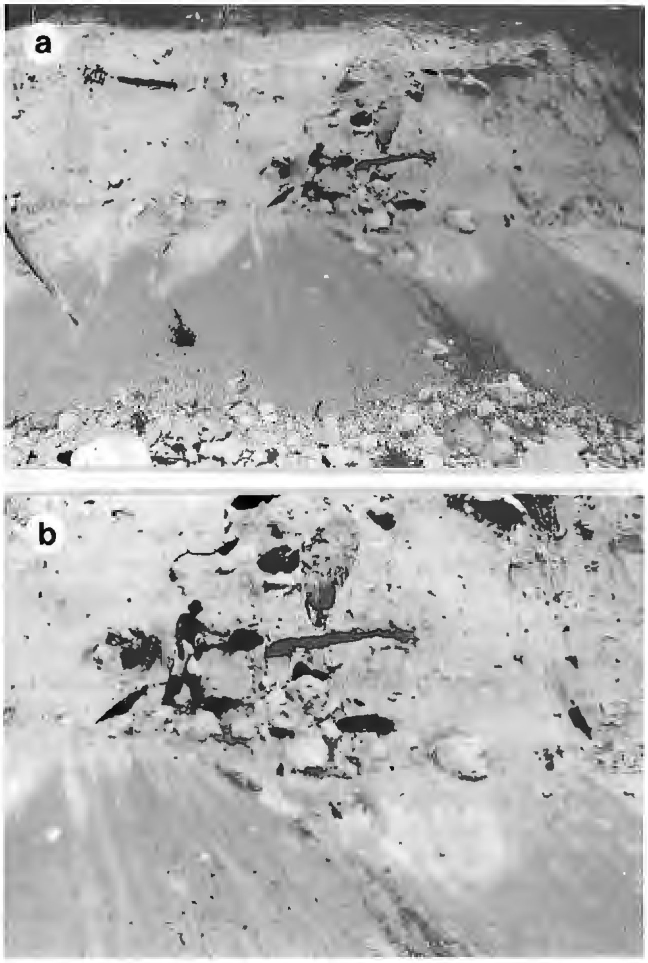

Figure 4. Pyroclastic flow within the Bridge River Assemblage **(a).** Note the large charred logs that have been incorporated into the pyroclastic flow **(4).** MVS appears in both plates as a scale. Photographs taken approximately 7 km downstream from the volcanic vent inferred by Read (1977). See text for description and discussion.

These distinct lithological changes probably relate to a significant change in eruptive style. The upper beds may, in fact, mark the onset of eruption column collapse and are perhaps, the attendant ground layer or basal surge deposit (Fisher and Schmincke, 1984).

Another interesting feature of the subunits which comprise the fallout deposits is the lithological variation within the basal fine zones. At the base of each subunit the deposit is relatively depleted in felsic pumice, and only contains about 50% felsic pumice. The remainder of the basal deposit is made up almost entirely of equant accessory fragments of vitric dacite. Conversely, the top of each subunit is up to 80-90% felsic pumice. These deposits have many of the characteristics associated with phreatomagmatic eruptions where gas-driven explosive activity results from ascending magma intersecting porous water saturated near-surface rocks (Heiken and Wohletz, 1987). The ejecta comprises significant accidental and accessory lithic material from the original edifice and conduit walls. In the Bridge River fallout deposits these basal layers repeat several times, suggesting there were a number of phreatomagmatic eruptions. The frequency of these eruptions over such an apparently short time interval suggests that there was abundant surface water with which to recharge the groundwater, perhaps from the surrounding icecaps and snowfields. Such explosions provide an efficient mechanism with which to abrade and widen the eruption conduit and, therefore, could have accelerated the onset of pyroclastic flow eruption.

The reverse grading of the air fall subunits is a common feature in air fall deposits and can be considered in the light of quantitative models of fallout dispersal for eruption columns (Carey and Sparks, 1986). Such models indicate that reverse graded bedding is a result of increasing eruption intensity and column height, as in the case of the Avellino plinian deposit in Italy (Pescatore et al., 1987). However, others attribute increasing column height (and the ultimate cause of reversely graded fallout deposits) to the widening of the conduit by eruptive erosion (Wilson et al., 1980). We prefer the latter hypothesis for the graded bedding in the Bridge River fallout deposits because it fits well with the observed sequence of volcanic activity. Widening of the conduit leads to increasing column height with attendant reverse graded fallout deposits and subsequently would produce pyroclastic flows with the collapse of the eruption column (Wilson et al., 1980).

Pyroclastic flow deposit

The second phase of the eruption, a pyroclastic flow, was not recognized as such by Read (1972, 1977) but was simply grouped with the pumice and ash fallout deposits. Stasiuk and Russell (1989) recognized the unit as a small-volume pyroclastic flow exposed only at low elevations in the Lillooet River valley. Based on the apparently limited extent of the deposit and on the fact that the deposit did not knock down or even tilt trees of the paleoforest, they concluded that it was a relatively non-mobile and low temperature pyroclastic flow deposit. Furthermore, the deposit is non-indurated.

Two new exposures were found in summer 1989 which cast new light on the mobility of the flow. The first exposure, relatively near the source area, occurs in the same creek off the Lillooet River as the airfall just described. This locality is at an elevation 30 m above the river valley bottom where all previous exposures were observed. The deposit is 4 m thick, is emplaced against basement rock which forms a paleocliff, and overlies the pumice and ash fallout deposits. At this locality, the pyroclastic flow exhibits a lack of internal structure and does not scour the underlying unconsolidated pyroclastic material.

The second exposure occurs about 5 km downstream, just above the confluence of the Lillooet River and Pebble Creek. The Bridge River deposits at this locality were previously mapped as reworked tephra by Read (1977), but the majority of the section comprises unreworked pyroclastic flow. Unlike the previous exposures of pyroclastic flow deposits, this section exhibits internal structural features commonly seen in the distal parts of cool, poorly fluidized, pyroclastic flows. Figure 4a illustrates the general aspect of the exposure looking north. The deposit averages about 6 m in total thickness and comprises a 1 m thick fine-grained (ash) basal layer, a 4-5 m unsorted, matrix-supported layer of ash and pumice blocks, and an upper 10-20 cm bed of bedded fine ash. Pieces of completely carbonized wood occur throughout the deposit, with small fragments and branches in the fine grained sections and large trunks, up to 50 cm across and 3 m long, in the unsorted massive interior (Fig. 4b). The wood, where it is in branch or log form, is almost invariably oriented horizontally. The deposit is easily correlated with the pyroclastic flow exposed upstream by the occurrence of dense, fibrous blocks of pumice with streaks of mafic-intermediate material and by the identical radiocarbon ages for both deposits (charred wood; Read, 1977).

In summary the deposit has four main features which relate to its origins. First, regardless of size, the logs generally lie horizontal and the flow is characterized by abundant large (up to 2 m diameter), dense, pumice blocks. These features suggest a high energy, dense, turbulent, mobile pyroclastic flow. Second, the deposit pinches out abruptly at the downstream end of the exposure where there exists a vertical, only slightly charred tree. These points indicate that this locality represents the distal end of the pyroclastic flow. Third, the uppermost part of the deposit comprises bedded fine ash and probably represents the upper ash cloud of the pyroclastic flow into which material was elutriated. If true, this implies that at this point the flow was still fluidized. Fourth, the fine ashy basal zone is partly mixed with oxidized, rusty pebbles of the paleosoil beneath the deposit. This may indicate that the pyroclastic flow was able to scour the underlying surface. Such action would be consistent with turbulent rather than laminar flow behaviour and corroborates the turbulence suggested by the entrainment large pumice blocks and trees. This lowermost, ash-rich layer shares some similarities to base surge deposits, however it lacks complex internal structures such as climbing ripples and is quite unlike the ground layer of "classical" pyroclastic flow deposits (Fisher and Schmincke, 1984).

In summary, the upstream exposures of the pyroclastic flow probably represent the margin deposits of the proximal facies. These deposits were emplaced against the paleovalley walls and were relatively cooler, thinner, and less turbulent than the main stream of the pyroclastic flow (which has been largely removed by the Lillooet River). These relatively low energy deposits do not show the same ability to entrain large pumice blocks or to knock down forests. Furthermore, the pyroclastic flow deposits do not show well developed internal structure, which suggests a poorly fluidized nature prior to emplacement. Conversely, the downstream, distal pyroclastic flow was deposited after the pyroclastic flow had developed internal flowage structures related to volcanic gas fluidization, and where the river valley widened considerably to allow the flow to spread and slow.

Bridge River assemblage lavas

The next phase of eruption consisted of at least one glassy rhyodacitic lava eruption. Presently, there are at least two conflicting interpretations concerning the events of this final phase. Read (1977) mapped a vitroclastic breccia, up to 120 m thick, as a blocky lava flow representing the initial phase of this eruption. The map unit is a dense, well indurated deposit comprising porphyritic, glassy rhyodacite clasts in a grey matrix of comminuted rhyodacite fragments. Overlying this breccia, Read recognized a poorly consolidated debris flow deposit and a massive rhyodacite lava flow. Stasiuk and Russell (1989) found, however, that petrographic and bulk chemical similarities between the clasts and matrix of the vitroclastic breccia and the vitric lava flow made the two indistinguishable in terms of these two criteria. This suggests that the lava flow may be the parent of the breccia, and that the two lithologies represent different facies of the same volcanic event. A number of additional observations support the latter hypothesis.

First, in the stream valley on the south edge of the lava flow, the lava flow bottom was clearly exposed and no debris flow material lay beneath the flow, indicating that the debris flow does not necessarily separate the breccia and the lava flow but may instead be stratigraphically higher than both of them. If younger than the rhyodacite lava, the debris flow would be expected to leave at least a veneer of material on the surface of the lava flow and smooth the topographic expression of the lava flow front. There is, however, no debris flow material overlying the lava flow (Read, 1972, 1977), and the toe of the lava flow rises 20-30 m above the level of the debris flow material, documenting the minimum thickness of the lava flow. Whereas these observations are consistent with the lava flow being younger than the debris flow, it is more likely that they are the result of the topography at the time that the debris flow was deposited. The debris flow is not observed to be deposited at higher elevations than in the area of the lava flow front. Thus, the debris flow appears to have been controlled by topography to the extent that it was restricted to lower elevations of the valley.

The Lillooet River valley provides a number of cross-sections through the indurated vitroclastic breccia. These sections are marked by three structural characteristics which constrain the deposit's origins and suggest an origin other than a simple lava flow breccia. In none of the exposures examined (over 4-5 km, representing two-thirds of the units length), does the breccia have a core of massive lava mantled by fragmental material. This is in contrast to most blocky lava flows which are commonly cored by viscous tongues of lava around which there is a thick carapace of breccia (e.g. Williams and McBirney, 1979).

Secondly, the breccia unit contains evidence of multiple surges or pulses of deposition which have produced a single composite cooling unit. The effects of the multiple depositional events within the vitroclastic breccia unit are particularly evident in a tributary gully which cuts a cross-section through the lateral edge of the breccia unit. At this locality, four depositional units can be recognized by variations in degree of welding (strongly welded interiors and more weakly welded tops). Each internal unit is flat-topped and tabular and deposited parallel with the paleotopographic slope. The separate layers represent a rapid depositional succession as they make a single composite cooling unit where the breccia unit is thickest. Where the vitroclastic breccia is thinnest (lateral facies), the individual layers are less welded and are more discernible. The exposures along the Lillooet River depict an excellent example of how several separate cooling units can grade laterally into a single composite cooling unit as deposit dimensions thicken.

The third key feature distinguishing this deposit from a blocky lava flow is the general form of the deposit. The breccia has a smooth top and has edges which thin gradually to less than a metre of welded material. The top surface of the deposit, as well as the tops of the separate depositional layers, have less dip than the basal contact of the units. The dips are gentle and steepen as the beds approach and intersect the paleotopography. Furthermore, the vitroclastic breccia is deposited at higher elevations on the walls of the paleo-valley than at the axis of the ancient valley bottom. This may represent the expenditure of excess momentum during deposition of the breccia as it swept across the valley bottom. Blocky lava flows move by forming steep blocky fronts which exceed the maximum angle of repose for the blocks and cause flow front rockfalls. Thus, the edges of such flows are expected to be steep. Furthermore, blocky flow tops are not flat, but are undulatory; and blocky flows do not commonly push up valley walls but move down valleys.

The vitroclastic unit contains a number of tree molds which are oriented normal to the flattening fabric defined by the rhyodacite clasts. Several of the tree molds are in place and vertically oriented, suggesting that deposition of the vitroclastic breccia was not vigorous enough to fell the paleoforest. The deposit was, however, emplaced at high enough temperatures to carbonize and vaporize the trees as well as promote welding of the deposit. The hollow tree molds are immediately surrounded by poorly welded breccia with little flattening, probably the result of local cooling of the breccia against the trees.

From the above observations, the vitroclastic breccia cannot be interpreted as a simple blocky lava flow deposit. Its actual origin remains somewhat unclear. If the vitroclastic breccia is to remain a different facies of the volcanic event which produced the rhyodacite lava flows, a process is required to explain i) the complete fragmentation of the lava; ii) the efficient transport of the fragmental material for several kilometres; and iii) the retention of sufficient heat in the deposits (as thin as 1-2 m) to carbonize trees and promote welding within the breccia.

Lithologically similar types of deposits have been produced by dome collapse and spine collapse (Heiken and Wohletz, 1987; Fisher and Schmincke, 1984; Williams and McBirney, 1979), but these do not produce composite welded sequences or pulses of brecciated material. The Bridge River vitroclastic breccia is also similar to breccias produced by Merapi-type activity (Williams and McBirney, 1979; Fisher and Schmincke, 1984; Heiken and Wohletz, 1987; Rose, 1987). Merapi activity is defined as pyroclastic flow eruption which emanates from the toes of lava flows, or from lava flow fragmentation and collapse due to flow on overly steep slopes. However, the deposits resulting from flow collapse as well as other examples of Merapi activity are generally unwelded (W.I. Rose, pers. comm., 1989; Fisher and Schmincke, 1984). Additionally, the Bridge River vitroclastic breccia, unlike Merapi activity which produces gaseous pyroclastic flows from lava flows, does not contain any significant amount of pumice clasts.

Although the Bridge River vitroclastic breccia does not appear to be the result of lava flow collapse on steep slopes, it shares similarities with products of these hot rock avalanches. Francis et al. (1974) described deposits from Merapi type activity on San Pedro and San Pablo volcanoes which, although not welded, show similar thickness ranges (up to about 100 m), similar radial prismatic cooling joints in juvenile clasts, similar flow unit lengths (2-7 km), and similar parental flows with truncated toes. The only major difference between the two cases, the presence of welding at Meager Mountain, may be attributable to higher eruption temperatures or earlier fragmentation of the Bridge River lava flow.

CONCLUSIONS

The Bridge River Assemblage of the Meager Mountain volcanic complex is the result of three distinct stages of silicic volcanic activity. The first stage of activity began about 2350 BP and is marked by well-sorted, rhyodacite subaerial fallout deposits of pumice and ash which delimit a northeast-trending plume. The fallout deposits record at least five distinct eruptive phases separated by short time intervals. These separate eruptive phases comprise an initial phreatomagmatic phase followed by a strictly magmatic eruptive phase. Reverse graded bedding is commonplace and suggests that the fallout deposits accumulated from eruption columns of increasing height. The second stage of activity may relate to the ultimate widening of the volcanic conduit through the precursive phreatomagmatic and magmatic explosive activity of the first stage. The middle stage of the Bridge River Assemblage is marked by a pyroclastic flow which extends up to 7 km downstream from the vent area. Although the flow was hot and mobile, it appears to have been small in volume. The third and final stage of volcanic activity is represented by a rhyodacite lava flow at the vent and by a dense welded vitroclastic rhyodacite breccia in the Lillooet River valley. The clasts in the breccia are petrographically and chemically indistinguishable from the lava at the vent. Thus, the breccia deposit is assumed to represent the accumulation of hot fragmental material spawned from the rhyodacite lava. The welding and jointing of the deposit after accumulation indicate that the material must have been deposited at high enough temperatures to promote the annealing of the vitric clasts.

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1989 field activities and accomplishments of geophysical and geotechnical land-based operations and marine/fluvial surveys, Fraser River delta, British Columbia

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Luternauer, J.L., 1989 field activities and accomplishments of geophysical and geotechnical land-based operations and marine/fluvial surveys, Fraser River delta, British Columbia in Current Research, Part E, Geological Survey of Canada, Paper 90-1F, p.235-237, 1990.

Abstract

Land surveys were continued on the subaerial part of the delta to a) more precisely delineate anomalous subsurface reflectors probably resulting from the presence of gas, b) extend our shear-wave velocity database, and c) refine our knowledge of the liquefaction potential of the delta subsurface in the vicinity of the seaward dyke in the Municipality of Richmond. Marine/fluvial surveys focused on a) precisely measuring sediment flows in the lower reaches of the main channel to add to our understanding of the delta front's sediment budget and b) identifying evidence of instability of the delta slope in the vicinity of major coastal structures.

Résumé

Des levés géophysiques et géotechniques sur terre se sont poursuivis sur la partie subaérienne du delta en vue: a) de délimiter de façon plus précise des réflecteurs souterrains anormaux qui sont dus probablement à la présence de gaz, b) d'agrandir la base de données actuelle sur les vitesses des ondes de cisaillement, et c) d'améliorer le niveau de connaissances du potentiel de liquéfaction du sous-sol du delta au voisinage du dyke orienté vers la mer de la municipalité de Richmond. Des levés marins et fluviaux ont porté: a) sur la mesure précise des coulées de sédiments dans les biefs inférieurs du chenal principal en vue d'accroître leur compréhension au niveau du bilan sédimentaire du front du delta et b) sur la détermination d'indices d'instabilité de la pente du delta au voisinage d'importantes structures côtières.

INTRODUCTION

Growing demand for information on the potential earthquake response and present stability of the highly populated and industrialized Fraser River delta (Fig. 1) has maintained the need for multi-disciplinary geological research involving many agencies. These studies have been coordinated by the Cordilleran Geoscience Division.

On the subaerial part of the delta, the focus has been on acquiring data that will enhance assessment of potential earthquake damage from both liquefaction and ground motion amplification. Further to these objectives the subsurface structure of the delta is being studied by coring and high-resolution continuous seismic profiling (Jol, 1988; Williams, 1988; Pullan et al., 1989). The shear-wave velocity database is being extended and seismic/electric cone penetrometer geotechnical surveys are being done along the seaward dyke of the Municipality of Richmond (Finn et al., 1989).

Offshore investigations have focused on defining the stability of the delta front where large ports, a lighthouse, jetties, causeways and submarine pipelines are situated and where there are important wildlife habitats (Fig. 1). These studies involve measuring the bed and suspended sediment load in the lower reach of the main channel (Kostaschuk and Luternauer, 1987, 1989; Kostaschuk et al., 1989a; Kostaschuk et al., in press), describing the patterns and processes of sediment diffusion (Kostaschuk et al., 1989b) and documenting evidence for erosion and deposition on the delta slope (McKenna and Luternauer, 1987; Luternauer et al., 1989).

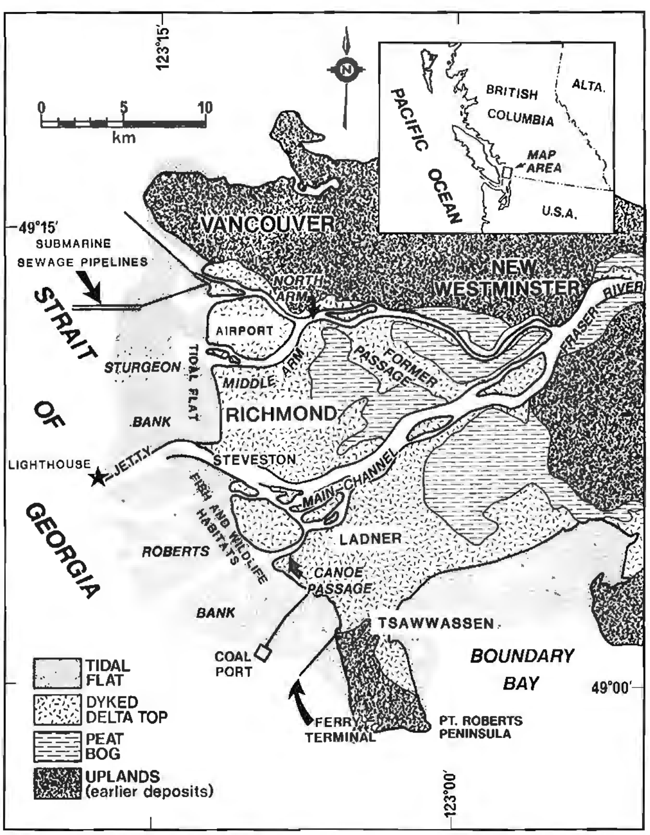

Figure 1. Geographic regions of the Fraser River delta and major developments.

FIELD ACTIVITIES AND ACCOMPLISHMENTS

The following summaries were prepared in consultation with the individual project chiefs listed in parentheses.

Seismo-stratigraphy (S.E. Pullan, Terrain Sciences Division)

A 1 km length of line 600 from the 1986 survey (Pullan et al., 1989) was reshot in the optimum-offset mode (Hunter et al., 1984, 1989) to check the position of an anomalous reflector with respect to a GSC borehole on the line. The borehole, found to be correctly positioned, had been drilled to identify any correspondence between lithology and the presence of the anomalous reflector. No conclusive correlation can be made, supporting the hypothesis that the shallow reflector may be related to the presence of gas within the sediments.

Another 1.5 km length of line 600 was reshot to allow recognition of deeper reflectors than was possible in the original survey. The deepest reflector observed on the 1986 survey was at approximately 200 ms. The present survey, made with a new seismograph having increased dynamic range, indicated the presence of low amplitude, low frequency reflectors to 400 ms which may be enhanced with further processing. It is thought that these deeper reflectors deteriorate in quality as a result of gas trapped below the major reflector at 200 ms.

In order to investigate the nature of a large reflector at shallow depth which constitutes acoustic basement for high-frequency work in the western part of the Tsawwassen-Ladner survey area, a hole was drilled by M.C. Roberts of Simon Fraser University using a Mobil Auger drilling rig. The target reflector was at about 40 m depth. Roberts encountered a gradational boundary (sand fining downwards to clayey silt) at approximately 28 m depth. Although no lithological boundary was encountered at 40 m depth, gas was observed to emanate from the drill hole, suggesting that the seismic boundary is probably gas trapped at an impervious boundary within the clayey silt. The hole was drilled to 50 m and cased with PVC for future logging.

A test record of 1 second duration was made along the Boundary Bay dyke to check the feasibility of mapping the top of Tertiary bedrock, which unpublished industry data suggest lies at about 600 m depth. This test utilized 4.5 Hz geophones and a 12 gauge downhole gun. Significant energy was detected from reflectors down to 400 ms, and one reflection group was detected at 690 ms (about 600 m depth). It appears possible, therefore, to design a continuous profiling survey to map this horizon in this area.

Several tests for shear-wave reflections were made in the Tsawwassen-Ladner area. Poor results were obtained, because only one or two low frequency shallow-depth reflectors were recognized. A dispersed package of ground roll interference results in prominent masking of reflection events. It is suggested that this area is not amenable to shear wave reflection shooting because extremely low-velocity surface sediments give rise to strong frequency attenuation as well as dispersion.

Shear-wave refraction measurements (J.A. Hunter, Terrain Sciences Division)

A downhole "shear-wave gun" was built as a modification to the existing 12-gauge "Buffalo gun" (Pullan and MacAulay, 1987) and used as the seismic source for the shear wave refraction measurements. Energy from the blank load is directed to one side of the gun using a 1.9 cm T-joint and an end plug. Sixteen shear wave refraction sites were occupied in the Ladner-Tsawwassen and Richmond areas, including sites where GSC boreholes have been drilled (Luternauer, 1988; Finn et al., 1989).

Urban seismic risk (D.J. Woeller, Conetec Investigations Ltd.)

These investigations, jointly supported by the GSC and the Municipality of Richmond, form part of our study of the geoarchitecture, evolution and seismic risk assessment of the Fraser River delta. The focus during the previous reporting period (Finn et al., 1989) and this past summer has been on understanding the character of the subsurface in the vicinity of the seaward dyke of the Municipality. During this past summer an additional 7 sites in the southern part of the dyke were probed with the electric and seismic cone penetrometer to depths as great as 50 m. The accumulated data suggest that the sediments below the dyke become progressively less dense and potentially more prone to liquefaction towards the south.

Fluvial sediment budget (R.A. Kostaschuk, University of Guelph)

Ongoing fluvial studies focused this past summer on more precisely discriminating the coarser-sediment flow components and mechanisms of transport in the lower reaches of the main channel. Preliminary results are reported in Kostaschuk et al. (1990).

Delta slope stability (J.L. Luternauer)

An intensive survey of the Fraser Delta slope involving detailed sidescan sonar and 3.5 kHz soundings complemented by sampling and deployment of a remote oceanographic video camera at selected targets was carried out from 24 July to 4 August, 1989. The data were gathered in response to concerns for the stability of the delta slope arising from the dredging program of Public Works Canada (Stewart and Tassone, 1989) and evidence of massive failure off the mouth of the main channel (McKenna and Luternauer, 1987). Detailed base maps of the slope surface can now be prepared to permit mechanisms and rates of recession to be more precisely monitored and understood.

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Relationships between bed-material load and bedform migration, Fraser River estuary, British Columbia

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Kostaschuk, R.A., Ilersich, S.A., and Luternauer, J.L. Relationships between bed-material load and bedform migration, Fraser River estuary, British Columbia; in *Current Research, Part E, Geological Survey of Canada, Paper 90-1E, p.239-244, 1990.*

Abstract

A predominance of bed-material transport in suspension in the Fraser River estuary, British Columbia, has formed in large, symmetric bedforms that do not exhibit flow separation on lee sides. An estimate of sediment transport based on bedform migration overestimates bed load and underestimates suspended and total bed-material loads. This indicates that most of the bed-material load is not involved in bedform migration but is advected through the channel in suspension or as a sheet.

Résumé

La prédominance du transport de matériaux de fond en suspension dans l'estuaire du fleuve Fraser, en Colombie-Britannique, a donné naissance à de grandes formes de fond symétriques qui ne montrent aucun signe de séparation d'écoulement sur les faces aval. Une estimation du transport des sédiments, fondée sur la migration des formes de fond, donne une surestimation de la charge de fond ainsi qu'une sous-estimation de la charge en suspension et de la charge totale en matériaux de fond. Cela indique que la majeure partie de la charge en matériaux de fond ne participe pas à la migration des formes de fond, mais qu'elle est transportée par advection dans le chenal, soit en suspension soit sous forme de nappe.

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INTRODUCTION

The Fraser River is the largest river reaching the west coast of Canada and has created a large delta in the Strait of Georgia (Fig. 1). The Main Channel of the delta is a salt-wedge estuary (Kostaschuk and Luternauer, 1988; Kostaschuk and Atwood, in press) and in spite of the extensive development occurring on the delta, the sediment budget of the estuary remains poorly defined (R. Kellerhals, unpublished report for Water Resources Branch, Environment Canada, 1984). To address this problem, an ongoing project involving the University of Guelph, Geological Survey of Canada, University of British Columbia, Sediment Survey Section of the Inland Waters Directorate, and Public Works Canada was begun in 1985 to examine sedimentary processes in the estuary and on the delta.

Sediment in the Fraser estuary is transported as wash load and bed-material load (Kostaschuk et al., 1989a). Wash load consists of fine material transported in continuous suspension that does not appear in appreciable quantities on the bed. Bed-material load is coarser bottom sediment that is transported close to the bed as bed load and episodically in suspension. Thus, the total suspended load consists of both wash load and suspended bed-material load. Kostaschuk et al. (1989b) examined over 100 bottom samples in the estuary and suggested that the boundary between wash and suspended bed-material loads is 0.125 mm.

Recently, Kostaschuk et al. (1989b) have used the characteristics of large bedforms, or dunes, to predict bed load transport in the estuary. Their procedure is based on the rate equation:

$$(1 - p)\ \delta y/\delta t + \delta q_b/\delta x = 0 \qquad (1)$$

in which y is the bed elevation above an arbitrary datum, p is sediment porosity, t is time, q_b is the volumetric transport rate of bed-material and x is distance in the downstream direction. Solution of this equation yields (Simons et al., 1965):

$$q_b = (1 - p)\ U_{bm}\ y + C \qquad (2)$$

in which U_{bm} is bedform migration rate and C is a constant of integration that represents transported bed-material not involved in the progress of the bedforms. A gravimetric estimate of sediment transported in a migrating bedform is provided by:

$$g_b = D(1 - p)\ \beta H\ U_{bm} \qquad (3)$$

in which g_b is the bed load transport rate per unit width of channel, D is sediment density, β is a bedform shape factor and H is bedform height (Fig. 2)

Kostaschuk et al. (1989b) found strong statistical relationships between Fraser River discharge and sediment transport estimates using equation 3. However, it is uncertain what components of bed-material transport g_b represents. The assumption is often made (e.g. van den Berg, 1987) that it represents only the bed load. In environments such as the Fraser estuary (Kostaschuk et al., 1989) where significant bed-material transport occurs in suspension, this assumption cannot hold because bed-material in intermittent suspension must have residence time on the bed and must become involved in bedform migration. In 1989 we set out to evaluate the contributions made by suspended bed-material load and bed load to bedform migration in the Fraser by comparing measurements of bed load and suspended bed-material load with estimates from equation 3. This paper describes the methodology and gives some preliminary results and conclusions.

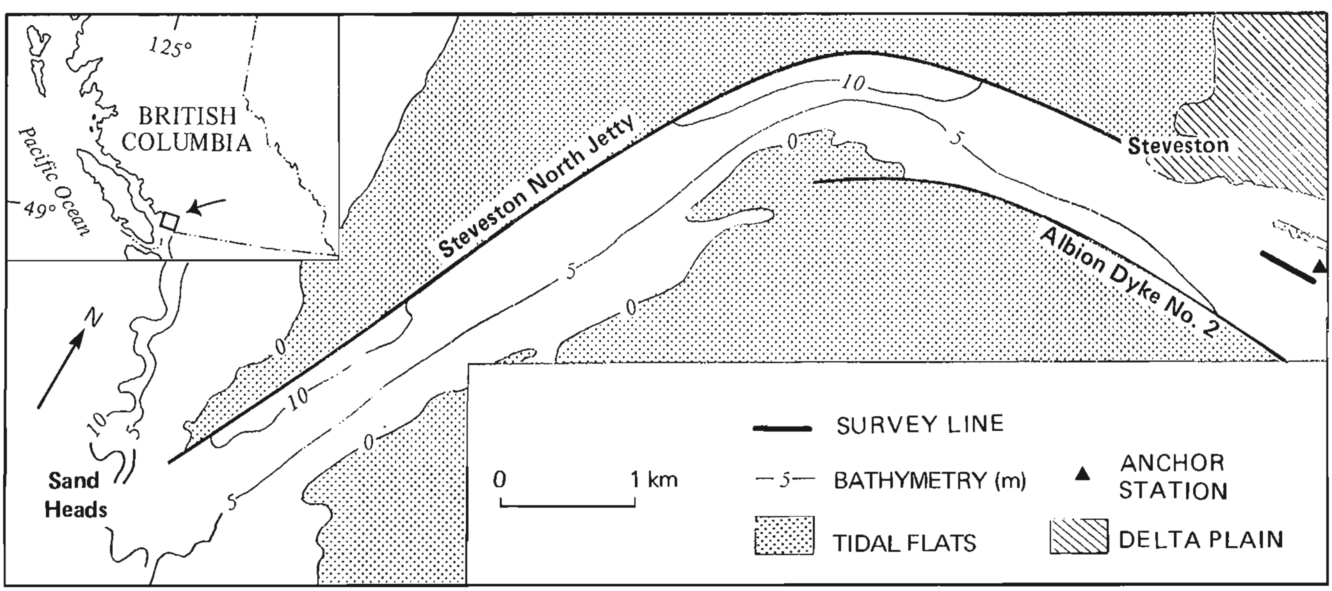

Figure 1. Lower Fraser River estuary, showing survey line and anchor station near Steveston.

METHODS

Eleven data sets were collected between May 30 and July 7, 1989. All measurements were taken aboard the survey launch CSL JAEGER near Steveston (Fig. 1) during a 3 hour period surrounding low tide (Fig. 3), the period when estuary currents are strongest and most bed-material is transported (Kostaschuk and Luternauer, 1989). Bedform geometry and migration rates were determined in two ways, using echosounding profiles obtained with a 200 kHz Apelco sounder. First, general changes in bedform characteristics were monitored along a 500 m survey line near the centre of the channel, with shore markers for positioning (Fig. 1). More precise profiles of individual bedforms were obtained by anchoring the vessel to a navigation buoy with a 200 m rope marked at 5 m intervals. This method provides extremely accurate position fixes. Soundings were obtained at the beginning and end of the survey period to estimate migration rates.

At five or six locations along the profile of each bedform, data on current speed and direction (using a Marsh McBirney 527 electromagnetic flow meter) and suspended sediment concentration (using a pump sampler and a D&A Instruments optical backscatter transmissometer) were collected at a number of levels above the bed. The current meter and transmissometer data were logged at 0.35 s intervals with a portable computer. These data were used to estimate the total suspended load at each location. At the lowest level (0.25 m), a 4 L pump sample was taken. The grain size distribution of this sample was obtained with the bottom withdrawal method (McCave, 1979) and used to separate the wash and bed-material components of the suspended load. A bed load sample was also taken at each location with a Helley-Smith sampler. A sample of the bed was obtained for each survey and the grain size distribution determined with a fall column and standard methods.

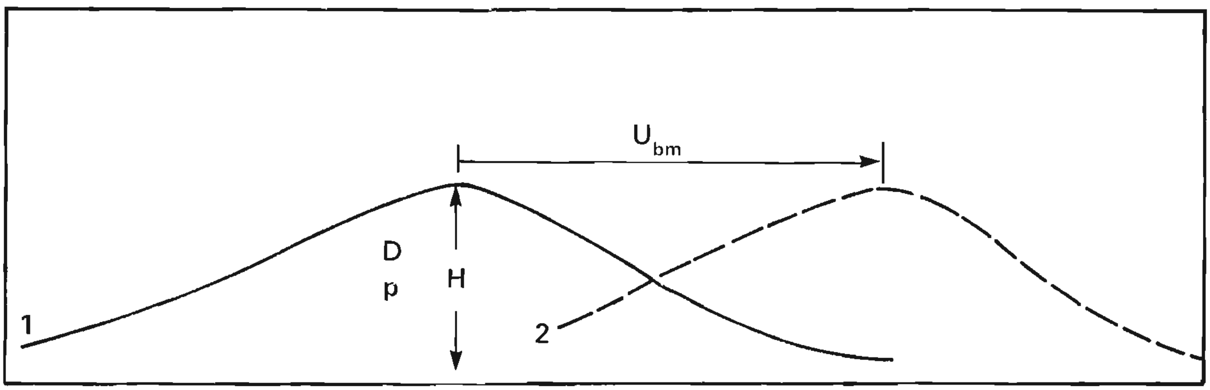

Figure 2. Definition diagram for estimating sediment transport from equation 3. U_{bm} is bedform migration rate ($m \cdot s^{-1}$), D is sediment density (2650 $kg \cdot m^{-3}$), p is porosity (0.4) and H is bedform height (m).

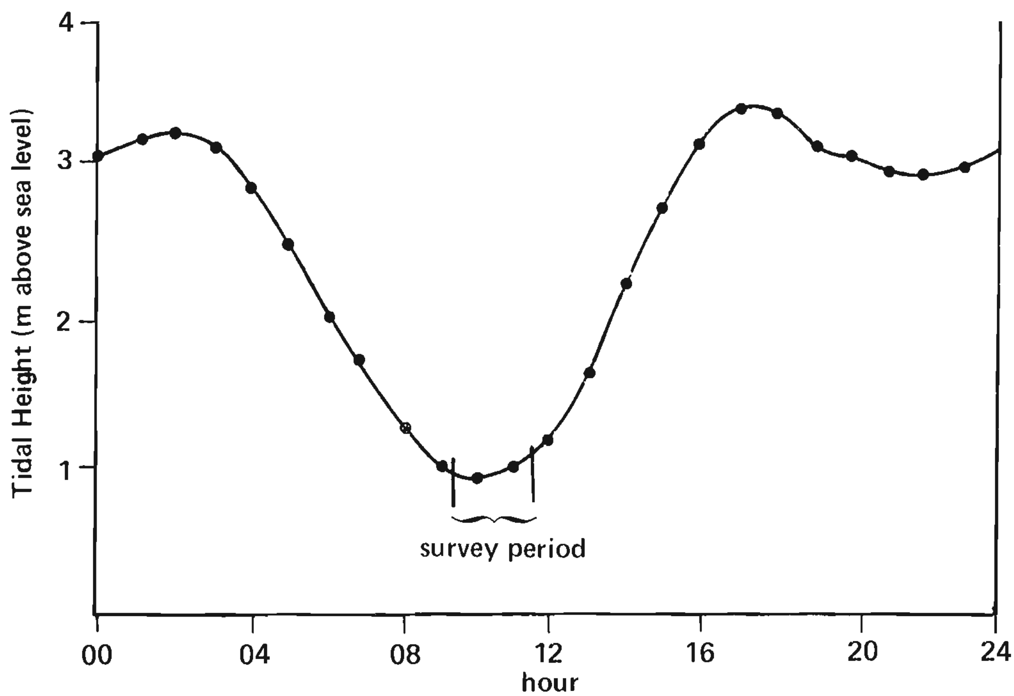

Figure 3. Tidal height predictions for 16 June, 1989 at Steveston. Predictions are from the Fraser River Mathematical Model, Institute of Ocean Sciences, Sidney, British Columbia.

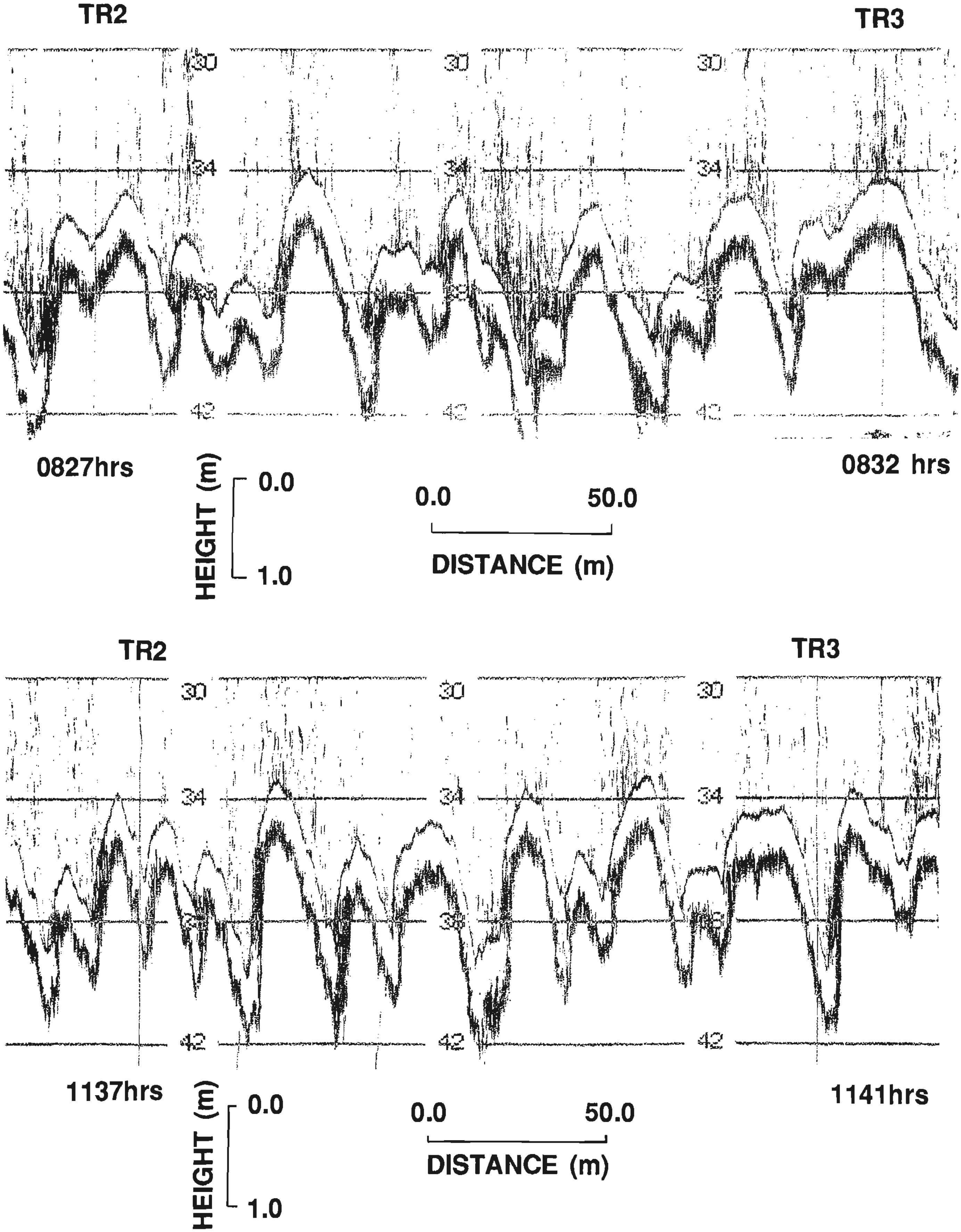

Figure 4. Echosounding profiles along the 500 m survey line (Fig. 1) for 15 June, 1989. The top profile was obtained before the survey and the lower profile after. The 'dirty' nature of the records is due to high suspended sediment concentrations. Downstream is to the left.

RESULTS

Our results can be illustrated using the case of 16 June, a survey with a moderately low tide (Fig. 3). The dunes during the entire study period were relatively symmetrical in profile with a slightly steeper downstream lee side and maximum lee side slopes of 16° (Fig. 4). The height of the dune we examined on 16 June was 0.85 m (Fig. 5) and it migrated 14.7 m in 3.5 hours. If we reasonably assume a sediment density $D = 2650$ kg $\cdot$ m^{-3}, a porosity $p = 0.4$, and a shape factor $\beta = 0.6$ (van den Berg, 1987; Kostaschuk et al., 1989b), equation 3 predicts a transport rate per unit width of channel of 0.94 kg $\cdot$ m^{-1} $\cdot$ s^{-1}.

Figure 5 shows that all currents measured over the bedforms were downstream-directed; there is no evidence of flow separation on the lee side of the dunes. Current velocity increases with distance from the bed. Total suspended sediment concentration decreases with height above the bed and there is a reduction in total suspended load and bed load on the lee side as sediment is deposited on the bedform.

In order to estimate the bed-material load transported in suspension, the concentration of wash load must be subtracted from total suspended concentration. This was accomplished by determining the concentration of sediment <0.125 mm from the grain size analysis of the 4 L samples taken 0.25 m above the bed (Fig. 6). This wash load concentration was assumed constant at all other levels above the bed at that location and concentrations of suspended bed-material determined. These concentrations were then used to determine the suspended bed-material load at each location. The mean suspended bed-material load for the six locations is 5.18 kg $\cdot$ m^{-1} $\cdot$ s^{-1} and the mean bed load is 0.04 kg $\cdot$ m^{-1} $\cdot$ s^{-1} (Fig. 5), indicating that transport of bed-material in suspension is over two orders of magnitude higher than bed load transport!

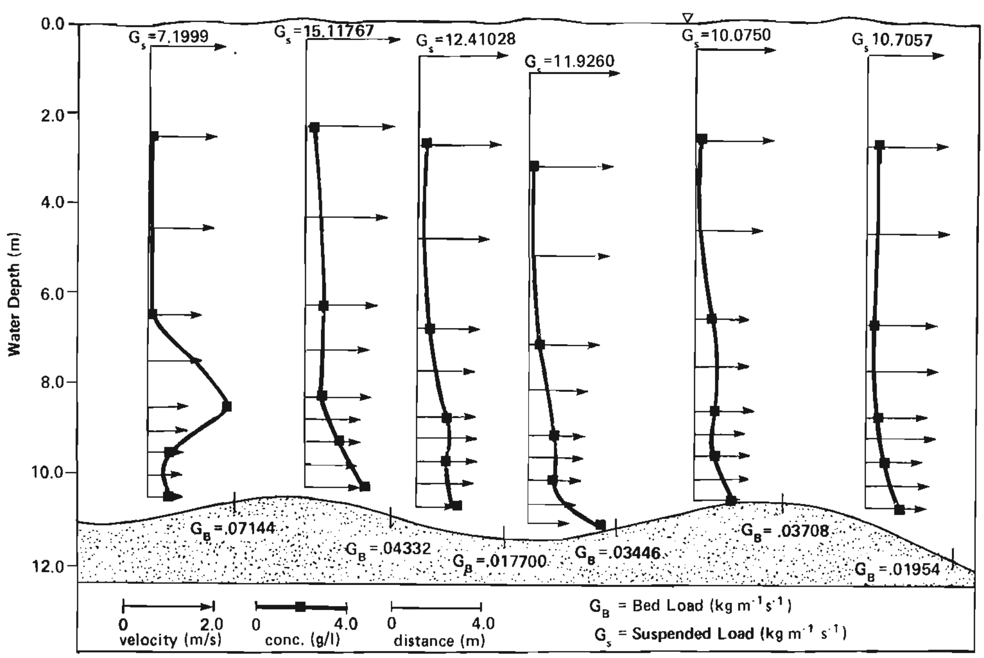

Figure 5. Velocity profiles, suspended sediment concentrations, suspended load and bed load for 16 June, 1989. Downstream is to the right. The profiles were obtained over a 3 hour period starting at the upstream end.

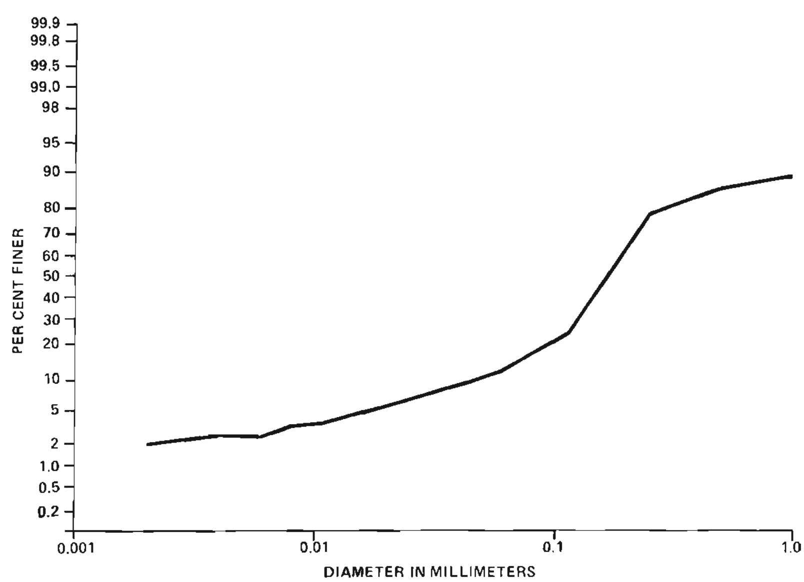

Figure 6. Grain size distribution for a suspended sediment sample obtained 0.25 m above the bed on 16 June, 1989. The sample is from the fourth profile from the left, Figure 5. The wash load/bed-material load boundary is estimated at 0.125 mm.

DISCUSSION

It is apparent that the dunes in the Fraser are symmetrical features with no separation zones on the lee side. This is contrary to standard models for dunes that describe angular features with short, steep lee sides (>30°), pronounced flow separation and migration in response to lee side avalanching of bed load immediately downstream of the crest (e.g. Allen, 1986). Similar symmetrical features to those in the Fraser, however, have been described in the Columbia River by Smith and McLean (1977) and attributed to a predominance of suspended bed-material load over bed load. They suggest that high suspended loads result in massive deposition from suspension on the lee side of the bedform, resulting in progressive removal of the separation zone, a reduction of the slope angle and a more symmetrical bedform. Our results show that suspended bed-material transport dominates bed load transport in the Fraser, indicating that Smith and McLean's explanation for symmetric bedforms probably applies bere as well. The overwhelming dominance of bed-material transport in suspension is not surprising, considering the grain size characteristics of bottom samples in the estuary. Kostaschuk et al. (1989b) found that most bottom sediment had mean grain sizes at the transition between medium and fine sand. According to Sundborg (1967), bed sediment of this size goes directly into suspension upon entrainment rather than being transported as bed load first.

The measured mean bed-material transport (5.22 kg $\cdot$ m^{-1} $\cdot$ s^{-1}) on 16 June is over five times higher than the transport predicted form the migrating bedform (0.94 kg $\cdot$ m^{-1} $\cdot$ s^{-1}). The gravimetric value of the integration constant C from equation 2, representing bed-material load not involved in bedform migration, is thus 4.28 kg $\cdot$ m^{-1} $\cdot$ s^{-1}. This analysis indicates that equation 3 severely underestimates total bed-material load and overestimates bed load. In addition, it is apparent that some suspended bed-material load is involved in bedform migration, but the bulk of it is being advected directly through the system. Kostaschuk et al. (1989b) suggested that advection could result if bed-material were transported as a uniform 'sheet' that would not be detectable in the bedform geometry or kinematics, or if the predominant transport mechanism was suspension. In addition, it may be that our estimate of the wash/suspended bed-material load break of 0.125 mm is an underestimate. However, a value of 0.2 mm still leaves significant quantities of bed-material not accounted for by bedform migration. Finally, measurement errors may contribute to these results. We expect, though, that the pump and Helley-Smith samplers would underestimate bed-material load rather than overestimate it. Predictions from equation 3 would be subject to positioning errors affecting values of bedform height and migration rates, plus errors from assumed values of sediment density and porosity. It is unlikely, however, that these would produce the large differences apparent between the measured values and those estimated from equation 3.

CONCLUSIONS

1. A dominance of bed-material transport in suspension in the Fraser estuary results in symmetrical bedforms without flow separation on the lee side.
2. The bedform migration equation, equation 3, overestimates bed load transport and underestimates suspended bed-material transport.
3. Sediment transport rates based on equation 3 are only 20% of values from direct measurements of bed-material load in the water column. This indicates that most of the bed-material load is not involved in bedform movement but is being advected through the system in suspension.
4. Future research should focus on the exchange between the bed and water column over entire tidal cycles, to better define bedform migration and sediment transport relationships.

ACKNOWLEDGMENTS

We thank Ray Sanderson for piloting the **Jaeger** and Professor M.A. Church and the Department of Geography, University of British Columbia, for technical assistance and the use of laboratory facilities. Financial support was provided by a Natural Sciences and Engineering Research Council Operating Grant to Kostaschuk and Luternauer's Geological Survey of Canada Project 860022.

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Holocene sea level change and crustal deformation, southwestern British Columbia

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Abstract

The pattern and timing of Holocene sea level change in southwestern British Columbia are being studied, in part, to elucidate recent crustal movements in the region. Studies of Holocene sediments at coastal sites near Victoria and Vancouver indicate that the sea has risen up to a few metres relative to the land during the last half of the Holocene. Sea level, however, has not fluctuated more than 1 m from its present position during the last 2000 years, and there is no evidence during this period for regional coseismic subsidence or uplift which might be expected during great subduction earthquakes on the Cascadia subduction zone.

Résumé

Des études sont actuellement en cours portant sur la configuration et l'âge du changement des niveaux de la mer tels qu'ils se présentaient pendant l'Holocène dans le sud-ouest de la Colombie-Britannique, en partie pour élucider les mouvements récents de la croûte dans la région. Des études menées sur les sédiments holocènes dans des emplacements côtiers près de Victoria et de Vancouver, indiquent que la mer s'est élevée de quelques mètres par rapport à la terre au cours de la deuxième moitié de l'Holocène. Cependant, le niveau de la mer n'a pas fluctué de plus d'un mètre de sa position actuelle au cours des 2000 dernières années; en outre, il n'existe aucun indice, au cours de cette période de subsidence ou de soulèvement cosismique régional tel qu'on pourrait s'y attendre au cours de grands séismes par subduction qui auraient touché la zone de subduction de Cascadia.

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INTRODUCTION

Western British Columbia is the most seismically active area in Canada. Large earthquakes occur at the boundary of the America lithospheric plate west of the continental margin. In addition, the southern coastal region is affected by moderately large (M6-7+) intraplate earthquakes, the most recent of which was a M7.2 event in 1946, centred northwest of Comox on central Vancouver Island (Hodgson, 1946; Rogers and Hasegawa, 1978).

Recently, concern has been expressed that major urban centres on the southern British Columbia coast could experience an unprecedented "great" earthquake, comparable in size to the 1960 Chilean and 1964 Alaskan quakes, which were responsible for widespread damage and considerable loss of life (Rogers, 1988). Such an earthquake might occur along the interface between the America plate and the subducting Juan de Fuca plate west of southern Vancouver Island, Washington, or northern Oregon ("Cascadia subduction zone"). One might also occur west of central Vancouver Island where the Explorer plate apparently is underriding the America plate. Stratigraphic and geochronological evidence from estuaries on the west coast of Washington and Oregon suggests that such earthquakes may have occurred on average once every several hundred years during the late Holocene, the most recent about 300 years ago (Atwater, 1987; Peterson and Darienzo, 1988).

In response to public concern over the possibility of such potentially devastating earthquakes, the Geological Survey of Canada has initiated a program of geological studies aimed at determining the character and extent of Quaternary crustal movements in western British Columbia and at relating these to the historical and contemporary stress field (Clague, 1989). These studies may provide information on the frequency and magnitude of large earthquakes in this region during the Holocene, which in turn can be used to assess the likelihood of similar earthquakes in the future.

A major component of the GSC's west coast paleoseismicity program (Clague, 1989) involves the reconstruction of Holocene sea level change. The sea surface provides a datum for determining vertical crustal movements, although other factors such as mass and phase changes in the mantle, variations in tidal amplitude, and steric effects can alter land-sea positions over time. Regional uplift and subsidence during large earthquakes cause sudden shifts in the level of the sea relative to the land. These shifts commonly are recorded by raised beaches and related littoral landforms in uplifted areas and by submerged marshes and soils in areas of subsidence.

The pattern and chronology of relative sea level movements are being documented at several places on the British Columbia coast in an attempt to determine what factors have controlled these movements and whether or not large earthquakes have occurred in the recent past. This is done through airphoto interpretation, stratigraphic and sedimentological logging of natural and artificial exposures and cores, paleoecological analysis, and radiocarbon dating of organic material. This report summarizes the current status of these investigations and presents some preliminary results (see also Clague, 1989).

STUDY AREA

To date, the study has focused on the coasts of southeastern Vancouver Island and mainland British Columbia near Vancouver (Fig. 1). The main reasons for choosing this area are that it includes the largest population centres in the province and that it would be subjected to strong and long shaking during an earthquake on the Cascadia subduction zone. By analogy with similar subduction zones in other regions, parts of the study area might be expected to subside one or more metres during a great earthquake (Dragert and Rogers, 1988). Subsidence of this magnitude almost certainly would leave a stratigraphic record at favourable coastal sites. Considerable damage would also result from a moderately large intraplate earthquake, several of which have occurred during this century immediately north and south of the study area. It is not clear, however, if coseismic uplift and subsidence associated with such quakes would be sufficiently large to leave a geological record.

STATUS OF INVESTIGATIONS

Several sites on southern Vancouver Island, three on the British Columbia mainland south of Vancouver, and one on Galiano Island were investigated during the summers of 1988 and 1989 (Fig. 1, Table 1). These sites were chosen

Table 1. Sites and types of data.

Locality	Location	Types of data[a] Pollen	Diatoms	Foraminifera	14C
Montague Harbour	48°54.0′N, 123°24.4′W				X
Tsehum Harbour	48°40.5′N, 123°25.3′W		X	X	X
Island View Beach	48°34.7′N, 123°22.4′W	X	X	X	X
Telegraph Cove	48°27.8′N, 123°16.7′W				
Gyro Park	48°27.6′N, 123°17.5′W	X	X	X	X
Portage Inlet					
Helmcken Park	48°27.6′N, 123°25.7′W	X	X	X	X
Highway 1 rest area	48°27.8′N, 123°25.4′W	X	X	X	X
Colquitz River Park	48°27.4′N, 123 23.9′W	X		X	X
Witty's Lagoon	48°23.1′N, 123°31.1′W	X	X	X	X
Sooke River	48°23.3′N, 123°42.3′W				
Muir Creek	48°22.9′N, 123 51.8′W	X	X	X	X
Burns Bog	49°07.6′N, 123°01.1′W	X	X	X	X
Serpentine River					
Colebrook Road	49°05.9′N, 122°49.9′W	X		X	X
152nd Street	49°05.5′N, 122°48.1′W	X	X	X	X

[a] Analyses are in progress.

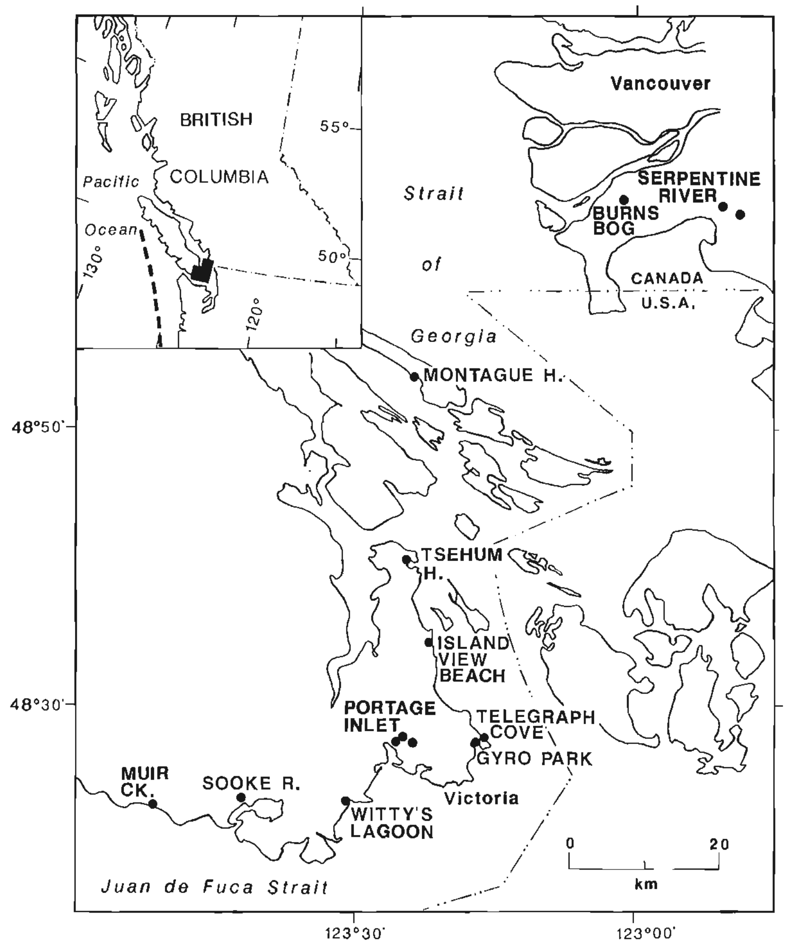

Figure 1. Location map showing sites investigated during 1988 and 1989. The dashed line on the inset indicates the approximate boundary between the America and Juan de Fuca plates (Cascadia subduction zone).

because they are wetlands at or just above the limit of tides and because they are underlain by Holocene sediments. Cores were obtained at some sites, and backhoe trenches and pits were dug at others. The stratigraphy of each core and exposure was logged in detail, with information recorded on texture, colour, sedimentary structures, fossil content, and contact relationships. Samples were collected for radiocarbon dating and for grain size, pollen, diatom, and foraminiferal analyses. Stratigraphic and sedimentological work is now complete at all sites shown in Figure 1. Some geochronological and paleoecological data are available for several of the sites, but most of the radiocarbon, pollen, diatom, and foraminiferal analyses have not yet been completed.

PRELIMINARY RESULTS

Most of the Vancouver Island sites, as well as the site on Galiano Island show evidence of a late Holocene, relative rise in sea level. Terrestrial and freshwater peats occur below the upper limit of tides at Portage Inlet, Gyro Park, Telegraph Cove, Island View Beach, Muir Creek, Witty's Lagoon, and Montague Harbour.

The peat at Portage Inlet, which is covered by marine mud, is up to 3 m thick and extends down to 7 m below mean sea level. It has yielded radiocarbon ages ranging from 9250 ± 140 BP to 5470 ± 115 BP (I-3676 and I-3673; Foster, 1972).

Peat at Gyro Park is up to 6 m thick and extends from high tide level seaward beneath thin littoral sand and gravel. The base of this peat at a site approximately 150 m inland from the present shoreline has been radiocarbon dated at 4120 ± 70 BP (GSC-4869, uncorrected age). The peat lacks interbeds of detrital mineral sediments that might be expected during a transgression, thus it apparently accumulated as rapidly as, or more rapidly than, any relative rise in sea level. Pollen analysis, however, indicates that the upper part of the peat was deposited in brackish water (i.e., upper intertidal zone), whereas the lower part accumulated dominantly in a freshwater swamp or bog. The peat may thus record a rise in sea level similar to that found at Portage Inlet and other sites mentioned below.

Terrestrial peats at Island View Beach (Clague, 1989, his Fig. 3), Telegraph Cove, and Muir Creek are a few tens of centimetres thick and are overlain by intertidal sediments. Fossil stumps are rooted in the upper parts of the peats at Island View Beach and Muir Creek. At Island View Beach, a stump, protruding through modern beach gravel about 1-1.5 m below high tide, has been dated at 2040 ± 130 BP (GSC-252; Mathews et al., 1970). This, in conjunction with other radiocarbon dates on associated peat (Clague, 1989), shows that a minor transgression occurred at this locality between about 2500 and 2000 BP. At Muir Creek, peat with in situ roots and stumps has been dated at 3530 ± 60 BP (GSC-4758). A radiocarbon age of 3120 ± 70 BP (GSC-4820) on charcoal in estuarine sediments overlying this peat shows that the forest floor was transgressed by the sea shortly after 3500 BP.

A midden at Montague Harbour (archeological site DfRu-22), on Galiano Island, has been partially transgressed and eroded during the late Holocene. Cultural material overlies a peat which extends seaward from the backshore area (Fig. 2). Both the peat and the seaward edge of the midden are overlain by modern beach gravel.

In summary, there is evidence for one or more transgressions on southeastern Vancouver Island during the late Holocene. At Muir Creek, there are several alternations of organic and mineral sediment that may record complex fluctuations in sea level, but there is evidence for only one transgression at all other localities. This transgression ended about 2000 BP, and since then, there has been little or no change in the level of the sea relative to the land on southeastern Vancouver Island.

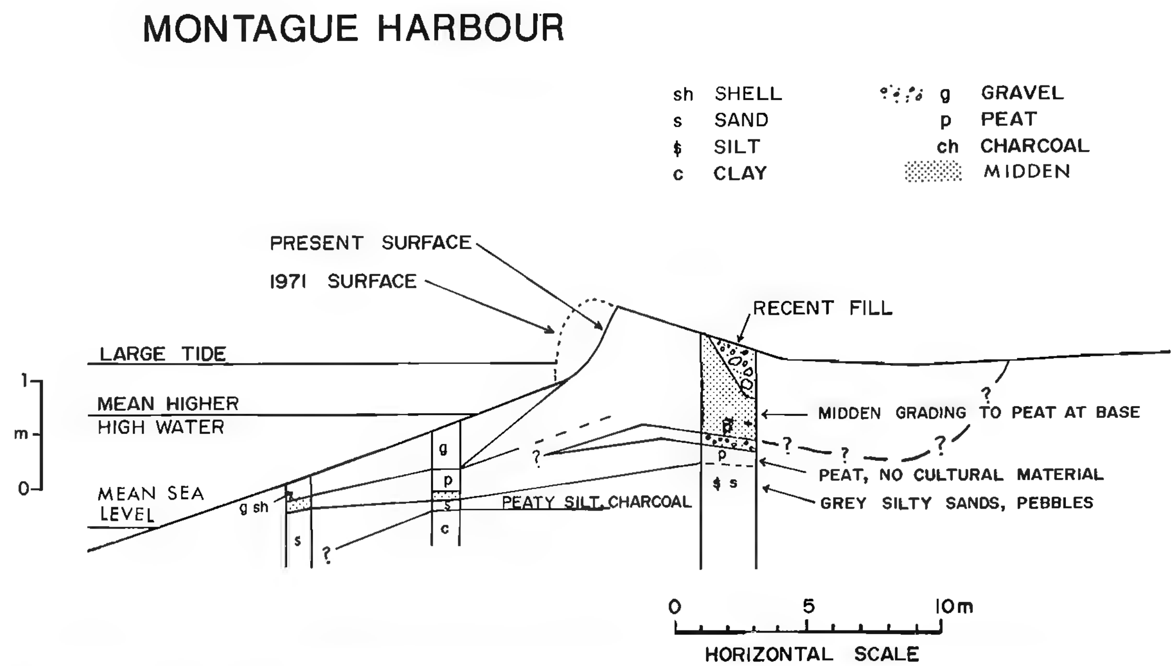

Figure 2. Stratigraphic sections at archeological site DfRu-22, Montague Harbour. Peat underlies modern beach gravel and cultural material and was found to at least 1.5 m below the upper limit of tides.

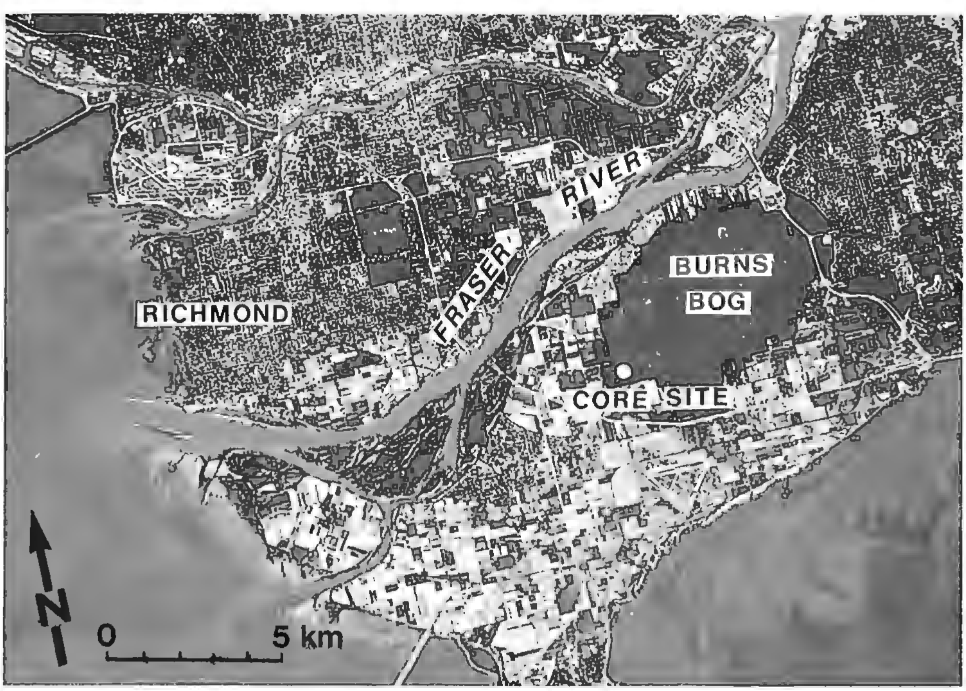

Figure 3. Satellite image of the Fraser delta and surrounding region, showing the location of the core site in Burns Bog. The image was taken on September 21, 1987.

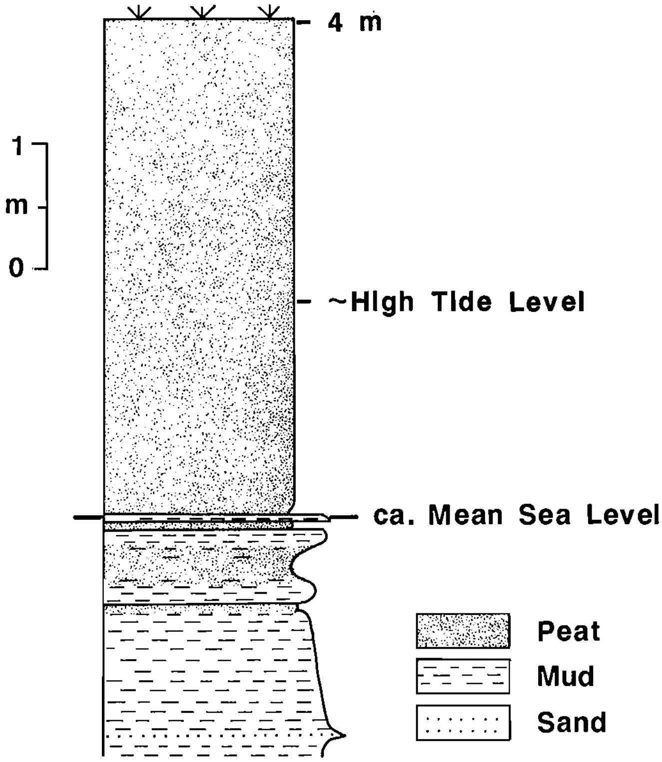

Figure 4. Stratigraphy of Burns Bog core (see Fig. 2 for location). Peat from the surface to 4 m depth accumulated above high tide level. The interfingering of peat and mud below 4 m depth records gradual emergence and colonization of the delta surface by vegetation. The base of the thick peat directly above the highest mud bed is morethan 4000 radiocarbon years old (a date of 4125 ± 110 BP (I-7627; Hebda, 1977) was obtained on peat at this level in a core 1 km to the south-southeast).

Cores taken from bogs and fens at three low-elevation sites near Vancouver record continuous deposition of peat over the last several thousand years. As at Gyro Park, variations in peat type (e.g., freshwater vs. brackish) may record minor variations in sea level. Interbeds of mineral sediment, however, are not present in the upper parts of the cores, thus it appears that these sites have not subsided significantly during the late Holocene. This is demonstrated by a core taken from western Burns Bog on the Fraser River delta (Fig. 3, 4). Four metres of terrestrial and freshwater peat overlie interbedded muddy peat and organic-rich mud (Fig. 4). These sediments form the top of an extensive delta plain that is graded to present sea level. Peat extends from about 0 to 4 m above mean sea level (-2 to $+2$ m above high tide) and began to accumulate before 4000 BP (Hebda, 1977). At no time during this period has the sea transgressed the bog; thus the delta plain on which the bog rests could not have subsided more than 2 m or the sea have risen more than 2 m above its present level.

DISCUSSION

Our evidence shows that the sea has risen relative to the land in coastal southwestern British Columbia during the late Holocene. This transgression culminated about 2000 BP when the sea reached its present level with respect to the land. Since then, shorelines in this region probably have been very near their present positions (± 0.5 m vertically).

These conclusions are in general agreement with previous findings (Mathews et al., 1970; Clague et al., 1982; Clague, 1989). The relative stability of sea level over the last 2000 years, however, seems surprising, given the intense crustal deformation in the region and its location near a plate margin. Sea level change, of course, is the integrated response of many factors, and it is possible that tectonic movements are being masked by eustatic changes or by residual glacioisostatic movements related to the final disappearance of the Cordilleran Ice Sheet 10 000-11 000 BP. Residual glacioisostatic movements during the last 2000 years, however, probably have been very small (Mathews et al., 1970), and any significant eustatic rise during this period would require compensatory crustal uplift to produce no net sea level change along the southern British Columbia coast.

The pattern of late Holocene sea level change in southwestern British Columbia differs markedly from that in Washington and Oregon to the south. In estuaries along the Pacific coast of Washington and northern Oregon, peat beds alternate with intertidal mud units in a cyclic fashion. The contacts between peat beds and overlying mud units are very sharp, indicating that terrestrial surfaces were rapidly transgressed by the sea. This stratigraphy has been attributed to episodic coseismic subsidence associated with great earthquakes on the Cascadia subduction zone (Atwater, 1987). Radiocarbon dating of the peat beds in these estuaries indicates transgressions, and possibly great earthquakes, about 300, 1000, 1500, 1700, 2500, 2800, and 3500 BP. Multiple peat beds have not been found in the study area, although there are several organic-rich detrital layers in the estuarine sequence at the mouth of Muir Creek, the most westerly of the sites examined to date. Even here, however, there is little or no evidence for significant subsidence during the last 2000 years, at least within the limit of resolution of the data (± 0.5 m). This does not deny the possibility that great earthquakes have occurred during the late Holocene on the Cascadia subduction zone. It does suggest, however, that these earthquakes have not produced significant regional subsidence in southwestern British Columbia comparable to that observed in historical subduction earthquakes (e.g., Chile, 1960).

The dissimilarity between the late Holocene sea level records of the northwestern United States and British Columbia is perhaps not surprising, given the fact that the tectonic regimes of the two areas are different. Although both areas are situated "inboard" of the Cascadia subduction zone, there are differences in the geological structure of the America plate between the two. Also, the interface between the America and Juan de Fuca plates has a different orientation north of Juan de Fuca Strait than to the south. Finally, crustal deformation on southern Vancouver Island is probably influenced by complex, poorly understood interactions of the Juan de Fuca and America plates with the Explorer plate, whereas deformation along the Washington and Oregon coasts is not affected by the Explorer plate. Given these differences, one would not expect similar patterns of deformation in the two regions.

ACKNOWLEDGMENTS

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Holocene sediments from Saanich Inlet, British Columbia, and their neotectonic implications

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Abstract

Piston cores collected from Saanich Inlet, British Columbia, in January 1989, contain a record of episodic, Holocene, sediment gravity flows that possibly have been triggered by large earth quakes. Beds of massive silty clay, emplaced by grain flows or debris flows, are interstratified with rhythmically laminated sediments consisting of annual couplets of diatoms and mud. Flows occur, on average, about once every 100 years. This may give some indication of the recurrence interval of large earthquakes in this region.

Résumé

Des carottes prises par carottier à piston dans l'inlet Saanich, en Colombie-Britannique, en janvier 1989, attestent qu'il s'est produit des coulées par gravité de sédiments épisodiques, holocènes, qui ont probablement été déclenchés par de grands séismes. Des couches d'argile silteuse massive, mis en place par des coulées de matériaux granulaires ou des coulées boueuses, sont interstratifiés avec des sédiments stratifiés d'une façon rythmique qui sont constitués de doublets annuels de diatomées et de boue. Les écoulements se font en moyenne environ une fois tous les 100 ans. Ils peuvent donner une certaine indication de l'intervalle de récurrence des grands séismes dans cette région.

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INTRODUCTION

In January 1989, the Geological Survey of Canada collected a series of sediment cores from Saanich Inlet, British Columbia. The objective of this exercise was to obtain a record of large Holocene earthquakes in the region. Saanich Inlet sediments are varved (Buddemeier, 1969), thus the potential exists for precisely dating disturbed or resedimented layers produced during earthquakes. This research endeavour may improve our ability to forecast large intraplate and thrust earthquakes that might affect the densely populated southwestern corner of British Columbia (Rogers, 1988).

During the last three decades, a considerable amount of general research has been done in Saanich Inlet (Juniper and Brinkhurst, 1984). Much of this work is concerned with the sediments that underlie the floor of the inlet (e.g., Gucluer, 1962; Gross et al., 1963; Gucluer and Gross, 1964; Buddemeier, 1969), but to our knowledge none of the work has addressed our objective.

Cores from Saanich Inlet obtained by the University of Washington in the 1960's were observed to comprise alternating dark and light laminae (Gross et al., 1963; Buddemeier, 1969). Detailed examination of these sediments indicates significant differences in the composition of the two types of laminae. The light laminae consist mainly of diatoms deposited during spring and summer blooms (Gross et al., 1963; Sancetta and Calvert, 1988). The dark laminae are dominantly fine mineral detritus derived from rivers outside the inlet (mainly Cowichan River) and deposited from suspension during fall and winter. Numerous radiocarbon age determinations on organic carbon at various depths below the seafloor in Saanich Inlet confirm that the rhythmites are varves (Buddemeier, 1969; Yang, 1971).

The varved sequences are interbedded with massive beds ranging from a few centimetres to several tens of centimetres thick. Previous workers have ascribed these beds to deformation resulting from the coring process (Buddemeier, 1969; Powys, 1987). At least two other explanations, however, seem more plausible: (1) the massive beds were originally laminated sediments, but have been liquefied during earthquakes or (2) the beds are products of sediment gravity flows. During an earthquake, water-saturated sediments directly underlying the seafloor in Saanich Inlet might liquefy, producing a massive bed above undisturbed varved sediments. According to this scenario, each massive bed would record one earthquake. If, on the other hand, the massive beds were emplaced by sediment gravity flows, a direct link with earthquakes cannot be proven. Earthquakes, however, would be the most likely triggers of sediment gravity flows in an inlet such as this with no significant inputs of fluvial sediment.

STUDY AREA

Saanich Inlet is a fiord located at the southern end of Vancouver Island directly northwest of Victoria (Fig. 1). The inlet is 26 km long, up to 8 km wide, has an average depth of 120 m, and a maximum depth of 236 m. Goldstream Creek, at its south end, is the only local source of sediment and freshwater of any significance. It contributes only a small percentage of the 9 × 10^4 tonnes of sediment that accumulates annually in Saanich Inlet (Gross et al., 1963). The main source of terrigenous sediment is Cowichan River which empties into Satellite Channel 12 km northwest of the mouth of the inlet.

A bedrock sill at the north end of Saanich Inlet rises to within 70 m of the surface and restricts normal water circulation. The lower part of the water column is anoxic. The boundary between oxygenated and anoxic waters ranges from 70 m below sea level in October to 150 m below sea level in December (Gross et al., 1963). There is an absence of epifauna and infauna in this anoxic environment, which accounts for the excellent preservation of stratification. Minor sediment mixing, however, does occur due to the release of hydrogen sulphide gas.

METHODS

Continuous cores were obtained at three locations in Saanich Inlet using a large (10 cm diameter) piston corer: **core 1** (length = 885 cm) — 48°33.73"N, 123°30.45"w, water depth = 227 m; **core 2** (1178 cm) — 48°38.41"N, 123°30.17"W, 198 m; **core 3** (1180 cm) — 48°36.10", 123°30.00"W, 227 m (Fig. 1). Upon recovery, the three cores were cut into manageable lengths of approximately 1.5 m and stored in a cold room, first on the ship and later at Pacific Geoscience Centre.

All core material was first X-rayed at Industrial Nondestructive Testing Ltd. in Vancouver using a General Electric Maxi-Mar X-ray machine which houses a 250 kvp - 15 ma - 100 % duty cycle X-ray tube. The cores were then returned to Pacific Geoscience Centre and split longitudinally using an electric hand saw and steel wire. This produced working and archival core halves which were stored in sealed plastic D-tubes. Sediment in the working core halves was subsequently logged by documenting the thickness, character, colour, and fossil content of the sediments, internal structures, and contact relationships between varved and massive units.

Colour photographs of the archival core halves were taken with a camera mounted on a quadraped. The developed prints were assembled into sequential strips. Colour slides and black and white photographs were taken of all contacts between massive beds and varved sequences.

Fifty-nine sediment samples were removed from the working core halves for grain size analysis. An attempt was made to sample sediment at a range of depths in all three cores. Closely spaced samples were also collected through individual massive beds into overlying and underlying varved units. This sampling methodology allowed the textural characteristics of each massive bed to be compared to those of the enclosing varved sediments. Grain size analyses were performed at Pacific Geoscience Centre using a Micromeritic SediGraph 5100.

Several sediment samples, primarily near the base of each core, were removed for radiocarbon dating and sent

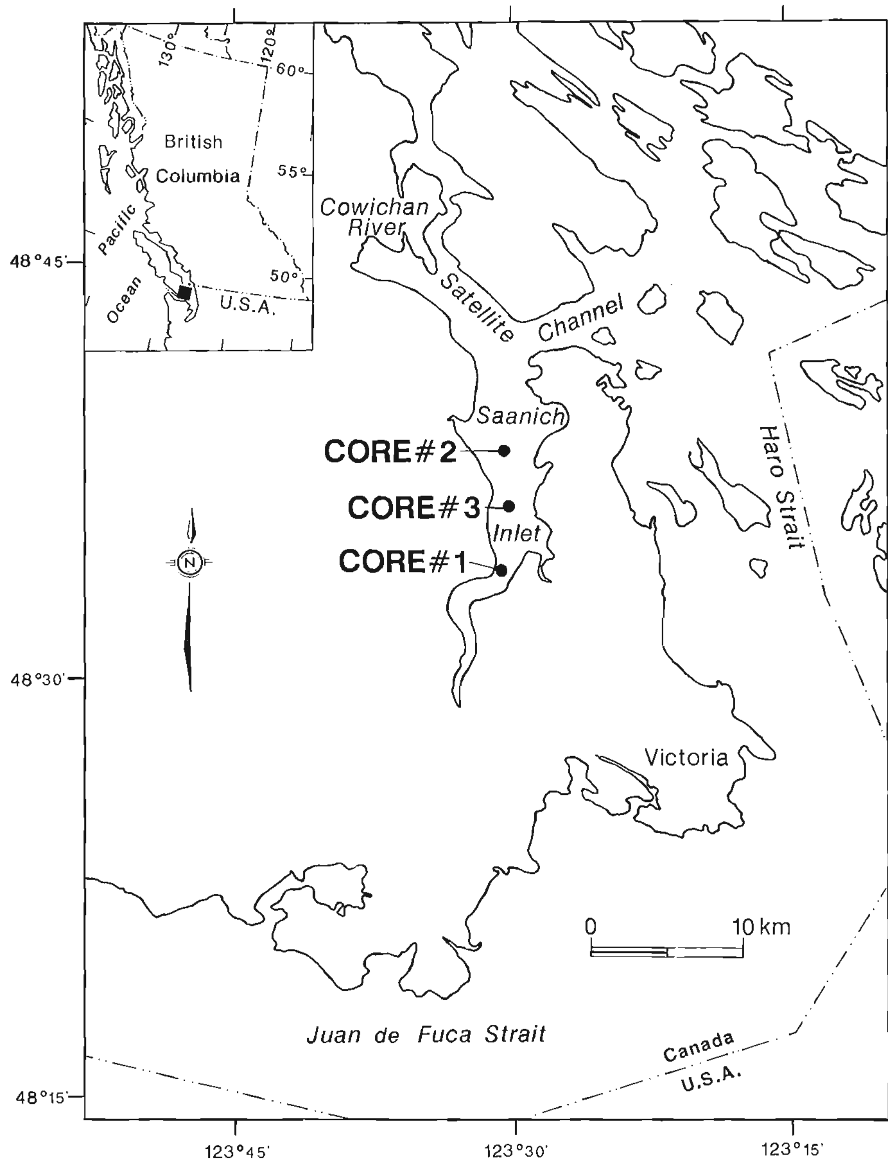

Figure 1. Location map showing Saanich Inlet and the core sites.

to the Geological Survey of Canada Radiocarbon Laboratory in Ottawa. Additional samples were taken at 2 cm intervals through the uppermost 40 cm of core 2 and submitted to TRIUMF at the University of British Columbia for ^{137}Cs analysis. This was done to identify the first occurrence and maximum concentration of ^{137}Cs in the core. The former would correspond to the first appearance of ^{137}Cs in the atmosphere in 1954, and the latter to the 1963 peak identified in many other sediments (Ritchie et al., 1975; Ashley and Moritz, 1979). Either would provide an absolute datum for back-dating the varve chronology.

RESULTS

Analyses have been completed for all three cores, but we present results mainly for core 3 in this preliminary report. Gaps in this core are restricted to the uppermost few metres and correspond to the tops of four of the cut sections. As the cores were stored upright, gradual dewatering resulted in minor compaction and hence the formation of empty spaces at the tops of sections.

Twelve massive beds, ranging from 2.0 to 61.5 cm thick, were identified in core 3 (Fig. 2, Table 1). The upper contacts of all massive beds are sharp, whereas some basal contacts are sharp and others are gradational. Core 3 contains 1394 (±5 %) couplets, averaging 4 mm in thickness. The number of couplets between successive massive beds ranges from 14 to 328 (Table 1). All contacts between layers within varved units are conformable and planar to slightly curvilinear (Fig. 3).

Light-coloured laminae in the varved sequences are olive (5Y4/3) to olive brown (2.5Y4/4; note: all Munsell colour determinations were made on wet sediment). Dark laminae are dark olive grey (5Y3/2) to very dark greyish brown (2.5Y3/2), and massive beds are dark olive grey (5Y3/2). Several of the massive beds are capped by a thicker-than-normal, light-coloured lamina (see Fig. 3 for an example and Fig. 2 for the distribution of these layers). These capping light layers have sharp upper contacts and gradational lower contacts.

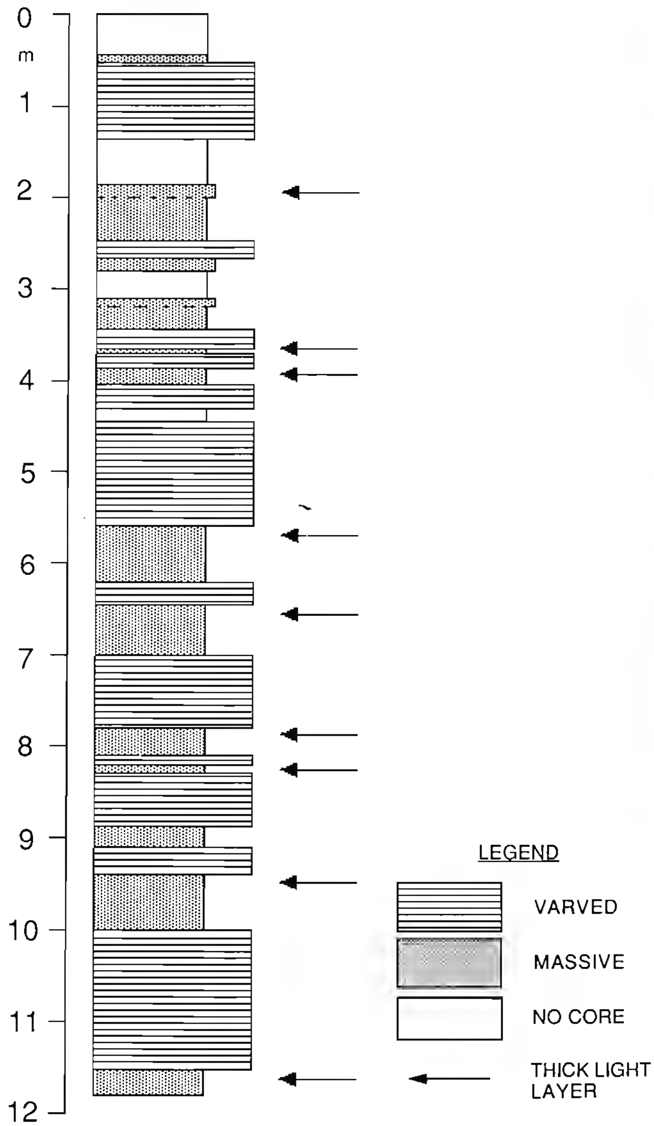

Figure 2. Stratigraphy of core 3.

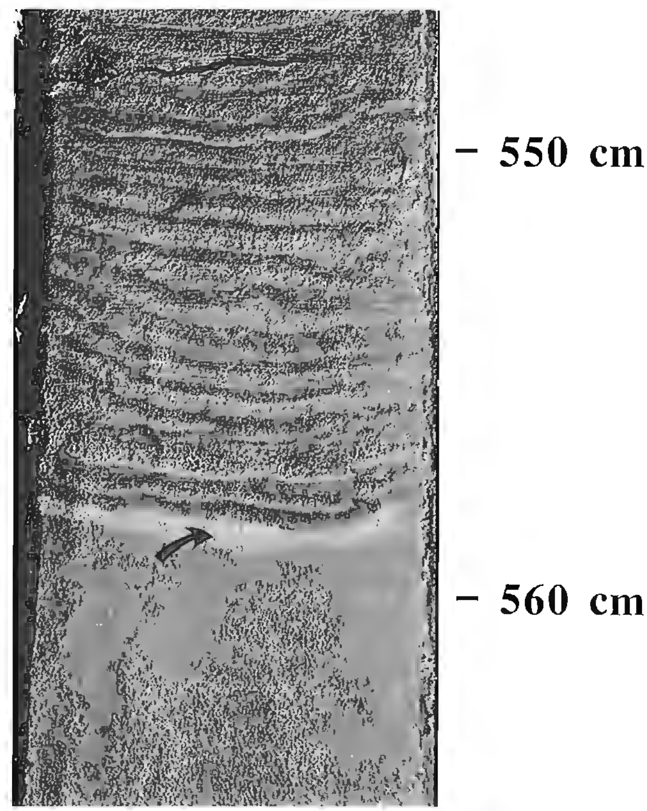

Figure 3. Massive bed overlain by rhythmically laminated sediments. Scale in centimetres. Note anomalously thick, light-coloured layer at the top of the massive bed (arrow).

Table 1. Stratigraphic data for core 3, Saanich Inlet.

Massive bed	Depth[1] (cm)	Thickness (cm)	No. varves above bed[2]	Cumulative no. varves above bed[2]
1	46.0-52.5	6.5	0	0
2	200.0-248.0	48.0	235	235
3	319.5-344.5	25.0	109	344
4	367.0-369.0	2.0	70	414
5	386.0-404.5	18.5	42	456
6	560.0-621.5	61.5	256	712
7	646.5-699.5	53.0	43	755
8	779.0-812.0	33.0	133	888
9	819.0-828.5	9.5	14	902
10	887.5-910.0	22.5	106	1008
11	939.5-999.5	60.0	58	1066
12	1152.0-1180.5	28.5	328	1394

[1]Depth below top of plastic core liner. No adjustment for compaction and dewatering.
[2]Accuracy = ±5%.

The X-radiographs proved to be of limited value. Although the massive beds and stratification within the varved sequences were recognized in the radiographs, fine structures could not be easily discerned. This problem is probably due to the large diameter of the core and the absence of a protective shield during X-raying.

Particle size distributions were determined for 17 samples from core 3 (Fig. 4). The analyses indicate that the sediments are silty clay with less than 1 % sand. Because clay size sediment is dominant in all samples, summary statistics are presented for this fraction. The clay content of the varved sediments ranges from 72.8 % to 78.6 % (average = 74.6 ± 2.0 %). The massive sediments are coarser, with clay contents of 65.0 % to 72.0 % (average = 67.2 ± 2.0 %). Results from a typical massive bed and bounding rhythmites are shown in Figure 5.

No ^{137}Cs was found in the uppermost sediments of core 2. This suggests that at least the last 45 years of sediments were not recovered during coring at this site. This is perhaps not surprising, given the low strength and very high water content of sediments directly underlying the seafloor in Saanich Inlet (Powys, 1987).

DISCUSSION

Sedimentological research and radiocarbon dating over the last three decades have shown that the rhythmic couplets in Saanich Inlet are varves.

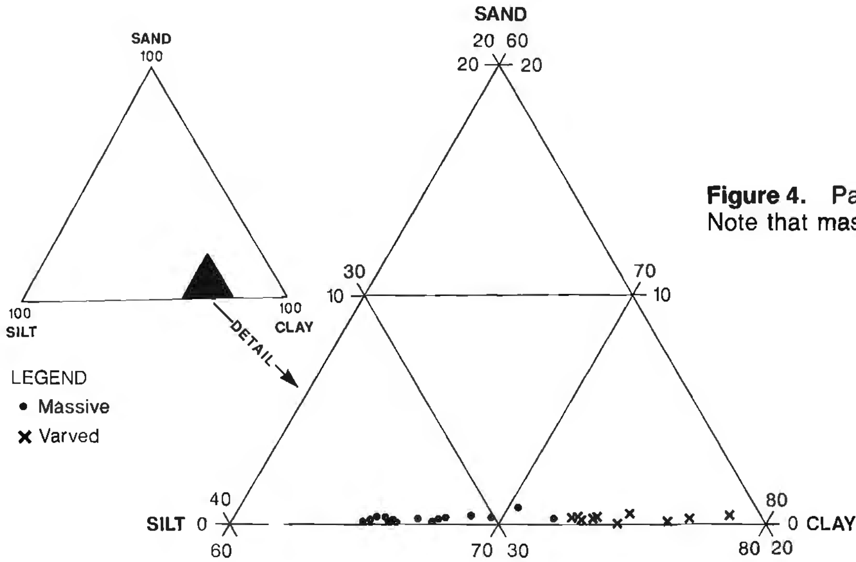

Figure 4. Particle size distribution of samples from core 3. Note that massive beds are coarser than the varves.

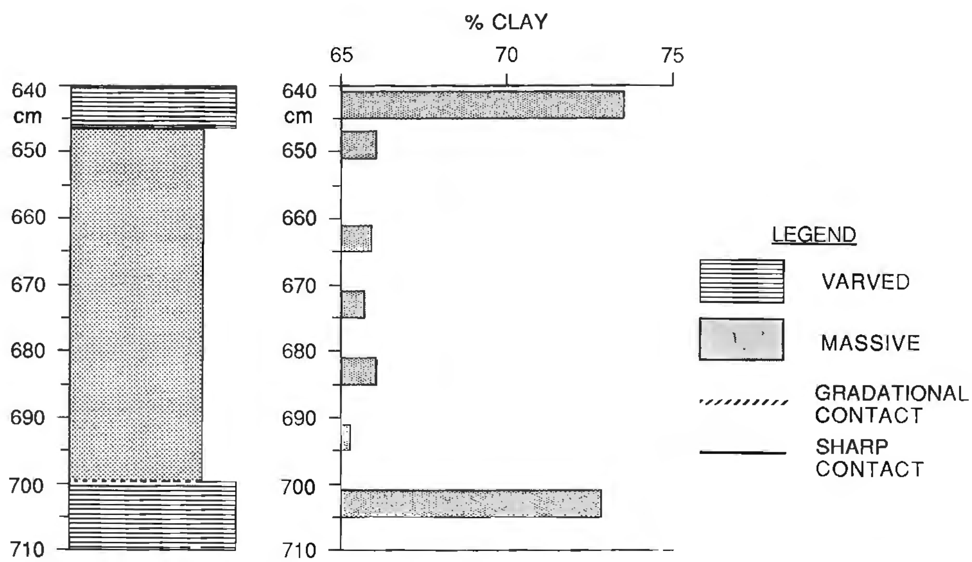

Figure 5. Clay percentage in samples from a typical massive bed and bounding varves.

The massive beds intercalated with the varves are more problematic. It is unlikely that these beds formed as a result of disturbance during coring and storage because they are found to considerable depths below the seafloor. Although the upper, water-saturated sediments of Saanich Inlet might be disturbed during coring, the lower, compact and essentially dewatered sediments would not behave in a fluid manner required for homogenization. Furthermore, the sharp upper contacts of massive beds and the presence of capping, anomalously thick, light-coloured layers argue against disturbance during coring.

More plausible explanations for the massive beds include sudden changes in the style and rate of sedimentation, liquefaction, and resedimentation. The first explanation is unlikely because there probably have not been sudden and repeated changes in the supply of sediment to Saanich Inlet during the Holocene. Laminated sediments which have undergone in situ liquefaction due to earthquake shaking commonly show Kuenen-type structures, including low amplitude folds and load features (Sims, 1973, 1975). Kuenen-type structures were not observed in any of the cores obtained in this study. If the massive beds were originally laminated and have been liquefied by shaking, it follows that the particle size characteristics of these beds would be comparable to those of the enclosing varved sediments. Data presented in this report indicate that this is not the case. The coarser nature of the massive beds thus supports an extra-local, or allogenic, sediment source and not in situ liquefaction.

Resedimentation best explains the massive beds. The beds do not, however, possess any structures commonly observed in turbidity current or fluidized/liquified flow deposits (Middleton and Hampton, 1976). The absence of grading, the presence of flat upper contacts, and the presence of deformed varves beneath a few of the massive beds indicate that they are products of either grain flows or debris flows (using the terminology of Middleton and Southard, 1984).

Seismically triggered, sediment gravity flows are common in lacustrine and marine environments. In the eastern North Pacific Ocean west of Washington, for example, large recurrent turbidity flows have been ascribed to thrust earthquakes on the Cascadia subduction zone (Adams, in press). Sediment gravity flows in Saanich Inlet may record these earthquakes, as well as large intraplate quakes centred on or near Vancouver Island. The number of massive beds deposited within the 1400-year time span of core 3 is compatible with a seismic trigger, since large earthquakes probably affect southern Vancouver Island, on average, once every 50-200 years. Unfortunately, we are unable at present to date the massive beds precisely because the uppermost sediments in the inlet were not recovered during coring. We hope to resolve this problem by attempting to sample the topmost 1 m of sediment with a specially designed box corer.

In conclusion, the Holocene fill in Saanich Inlet comprises fine grained varved sediments which were deposited slowly from suspension and resedimented beds, some or all of which were probably emplaced during earthquakes.

ACKNOWLEDGMENTS

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Pleistocene tephra in central British Columbia

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Clague, J.J. Pleistocene tephra in central British Columbia; in Current Research, Part E, Geological Survey of Canada, Paper 90-1E, p. 257-261, 1990.

Abstract

More than 25 thin layers of silicic tephra (''Mexican Hill tephra'') have been found in a bluff in the valley of Lightning Creek, central British Columbia. No other Pleistocene or Holocene tephras are known from this region. Mexican Hill tephra is the product of one or more eruptions from a distant volcano some time before 32 000 BP and probably before 51 000 BP. The tephra resembles fine grained rhyolitic and dacitic tephras in southern British Columbia a few hundred kilometres to the south, but the chemical and mineralogical differences are sufficiently great that no correlations can be made. Possible sources of the Mexican Hill tephra include volcanoes in the Cascade Range of Washington and Oregon, the Coast Mountains of British Columbia, and the Wrangell Mountains or eastern Aleutian arc of southern Alaska.

Résumé

Plus de 25 couches minces d'un téphra siliceux (« téphra de Mexican Hill ») ont été trouvées dans un escarpement de la vallée du ruisseau Lightning, dans la partie centrale de la Colombie-Britannique. On ne connaît pas d'autres téphras du Pléistocène et de l'Holocène dans cette région. Le téphra de Mexican Hill est le produit d'une ou plusieurs éruptions d'un volcan éloigné survenues il y a plus de 32 000 ans et probablement, il y a plus de 51 000 ans. Le téphra ressemble à des téphras rhyolitiques et dacitiques à grain plus fin de la partie sud de la Colombie-Britannique, situés à quelques centaines de kilomètres au sud, mais les différences chimiques et minéralogiques sont suffisamment grandes pour interdire toute corrélation. Les sources possibles du téphra de Mexican Hill comprennent des volcans de la chaîne des Cascades des états de Washington et de l'Oregon, de la chaîne côtière de la Colombie-Britannique et de la chaîne de Wrangell ou de la région est de l'arc aléoutien de la partie sud de l'Alaska.

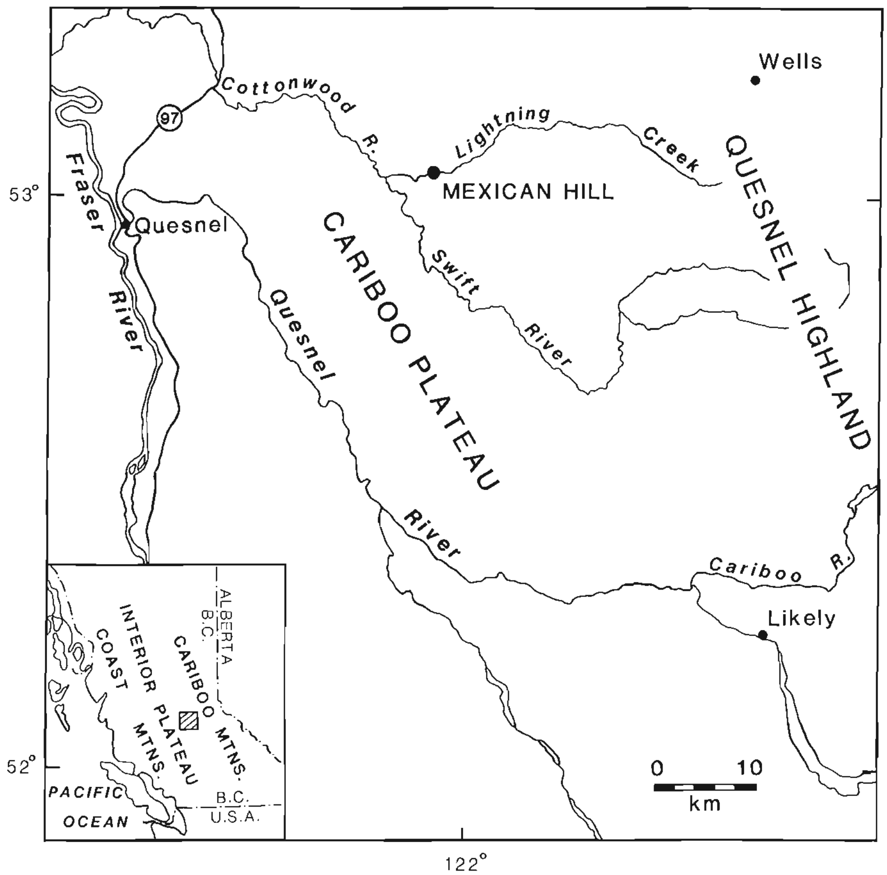

Figure 1. Location map.

INTRODUCTION

Thin layers of silicic tephra have been found in an exposure along Lightning Creek in central British Columbia ("Mexican Hill", Fig. 1). This is an interesting and unusual find because the site is 200 km from the nearest possible source and because there are no other known dacitic or rhyolitic tephras in this region (the site is even beyond the limits of Mazama tephra, the most widespread late Quaternary tephra in western North America).

This report decribes the Mexican Hill tephra and its stratigraphic context, discusses its possible relationship to Pleistocene tephras in southern British Columbia (Westgate and Fulton, 1975), and comments on possible sources.

SETTING AND STRATIGRAPHIC CONTEXT

The tephra site is located on Cariboo Plateau west of Quesnel Highland (Fig. 1; Mathews, 1986). This part of the Interior Plateau is underlain by sedimentary and volcanic rocks of Proterozoic to Miocene age and by minor Mesozoic granitic rocks (Tipper et al., 1979). These rocks are mantled by Quaternary sediments, mainly till and glaciofluvial gravel. The plateau surface east and southeast of Quesnel is incised by streams such as Cottonwood River, Quesnel River, and Lightning Creek which head in Quesnel Highland and the Cariboo Mountains farther east.

The tephra is exposed in a ravine on the north side of Lightning Creek valley, 30 km east of Quesnel (Fig. 1). It occurs within nonglacial sediments which are underlain and overlain by glacial deposits (Fig. 2). At the base of the section, sharpstone diamicton (colluvium) with interbeds of silt and sand overlies bedrock. The diamction, in turn, is overlain by a complex sequence of bedded sand and gravel (proglacial? deltaic sediments). The deltaic sequence coarsens near the top and is overlain by weakly stratified diamicton with striated clasts and lenses of poorly sorted gravel (till). A possible yellow-brown paleosol is present in the upper part of this unit. The diamicton is capped by patchy thin oxidized gravel.

These weathered sediments are overlain by about 2.5 m of interbedded silt and sand containing fossil plant material and, near the bottom, small scattered pebbles (alluvium or slopewash passing upward into lacustrine sediments). More than 25 layers of white tephra, ranging from 0.2 to 4 mm in thickness, are present in the upper part of the unit (Fig. 3).

The plant-bearing silt and sand are overlain by unfossiliferous laminated mud (lacustrine or glaciolacustrine sediments). This mud grades upward into stratified sand and minor silt (deltaic sediments), and the latter are unconformably overlain by diamicton (till and colluvium) and gravel (glaciofluvial sediments).

Twigs collected from silty and sandy lacustrine sediments 25 cm below the lowest tephra layer yielded conventional radiocarbon ages of >40 000 BP and >51 000 BP (GSC-4383 and GSC-4611HP). A twig from the same unit,

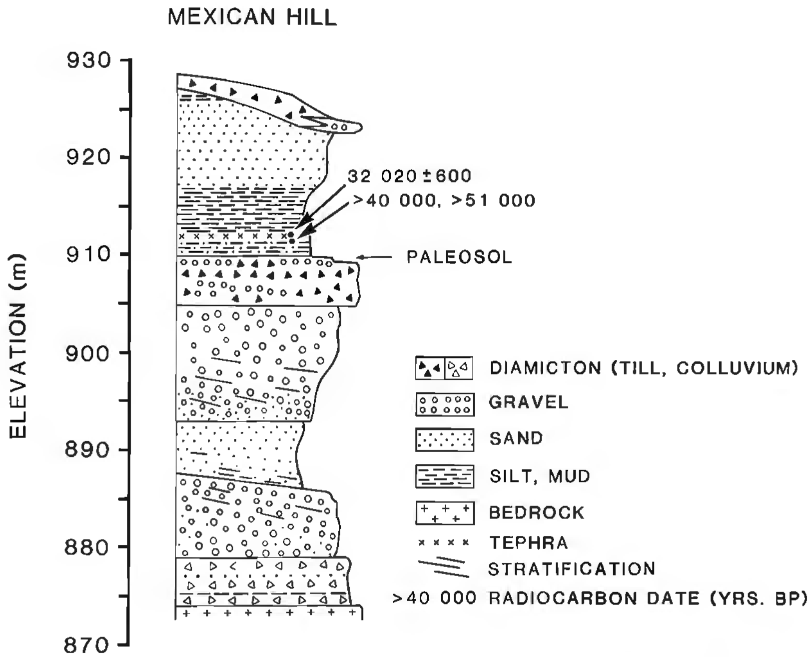

Figure 2. Geological section at Mexican Hill, showing stratigraphic position of tephra layers.

16 cm above the lowest tephra layer and 4 cm below the highest tephra layer, was dated by the accelerator mass spectrometry method (AMS) at 32 020 ± 600 BP (TO-532). Because the sediments between the dated horizons record apparently continuous lacustrine sedimentation, there is no reason to suspect a large difference in age. It is possible that the "infinite" ages are from reworked material. Many of the dated twigs, however, have delicate branches that probably would not survive reworking from an older unit. The 32 020 age thus may be too young and should be viewed as a minimum.

This leaves some uncertainty about the age of the Mexican Hill tephra. Regional considerations indicate that the upper till at this site is Late Wisconsinan. The upward-coarsening, lacustro-deltaic sequence directly underlying this till may record the formation and deepening of a proglacial lake during the early part of the last (Fraser = Late Wisconsinan) glaciation. It is difficult to reconcile this, however, with the evidence that the organic-rich sediments, which appear to conformably underlie the deltaic sand and laminated mud, are >32 000 years old (and probably >51 000 years old). It is possible, therefore, that the lacustro-deltaic sequence does not date to the early part of the Fraser Glaciation, but, rather, formed during an Early Wisconsinan or pre-Wisconsinan glaciation.

DESCRIPTION OF TEPHRA

Most of the Mexican Hill tephra layers consist largely or entirely of fine volcanic ash. Some, however, contain abundant nonvolcanic detritus that must have been washed into the lake along with the volcanic material. It is not clear if the layers are products of a single eruption or of several closely spaced eruptive events. If there was only one eruption, all tephra layers above the lowest occurrence must have formed by reworking of a primary airfall deposit that mantled slopes bordering the lake.

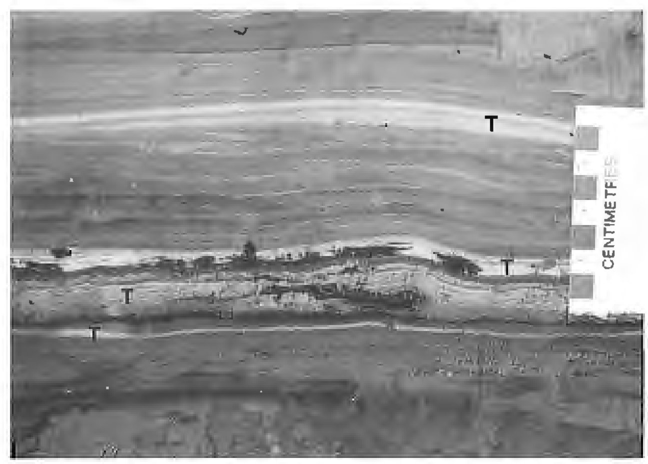

Figure 3. Tephra layers (T) in nonglacial lacustrine sediments at Mexican Hill.

Chemical and mineralogical analyses were performed on a 4 mm thick tephra layer, 4 cm above the base of the Mexican Hill sequence (Tables 1 and 2). The tephra is very fine grained (most particles < 60 µm diameter), which suggests a source at least a few hundred kilometres away. Glass with a refractive index of 1.508 ± 0.002 constitutes 91 % of the tephra by weight (Table 1). Light minerals (8 % by weight) consist largely of plagioclase and quartz. The dominant heavy minerals are hornblende, clinopyroxene (probably augite), hypersthene, epidote, magnetite, and ilmenite.

COMPARISONS WITH OTHER TEPHRAS AND POSSIBLE SOURCES

Several thin, fine grained rhyolitic and dacitic tephras of late Pleistocene age are present in southern British Columbia, about 300-400 km southeast of Mexican Hill (Fig. 4; Westgate and Fulton, 1975). Some of these share compositional

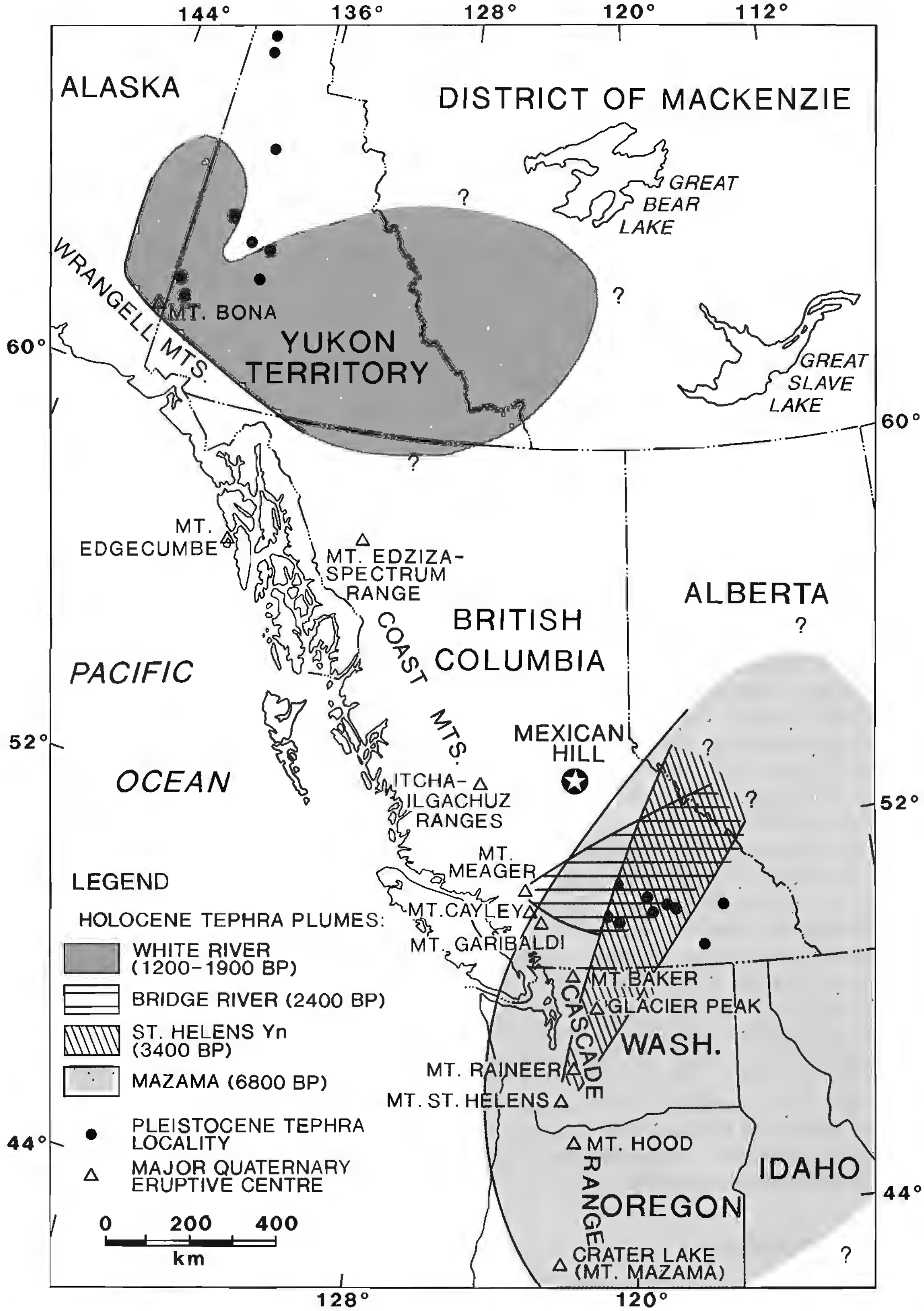

Figure 4. Distribution of widespread Holocene tephras and occurrences of Pleistocene tephras in British Columbia and Yukon Territory. Some possible sources of the Mexican Hill tephra are also shown. Adapted from Clague (1989, Fig. 1.11).

Table 1. Mineralogy of Mexican Hill tephra.

Glass — 91% by weight[a]	
Light minerals — 8% by weight	
Feldspar (mainly plagioclase)[b]	55%
Quartz	40%
Muscovite	5%
Chlorite	<1%
Heavy minerals — 1% by weight	
Opaque minerals and polycrystalline aggregates	26%
Hornblende	22%
Clinopyroxene (probably augite)	17%
Orthopyroxene (hypersthene)	15%
Epidote	11%
Chlorite	4%
Apatite	2%
Sphene	<1%
Tourmaline	<1%
Garnet	<1%
Muscovite	<1%
Calcite	<1%

Notes: Determinations made on 20-300 μm fraction by D.N. Eden, Division of Land and Soil Sciences, New Zealand Department of Scientific and Industrial Research. A solution of sodium polytungstate and water was used for mineral separations. 300 grains were counted for heavy mineral proportions. Percentages of light minerals were determined from X-ray diffraction peak intensities and were checked by optical examination.

[a] Refractive index = 1.508 $\pm$ 0.002.

[b] Andesine composition.

Table 2. Composition of volcanic glass from Mexican Hill tephra.

	Analyzed shard												
	1	2	3	4	5	6	7	8	9	10	11	x	σ
SiO_2	72.15	73.58	73.84	74.35	74.52	74.51	74.80	74.54	74.49	73.31	75.07	74.11	0.84
Al_2O_3	15.83	14.73	14.57	14.52	14.38	14.27	14.45	14.37	14.46	14.80	14.36	14.61	0.43
TiO_2	0.36	0.30	0.29	0.29	0.32	0.24	0.27	0.30	0.30	0.31		0.30	0.03
FeO*	1.35	1.59	1.44	1.41	1.37	1.49	1.49	1.43	1.42	1.83	1.25	1.46	0.15
MgO	0.38	0.53	0.40	0.37	0.39	0.38	0.36	0.39	0.40	0.49	0.38	0.41	0.05
CaO	2.47	1.55	1.48	1.31	1.34	1.32	1.32	1.43	1.38	1.76	1.33	1.52	0.34
Na_2O	4.57	4.35	4.50	4.27	4.20	4.13	4.09	4.27	4.23	4.37	4.28	4.30	0.15
K_2O	2.75	3.29	3.28	3.25	3.30	3.30	3.23	3.17	3.19	3.04	3.14	3.17	0.16
Cl	0.14	0.08	0.20	0.23	0.19	0.25		0.10	0.15	0.09	0.20	0.16	0.06
H_2O**	5.34	5.55	4.48	4.59	2.99	8.56	4.61	3.43	5.50	3.89	4.95	4.90	1.47

Notes: Determinations made with JEOL 733 electron microprobe with a 10 μm beam diameter and 8nA beam current by P.C. Froggatt (Victoria University, Wellington, New Zealand). Glass standards: comendite obsidian (KN-18) and basaltic glass (VG-99).

* All Fe calculated as FeO.

** Water, by difference.

affinities with the Mexican Hill tephra; the differences, however, are sufficiently great that it is not possible to correlate the latter with any of the former. Most of the tephras in southern British Columbia are rich in cummingtonite and have only small amounts of orthopyroxene. Westgate and Fulton (1975) stated that these tephras probably were erupted from Mount St. Helens in Washington. In contrast, the Mexican Hill tephra contains little or no cummingtonite and may have originated elsewhere. A distant source for this tephra is indicated by its fine texture and by the fact that there are no Pleistocene volcanoes within 200 km of Mexican Hill capable of producing silicic magmas. Possible sources include volcanoes in the Cascade Range of Washington and Oregon, the Coast Mountains of British Columbia, and the Wrangell Mountains or eastern Aleutian arc of southern Alaska (Fig. 4). Of these, Alaskan volcanoes seem the least likely because they are more than 1500 km from Mexican Hill.

ACKNOWLEDGMENTS

D.C. Froggatt (Victoria University, Wellington, New Zealand) and D.N. Eden (Division of Land and Soil Sciences, New Zealand Department of Scientific and Industrial Research) conducted the chemical and mineralogical analyses, respectively, under contract to the Geological Survey of Canada. C. Davis and R.N. Gopal drafted the figures.

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Rock avalanches in the Pelly Mountains, Yukon Territory

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Jackson, L.E., Jr. and Isobe, J.S. Rock avalanches in the Pelly Mountains, Yukon Territory; in Current Research, Part E, Geological Survey of Canada, Paper 90-1E, p. 263-269, 1990.

Abstract

Five large rock avalanches have occurred in Pelly Mountains in postglacial time. Four of these have occurred in noncarbonate rock units. Excessive travel distances and effective coefficients of friction of these rock avalanches are comparable to those reported for similar failures in Mackenzie Mountains. Rock avalanches in this region of the Cordillera have low effective coefficients of friction for their volumes compared to other rock avalanches reported in the literature.

Résumé

Il y a eu cinq grandes avalanches de pierre dans les monts Pelly pendant la période postglaciaire. Quatre se sont produites dans des unités de roches non carbonatées. Les distances importantes parcourues et les cœfficients effectifs de friction de ces avalanches sont comparables à ceux signalés pour des types d'avalanches semblables survenues dans les monts Mackenzie. Les avalanches de pierre dans cette région de la Cordillère ont de faibles cœfficients effectifs de friction pour leur volume par rapport à d'autres avalanches de pierre signalées dans la documentation publiée.

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INTRODUCTION

Rock avalanches or sturzstroms (Hsü, 1975) are catastrophic failures of large masses of rock from mountain sides. The dynamics of rock avalanche mobility are dictated primarily by the volume of the failed mass. Beyond a threshold range of $1x10^5$ to $1x10^6$ m^3, a failed mass flows or streams in a fluid-like way. This is in contrast to the free tumbling of individual blocks of rock noted in failures below this threshold (Eisbacher and Clague, 1984). These streaming avalanches commonly travel impressive distances beyond the distances predicted assuming sliding friction coefficients alone. Many hypothetical mechanisms have been suggested to account for streaming flow, excessive runout distances, and low apparent coefficients of friction (see Melosh, 1987 for a summary).

Surficial geology mapping in Pelly Mountains, Yukon Territory has identified five rock avalanche deposits in the 1-5 $x10^6$ m^3 range (Fig. 1). Unlike the rock avalanches studied by Eisbacher (1979) and McLellan (1983) in the carbonate terrain of the Mackenzie Mountains 250 km to the northeast, these rock avalanches occurred predominantly in noncarbonate lithologies. A study of these deposits was undertaken in order to compare and contrast these failures with those in the Mackenzie Mountains and elsewhere.

SETTING

Pelly Mountains are composed of allochthonous Mesozoic and Paleozoic sheared ophiolites and cataclastic rocks discontinuously preserved as klippen resting on folded and thrusted autochthonous sedimentary rocks of late Precambrian to early Mesozoic age. These were intruded during the late Cretaceous by quartz monzonite plutons (Tempelman-Kluit, 1979). The Pelly Mountains supported an extensive ice cap during the last (McConnell) glaciation (Jackson, 1989). They are characterized by classic alpine landforms such as horns, arêtes, and cirques. Elevations reach 2366 m with a maximum relief between ridge tops and adjacent major valley bottoms of 750 m over horizontal distances of 3 to 4 km.

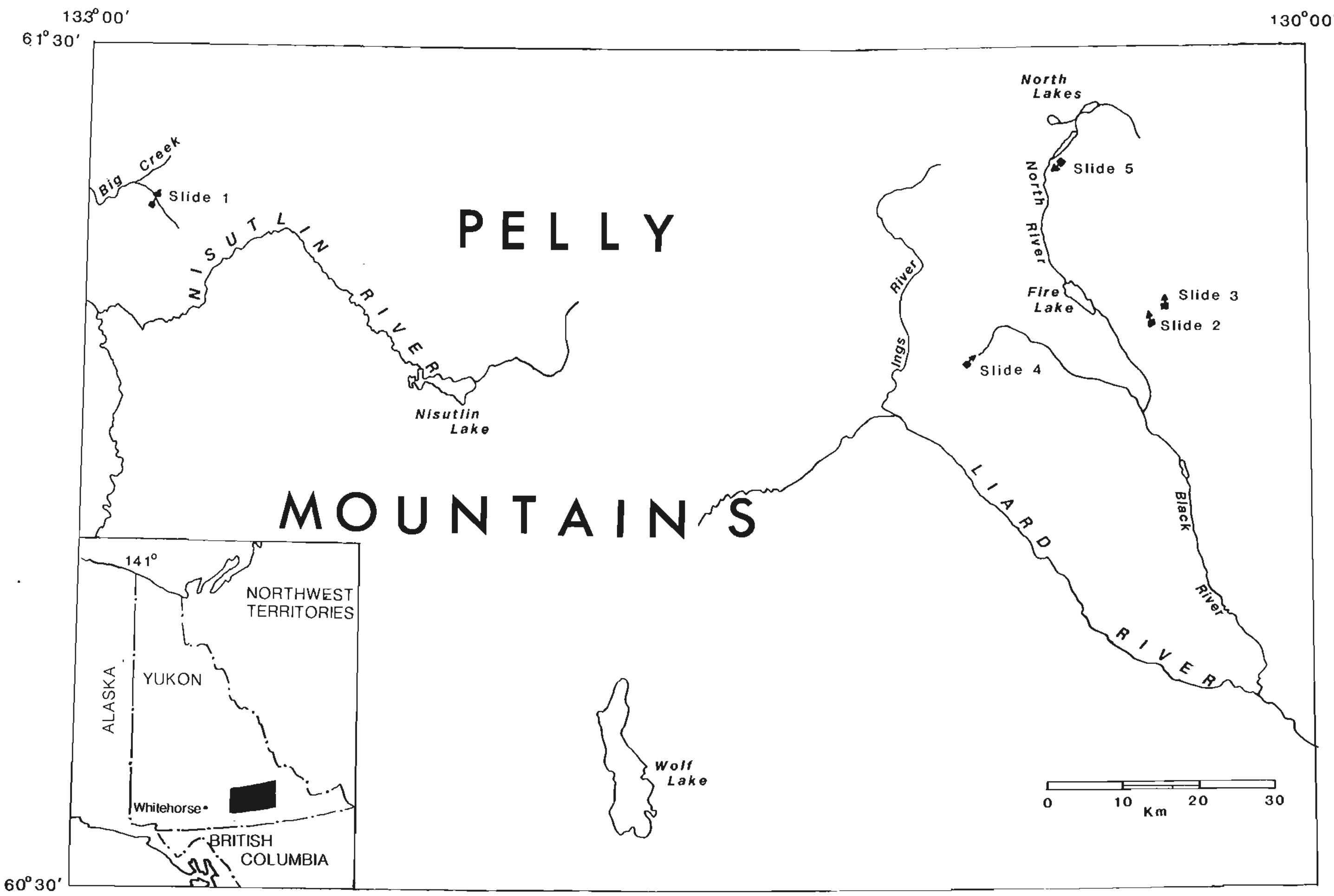

Figure 1. Pelly Mountains region and locations of debris avalanches. Arrows on symbols indicate the direction of rock avalanche movement.

Table 1. Debris avalanche summary

Slide	Location	Lithology[1]	Length (m)	Width[2] (m)	Volume[3] ($10^6 m^3$)	E.C.F.[4]	L_e[5] (m)
1	61°17.8′N 132°50.7′W	peridotite	3700	210 560 330	3.1	0.20	2520
2	61°10.8′N 130°24.8′W	quartz chlorite feldspar schist (mylonite)	2390	280 780 490	2.1	0.29	1300
3	61°12.6′N 130°22.2′W	recrystalized limestone/ marble blastomylonite	2340	260 900 630	2.5	0.17	1680
4	61°08.3′N 130°53.1′W	biotite,quartz monzonite	3400	410 1180 850	4.7	0.20	2310
5	61°21.0′N 130°36.5′W	muscovite,biotite, quartzofeldspathic gneiss	1580	100 400 310	0.98	0.27	890

1. After Templeman-Kluit, 1977
2. Upper value, minimum; middle value, maximum; lower value, mean
3. Estimate based upon assumed deposit thicknesses or reconstruction of detached rock mass in rock avalanche source area.
4. Effective coefficient of friction
5. Excessive travel distance

DESCRIPTION OF ROCK AVALANCHE DEPOSITS AND THEIR ENVIRONMENTS

Glacially induced oversteepening of valley sides is likely the ultimate cause of all the rock avalanches in Pelly Mountains (Fig. 2-6). All had their failure areas in arêtes separating adjacent steep-sided glaciated valleys (rock avalanches 2 and 3) or in cirque headwalls (rock avalanches 1, 4, and 5). All slides have occurred in postglacial time; none display collapse features or anomalous margins suggestive of contact with active or stagnant glacial ice. The physical dimensions of the rock avalanches are summarized in Table 1 along with parameters associated with transport dynamics.

Two parameters commonly used to measure and compare the mobilities of rock avalanches are excessive travel distance (L_e; Hsü, 1975) and effective coefficient of friction (E.C.F.; Shreve, 1968) or Fahrboschung (travel angle; Heim, 1932) where:

$$L_e = L - H/\tan 32° \quad (1)$$

$$\text{and E.C.F.} = H/L \quad (2)$$

L is the planimetric length of the rock avalanche deposits measured between the uppermost boundary of the failure area the toe of the deposits.

H is the height of the rock avalanche measured between the uppermost boundary of the failure area and the toe of the deposit.

Each rock avalanche is briefly described below. Volume for slide 1 was calculated two ways: measurement of the failed mass by extending contours across the failure area and as the product of length, width, and an estimated mean thickness (2.5 m). These two estimates agreed within 10%. The other rock avalanche volumes were computed as products of length, mean width, and estimated thicknesses alone. We regard these estimates to be conservative. Actual values could be as much as X5 larger.

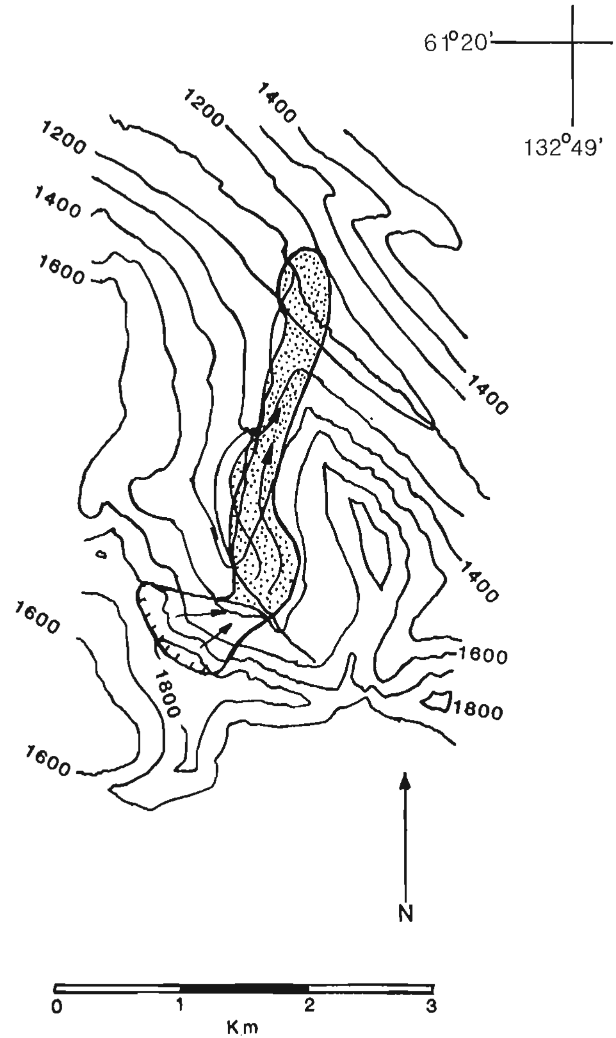

Figure 2. Rock avalanche 1.

Rock avalanche 1 (NAPL A25289 99-100)

This rock avalanche was the only one studied in the field. The failure area was along the head of a north facing cirque in a steeply inclined (approximately 60°), highly jointed, ultramafic dyke complex (Fig. 2, 7, and 8). The slope in the area of detachment was approximately 40°. The slide descended 280 m to the cirque floor then climbed 105 m up the adjacent slope travelling northward along a superelevated arcuate path. It then descended to the valley floor climbing 59 m up the opposite (west) side of the valley along a second superelevated arcuate path. From there it travelled 1400 m along the centre of the valley down an average gradient of 7°, finally running 20 m up the side of a confluent valley.

The superelevations defined by the deposits provided opportunity to determine the velocities of the rock avalanche at the superelevations using the forced vortex equation (Hungr *et al.*, 1984 after Mears, 1977):

$$h = kbv^2/rg \quad (3)$$

where h is the elevation difference between the two sides of the rock avalanche at the superelevation, b is the surface width of the debris avalanche, v is the mean velocity, g is the gravitational acceleration constant, r is the mean curvature radius, and k is a correction coefficient assumed to be 1 for rock avalanches (O. Hungr, Thurber Consultants Ltd., Vancouver, personal communication, 1989). Velocities of 26 m/s and 29 m/s were calculated for the flow of the rock avalanche along the upper and lower superelevated segments of its path. The flow of the rock avalanche was apparently nonturbulent: a secondary rockfall from a grey weathering ultramafic unit fell onto the west margin of the debris avalanche near the base of the detatchment area. The deposits of this failure were stretched out as a distinct stripe-like veneer less than 1 m thick as they were rafted along on top of the debris avalanche (Fig. 7).

Rock avalanche 2 (NAPL A25289 148-149)

This landslide detatched from the north side of an arête and descended down the valley of an unnamed tributary of North River. Detachment occurred on a slope of approximately 40°. The slide ran 40 m up the opposite valley wall along a broad superelevated arcuate path and then followed the centre of the valley to the west and southwest along a gradient ranging between 7° and 13°. A velocity of 25 m/s was calculated for the superelevated portion of the travel path using the forced vortex equation. The rock avalanche is very fresh in appearance with the bouldery surface lacking any discernable vegetative cover when viewed on air photographs. This rock avalanche and its environment are noteworthy on two additional counts. First, displacement has occurred along large cracks in the arête immediately east of the failure area (Fig. 3). Because of these extensive cracks and adjacent rock avalanche activity, further rock avalanche activity is likely in this area. Second, the toe area of the rock avalanche deposit overlies vegetated ridged and hummocky deposits which could be an older and possibly more extensive rock avalanche deposit or deposits. Field work at this site will be required to substantiate this.

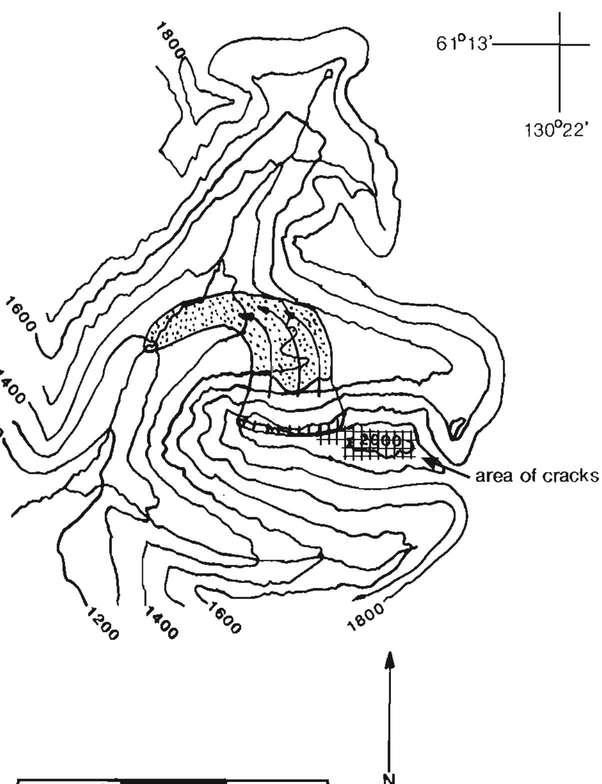

Figure 3. Rock avalanche 2.

Rock avalanche 3 (NAPL A25289 148-149)

This failure was the only one of the five studied to have occurred in carbonate rocks. It detached from the northeast face of an arête on a 25° slope, and travelled 700 m across the adjacent valley where it ran 20 m up the opposite valley side (Fig. 4). From there, part of the slide apparently slid back to the centre of the valley where it continued downvalley for another 600 m. Although the E.C.F. and L_e for this failure were calculated using this path, most of the volume of this slide did not travel past the runup area. The surface of the rock avalanche is obscured by vegetal cover with the exception of large bedrock blocks with dimensions of several metres along its southern margin.

Rock avalanche 4 (NAPL A25289 143-144)

The source of this rock avalanche could not be identified for certain through airphoto interpretation as there are three amphitheatre-like features near the top of the cirque wall above it that could be failure scars. The details of these features are obscured by permanent snow fields and a rock glacier. Depending on the source, detatchment of the failed

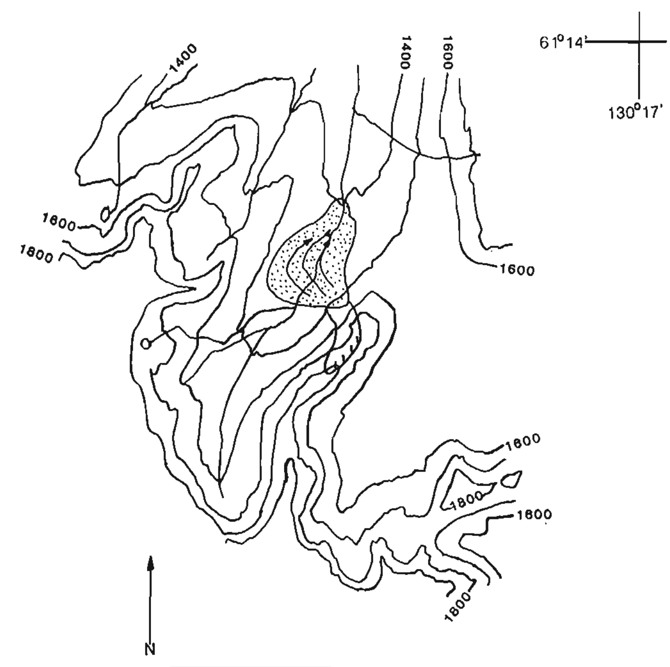

Figure 4. Rock avalanche 3.

mass occurred on slopes ranging from 32° to 44°. The slide traversed 700 m across a nearly horizontal bedrock bench where much of the slide mass piled up and remained. The balance of the avalanche descended 140 m down an average 20° bedrock slope travelling 500 m in a horizontal sense (Fig. 5). It then travelled an additional 1000 m along a 5° slope on the valley floor. The rock avalanche deposits are dotted by angular blocks with dimensions in the 5 to 10 m range likely reflecting the coarse jointing in the biotite quartz monzonite bedrock from which it originated. Although unvegetated in its upper three quarters, the lower one quarter of the slide is largely covered by alpine tundra vegetation.

Rock avalanche 5 (NAPL A25289 129-130)

This failure was the smallest investigated and possessed the smallest L_e and largest E.C.F values. Two types of failure are apparent here: rock avalanching and sliding of intact or semi-intact masses of the same mountain side. The detachment area is located on the southside of an arête separating adjacent cirques and has a slope of approximately 36°. The rock avalanche portion of the failure descended the mountain side and filled a small valley transverse to its path (Fig. 6). A small part of the rock avalanche turned 90° and followed this valley for about 700 m descending a slope of about 14° to 20°. The rock avalanche deposits are almost entirely unvegetated.

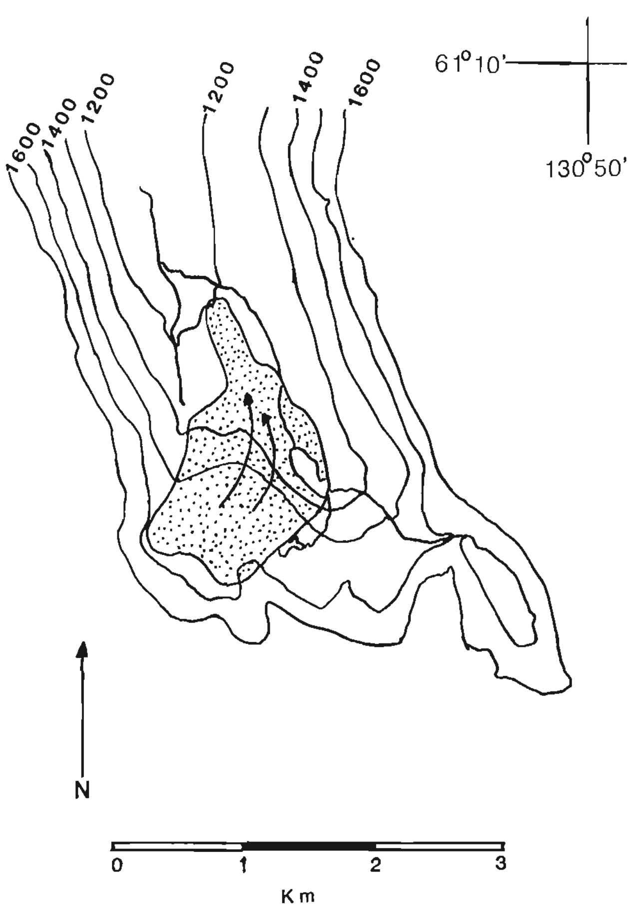

Figure 5. Rock avalanche 4.

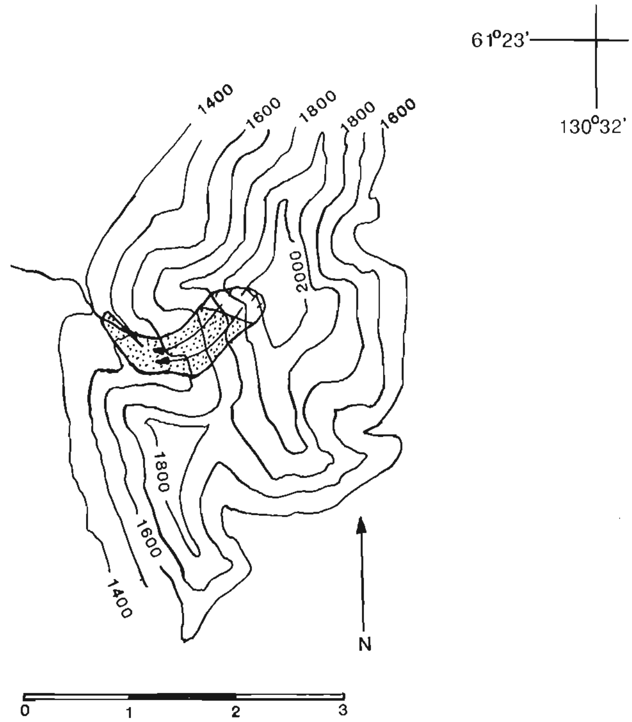

Figure 6. Rock avalanche 5.

DISCUSSION

The runout distances of large rock avalanches are broadly directly related to their volumes through a power relationship (Davies, 1982). As pointed out by Eisbacher and Clague (1984), this has led to the suggestion in the literature that E.C.F. is inversely related to rock avalanche volume. Figure 9 is a compilation of mobility and volume data for terrestrial, lunar, and martian rock avalanches by Melosh (1987) documenting this relationship. Values for Mackenzie Mountains and Pelly Mountains rock avalanches have been added to this plot. Although this inverse relationship between E.C.F. and rock avalanche volume does apparently hold, there is much scatter. It is interesting to note that E.C.F. values for the Mackenzie Mountains and Pelly Mountains are as low as some rock avalanches two or three orders of magnitude larger. Eisbacher (1979) and Eishbacher and Clague (1984) discussed the importance of factors other than volume in determining runout distance. Among these were complexities in avalanche path, lithology, and the incorporation of wet soil, snow, ice, or water into the rock avalanche. In their view, this increase in moisture content through the incorporation of snow, ice, or wet sediment would tend to change the character of a rock avalanche into that of a debris flow, further increasing its mobility. This was likely a factor in Mackenzie Mountains and Pelly Mountains rock avalanches. Snow and ice are typically present nine or ten months of the year in the alpine areas of this region. Valley floors are underlain extensively by saturated alpine tundra and bog during snow-free months. The abundance of snow and water along rock avalanche paths may explain the generally low E.C.F. values with respect to rock avalanche volumes in this area of the Cordillera.

Figure 7. View (south) of the upper two thirds of rock avalanche 1. S indicates detatchment area. Dashed lines mark superelevated margins of the rock avalanche path. Stipple area marks the grey weathering peridotite blocks discussed in the text.

Figure 8. Abrupt margin of upper superelevation of rock avalanche 1 (looking north). Arrow indicates man for scale.

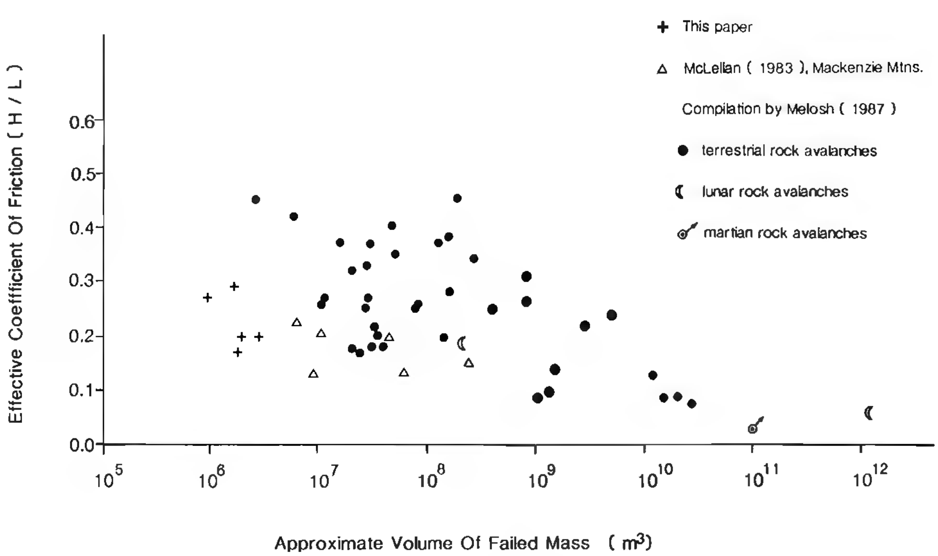

Figure 9. Effective coefficient of friction versus rock avalanche volume data from Melosh (1987) including rock avalanches from Mackenzie and Pelly Mountains.

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A jökulhlaup origin for boulder beds near Granite Canyon, Yukon Territory

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Lye, D., Jackson, L.E., Jr., and Ward, B. A jökulhlaup origin for boulder beds near Granite Canyon, Yukon Territory; in Current Research, Part E, Geological Survey of Canada, Paper 90-1E, p. 271-275, 1990.

Abstract

A bed of large boulders is exposed for 3.8 km along Pelly River below Granite Canyon, Yukon. The bed was deposited during a catastrophic draining of a large glacial lake through Granite Canyon. Minimum mean velocity was in the range of 3.5-7 m/s.

Résumé

Une couche constituée de grands blocs affleure sur 3,8 km le long de la rivière Pelly, au-dessous du canyon Granite au Yukon. Elle a été mise en place au cours du drainage catastrophique d'un grand lac glaciaire à travers le canyon Granite. La vitesse moyenne minimale était de l'ordre de 3,5 à 7 m/s.

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INTRODUCTION

Unusual beds of extremely large boulders were noted during surficial mapping in the Granite Canyon area along Pelly River, Yukon Territory. The boulders are situated in the area of the McConnell limit of the Selwyn Lobe of the Cordilleran Ice Sheet (Hughes et al., 1969). This report describes these boulder beds and investigates their genesis.

SETTING

The boulder gravels studied are exposed for 3.8 km along the upper part of a stream-cut escarpment on the north side of Pelly River valley, approximately 25 km upstream of the village of Pelly Crossing, Yukon Territory (Fig. 1 and 2).

The boulder beds are located at the downstream end of Granite Canyon, a 5 km long narrow canyon cut through Mesozoic biotite granite and andesite (Tempelman-Kluit, 1984). Granite Canyon is a geologically recent feature and trends southward from the wider Pelly River valley which is blocked immediately west of Granite Canyon by ice stagnation deposits. The deposits mark the approximate western limit of glacial ice in Pelly River valley during McConnell Glaciation (Bostock, 1966; Hughes et al., 1969). McConnell Glaciation began more than 26 000 years ago (Jackson, 1989) and had retreated at least 60 km east from Granite Canyon by ca. 12 600 years ago (Ward, 1989). The boulder bed is truncated by one of a series of terraces cut into McConnell outwash and ice stagnation deposits.

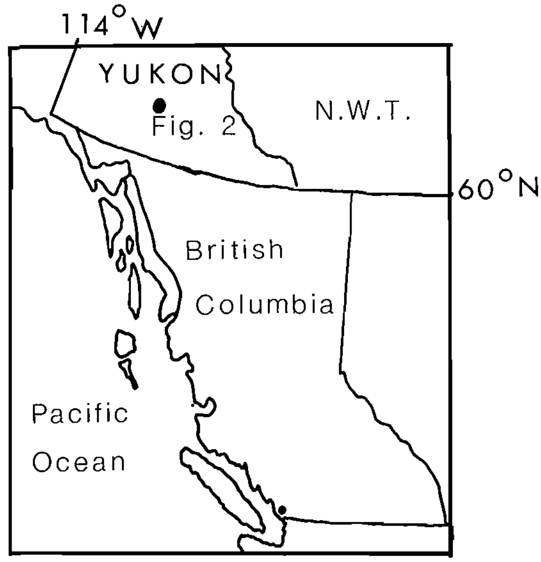

Figure 1. Location of the Granite Canyon area in southern Yukon Territory.

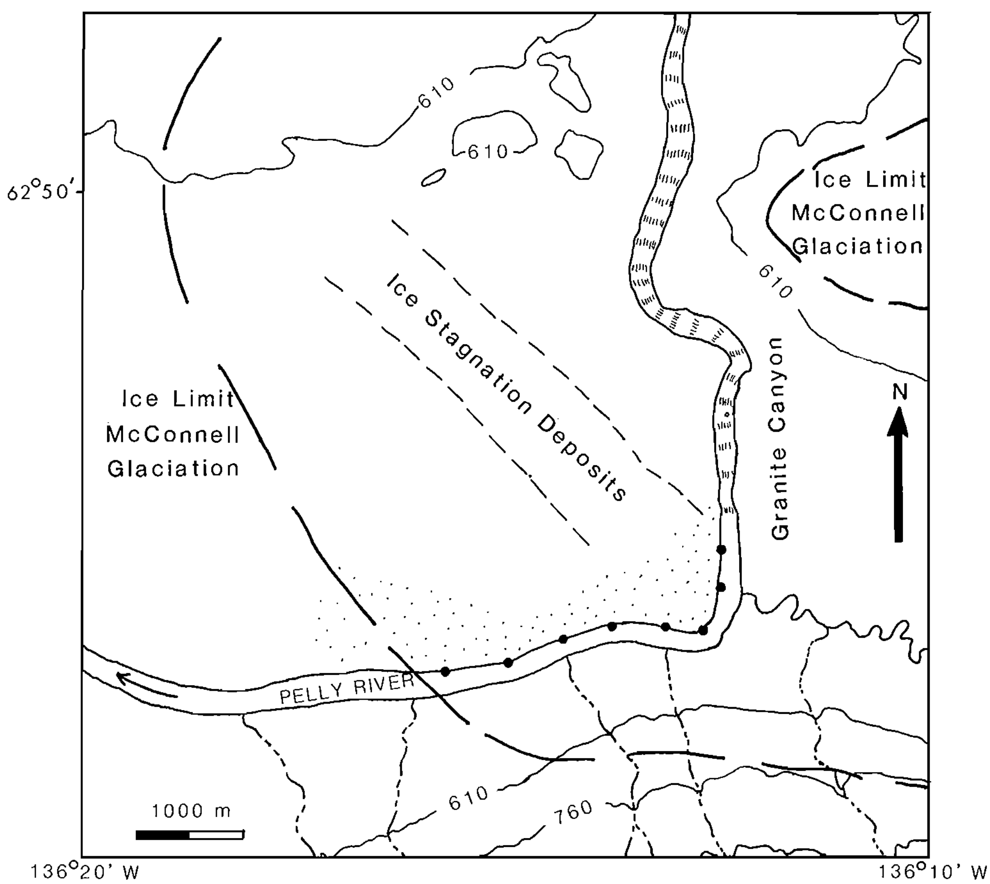

Figure 2. Study area. Dots indicate boulder measurement stations. The stipple pattern denotes the terrace underlain by boulder gravels.

DESCRIPTION OF THE BOULDER GRAVELS

General sedimentology and stratigraphy

Figure 3 depicts the stratigraphic setting of the boulder gravel. Beginning at river level, the lower 3-5 m of the section is covered by colluvium. Unit 1 consist of the next 26 m of unsorted coarse sands and gravels. Clasts in this unit are matrix-supported, subangular to well rounded and range from 3-35 cm in size. In local bouldery zones, clasts range to 50 cm.

Unit 1 is capped by Unit 2, which consists of 2-5 m of large, subangular to subrounded andesite boulders. The ultimate source of these boulders is Granite Canyon although the striations and polish on some of these clasts indicates that they may have been initially glacially eroded and deposited in drift before being redeposited in their present location. They are supported by a matrix of coarse sands and cobbles (Fig. 4). The contact between units 1 and 2 is depositional and gradational. The boulder gravel is capped by 1 to 1.5 m of sand in which a well developed Holocene soil has formed. The large clasts are discontinuously exposed along the terrace surface where they protrude through the overlying sand.

Clasts in unit 2 range in size from 15-160 cm in exposed intermediate axial dimension. Some of the clasts have glacially striated or polished surfaces. The unit is massive with no indication of any bedforms or stratification except for local imbrication of the boulders.

The largest boulders were found at the extreme upstream end of the deposit, though boulders exceeding 1.5 m were found throughout the entire length of the exposure. Although the boulders appear to be continuous along the length of the terrace, along one stretch approximately in the middle of the exposure, the boulder bed becomes discontinuous, laterally fining out in a downstream direction; however, the boulders again become evident immediately downstream.

Interpretation

The coarse matrix, rounded clasts, and poor sorting of unit 1 are consistent with a proximal glaciofluvial environment (e.g., Rust and Koster, 1984).

The boulder clasts of unit 2, commonly larger than 1 m and with a cobble matrix, are not compatible with a glaciofluvial flow regime driven directly by seasonal snow and ice melt. Such environments may routinely transport clasts up to perhaps 0.5 m (see data summaries in Nowak, 1973 and Church and Gilbert, 1975). Rather, the boulder beds are consistent with a catastrophic flood event (Baker, 1974, 1984; Bradley and Mears, 1980; Baker and Costa, 1987).

Interpretation of surficial geology mapping (by Ward and Jackson, in progress) suggests that the flood event was likely a jökulhlaup caused by the breaching of a drift dam with Granite Canyon being cut or exhumed by the resulting jökulhlaup. The source of the jökulhlaup waters was an extensive lake, delineated today by lake deposits (Fig. 5) which extended from near the upstream end of Granite Canyon to approximately 40 km up Macmillan and Pelly valleys (Fig. 5). This lake was apparently ponded at least in part by stagnant ice in the Granite Canyon area. The drainage history of the lake remains to be investigated in detail; however, with a volume we estimate to have been of the order of 100 to 150 km^3, drainage of even a small percentage of the lake through Granite Canyon would have generated a catastrophic flood with extreme velocities.

ESTIMATION OF JÖKULHLAUP VELOCITY

Method of analysis

Paleohydrological reconstructions of flood depth and peak discharge require definable channel boundaries in order to estimate water-surface slopes and channel geometries (Bradley and Mears, 1980). There is no field evidence of clear vertical or lateral jökulhlaup path boundaries in this area due to subsequent terrace cutting by Pelly River. It is possible, however, to infer some limited information from the boulder gravels themselves. Coarse clast deposits in sedimentary sequences have been used extensively to estimate flow velocities (Novak, 1973; Maizels, 1983). Grain size data from the large boulders present in unit 2 allow for the calculation of flow velocities through the use of an established engineering formulas.

Figure 3. Stratigraphic setting of boulder bed (unit 2). View to the north.

Figure 4. Typical exposure of the boulder bed (unit 2). Divisions on the rod are 0.3 m.

Data collection

In order to sample the 3.8 km of exposure, eight stations approximately 500 m apart were established along the river (Fig. 2). At each station, a stratigraphic section was measured and the largest of the boulders present in unit 2 were measured. This was done by measuring the intermediate axis of the boulder if it was well exposed. If the boulder was covered, the longest exposed axis was measured. The sample area for each station consisted of those boulders that appeared to be in situ, 25 m on either side of the established control point on the river bank.

A total of 225 boulders were sampled. Consistent with the sampling method employed by Bradley and Mears (1980), a mean axial dimension was calculated for the coarsest 25 % of all boulders sampled, which resulted in a mean boulder size of 104 cm. This was the grain size used in the calculations that follow.

Flood velocity

Four commonly applied engineering formulas calculating water velocity necessary to initiate the movement of bedload were employed in this study to calculate mean velocity. Three of the techniques yield a bottom velocity. For the purposes of this report mean velocities will be discussed. Mean velocity is considered to be at a height above the bed equal to 0.4 times the depth. As long as flow depth is several times particle size, a conversion factor of 1.4 can be used to convert bottom velocities to mean velocities (Bradley and Mears, 1980).

The four methods are as follows:

1) Mavis and Laushey (1949)

$$V_b = 0.5\ D^{4/9}(S-1)^{1/2} \quad (1)$$

where: V_b = bottom velocity in ft/sec,

D = grain size in mm

S = specific gravity of particles (2.5 assumed)

2) Peterka et al. (1956)

$$V_b = 2.57\ D^{1/2} \quad (2)$$

where: V_b = bottom velocity in ft/sec,

D = grain size in inches

3) Torpen (1956) A

$$D = 0.6\ V_b^2 \quad (3)$$

4) Torpen (1956) B

$$D = 0.08\ V^2 \quad (4)$$

where: D = grain size in inches

V_b = bottom velocity for sliding in ft/sec

V = mean velocity for rolling in ft/sec

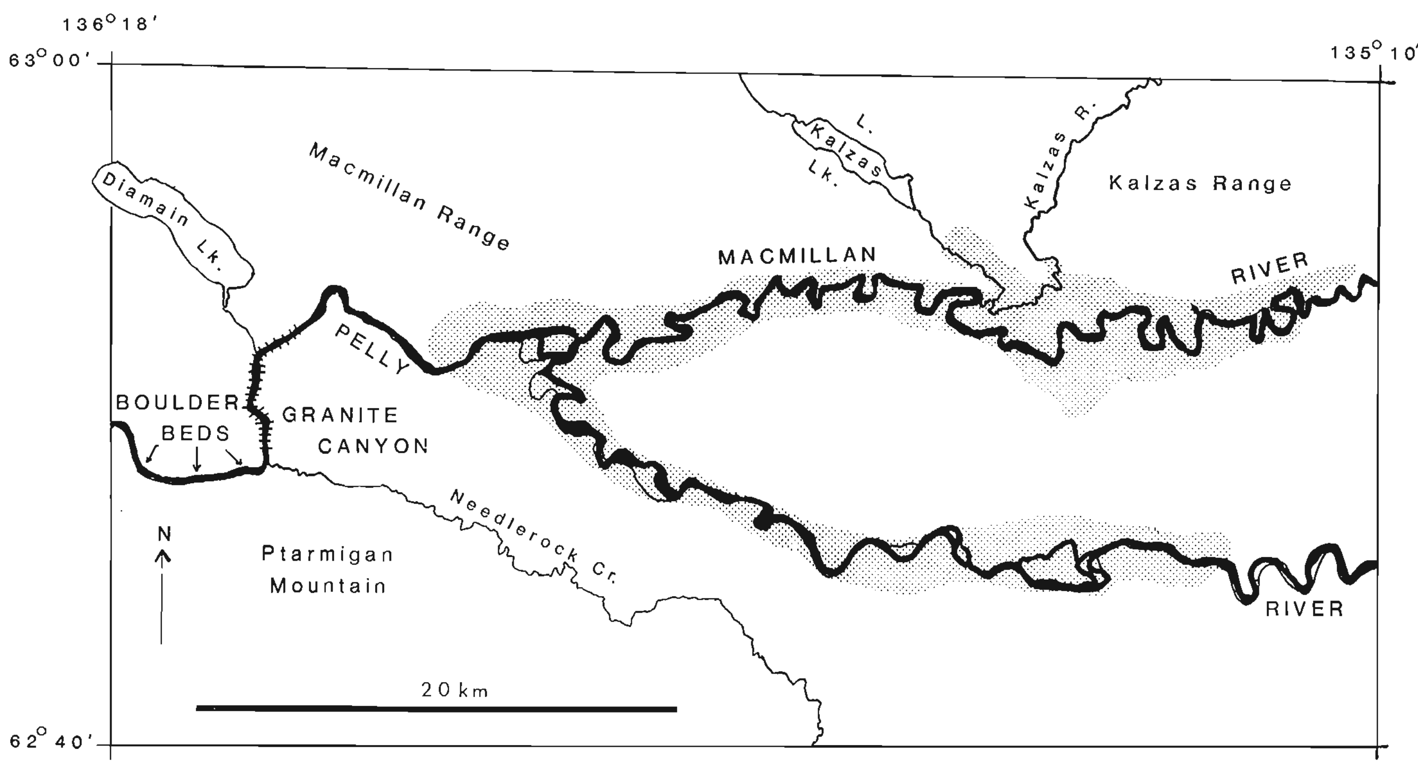

Figure 5. Extent of glacial lake deposits (shaded) upstream from Granite Canyon.

Table 1. Velocity calculation

Techniques	Mean velocity
1) Mavis and Laushey	5.71 m/s
2) Peterka et al.	7.02 m/s
3) Torpen A	3.52 m/s
4) Torpen B	6.89 m/s

Table 1 presents the results of these velocity calculations.

RESULTS AND DISCUSSION

Techniques 1, 2, and 4 (Table 1) yielded comparable values. The value calculated using Method 3, Torpen A, (3.52 m/s) is considerably lower than the other three. This is to be expected because this method calculates the velocity necessary to initiate sliding motion, whereas the other methods calculate the velocity necessary for initiating rolling motion.

The velocity values in Table 1 are consistent with other Pleistocene catastrophic floods (see Novak, 1973 for a summary) and in particular fall within the range of velocities computed for the Truckee River jökulhlaup using the Manning equation for open channel flow (Birkeland, 1968).

SUMMARY

Extensive boulder beds exposed along Pelly River below Granite Canyon are the product of a jökulhlaup which occurred during the deglaciation phase of McConnell Glaciation. This event likely occurred when a large glacial lake drained through Granite Canyon. Mean flow velocities in the 3.5 to 7 m/s range were required to initiate transport of the large boulders found in the boulder beds. This velocity range can be regarded as a minimum estimate of mean flow velocity during the jökulhlaup. This velocity range is also consistent with flow velocities calculated for other catastrophic floods.

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Magnetostratigraphy of early to middle Pliestocene basalts and sediments, Fort Selkirk area, Yukon Territory

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Abstract

Paleomagnetic study has shown that the Selkirk Volcanics and interstratified glacial and nonglacial sediments are excellent recorders of the paleofield. Comparison with the time scale of reversals of the paleofield shows that the Selkirk Volcanics in the vicinity of Fort Selkirk were laid down during the Matuyama Reversed Polarity Chron. The oldest pre-Reid Glaciation occurred prior to the start of the Olduvai Normal Polarity Subchron (> 1.87 Ma). The youngest pre-Reid Glaciation occurred in the post-Olduvai part of the Matuyama (0.79-1.67 Ma). The thick fill of glacial sediments in the adjacent Pelly River valley was deposited entirely within the Brunhes Normal Polarity Chron.

Résumé

Une étude paléomagnétique a montré que les roches volcaniques de Selkirk et des sédiments glaciaires et non glaciaires interstratifiés sont d'excellents enregistreurs du paléochamp magnétique. La comparaison avec l'échelle des temps des inversions du paléochamp montre que les roches volcaniques de Selkirk, au voisinage de Fort Selkirk, ont été mises en place au cours de l'époque de Matuyama de polarité inverse. La glaciation pré-Reid la plus ancienne s'est produite avant le début de l'événement d'Olduwai de polarité normale (> 1,87 Ma). La glaciation pré-Reid la plus récente s'est produite dans la partie post-Olduwai de l'époque de Matuyama (0,79 à 1,67 Ma). Le remplissage épais de sédiments glaciaires de la vallée de la rivière Pelly adjacente a été déposé entièrement au cours de l'époque de Brunhes de polarité normale.

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INTRODUCTION

Jackson (1989) presented evidence documenting a past eruption of a small volcano beneath an ice sheet during a pre-Reid Glaciation near the ghost town of Fort Selkirk, Yukon (Fig. 1). This edifice, informally called "Ne Ch'e Ddhäwa", its Northern Tutchone name, is a complex pile of altered (predominantly to analcite) hyaloclastite tuffs, breccias, and pillow breccias, with partly fused and unaltered lapilli beds and pillow basalts near its summit. He found Ne Ch'e Ddhäwa was partly underlain by sediments capped by a till. These sediments were apparently in fault contact with still older basalt flows of the Selkirk Volcanics (Bostock, 1936) that fill much of the Yukon River valley in the Fort Selkirk area.

This past interaction of volcanic activity and glacial ice provides an opportunity to date a previously undated glaciation using paleomagnetism as a tool. Field and laboratory work to these ends was carried out during 1989. In addition to paleomagnetic sampling at Ne Ch'e Ddhäwa, another key stratigraphic sequence within the Selkirk Volcanics and a nearby valley to Pelly River were sampled in order to further clarify Quaternary stratigraphy in the Fort Selkirk area. This report presents the results of this investigation.

SETTING

The Fort Selkirk area is located within the Yukon Plateau physiographic province (Bostock, 1948) in the vicinity of the confluence of Yukon and Pelly rivers (Fig. 1). The area is hilly upland, with local summits rising to 950 m elevation, cut by Yukon and Pelly river valleys and smaller tributary valleys such as that of Wolverine Creek. Local relief is about 500 m. The valleys contain extensive fills of basalt flows (Selkirk Volcanics). The contemporary re-excavated valleys have pronounced box or "V"-shaped cross sections. The basaltic fill is commonly in excess of 100 m thickness. Two major volcanic edifices, Volcano Mountain and Ne Ch'e Ddhäwa rise above basaltic valley fill.

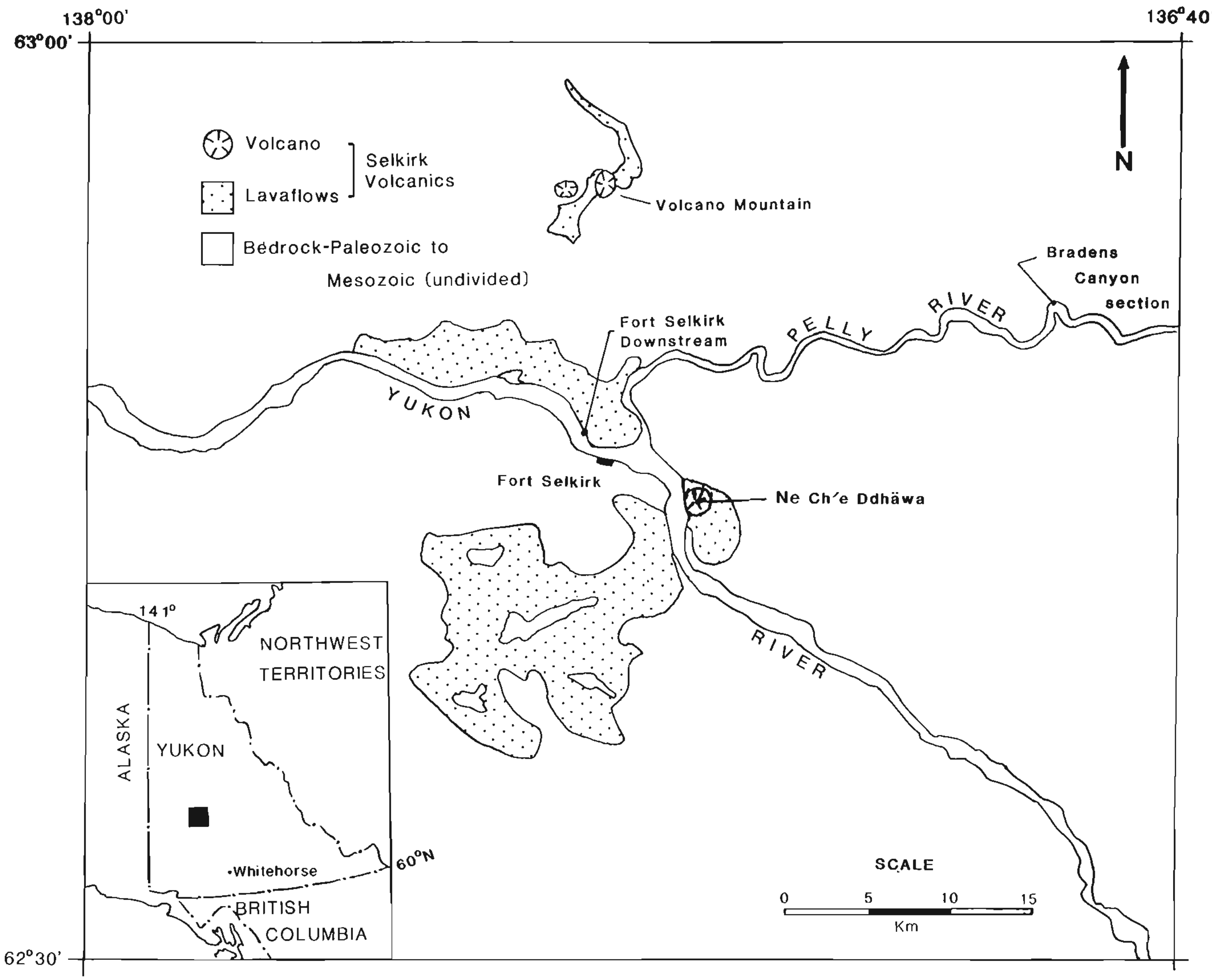

Figure 1. Location map showing Ne Ch'e Ddhäwa, Fort Selkirk Downstream and Bradens Canyon study sites. Stipple pattern indicates the Selkirk Volcanics (after Tempelman-Kluit (1984).

The area is densely forested or covered by bog. Exposures are confined primarily to cliffs along valley bottoms and to ridge tops.

PREVIOUS WORK

Geological investigations

Bostock (1936) assigned the name "Selkirk Volcanics" to late Tertiary to Holocene extensive basalt flows and volcanic edifaces in the Fort Selkirk area. He identified three major volcanic centres and sources for the basalt flows: upper Wolverine Creek, Ne Ch'e Ddhäwa (described by him as a unnamed "roughly conical hill") and Volcano Mountain. Owen (1959 a, b) and Bostock (1966) noted the presence of glacial and nonglacial sediments beneath approximately 100 m of basalt flows across and immediately down Yukon River from Fort Selkirk (Fort Selkirk downstream in Fig. 2). These sediments were described by Naeser *et al.* (1982). They dated a tephra (Fort Selkirk Tephra) enclosed in the sediments at 0.86± 0.13 Ma to 0.94± 0.40 Ma by fission track dating. The lowest overlying basalt yielded a whole rock K-Ar date of 1.08 + 0.05 Ma. Redating of the basalt has indicated it to be in the 1.5 Ma range (J.A. Westgate, personal communication, 1988).

Bostock (1966) and Hughes *et al.* (1969) delineated the limits of past glaciations in the Fort Selkirk area; they show the Fort Selkirk area 60 km beyond the limits of the McConnell Glaciation, at the limit of the penultimate Reid Glaciation, but well within the limit of pre-Reid glaciations. Bostock (1966) attributed striations on the surface of basalt flows near Fort Selkirk to the younger of two pre-Reid glaciations which he named "Klaza". He attributed the till underlying the basalts to the older pre-Reid glaciation which he named "Nansen". Sinclair *et al.* (1978) noted that the till on Ne Ch'e Ddhäwa was restricted to its base and was, by inference, related to a glaciation that postdated the last eruption of the mountain.

Paleomagnetism

The first paleomagnetic work on Quaternary deposits in the Yukon Territory was carried out by Du Bois (1959) who studied rocks from various localities, including basalts from near Fort Selkirk. Normal and reversed polarities were found and the mean direction, irrespective of sign, agreed well with the geocentric axial dipole. This early study was undertaken not for stratigraphic purposes, but as part of the global effort which was being carried out at that time to establish the general form of the paleofield. Naesar *et al.* (1982) reported reversed magnetization in Selkirk basalt, which can therefore be assigned to the Matuyama Reversed Polarity Chron in agreement with the whole rock K-Ar dates discussed above.

STRATIGRAPHY

The generalized stratigraphy of the three areas within which paleomagnetic sampling was carried out is presented in Figure 2. These three areas are refered to as Fort Selkirk down-

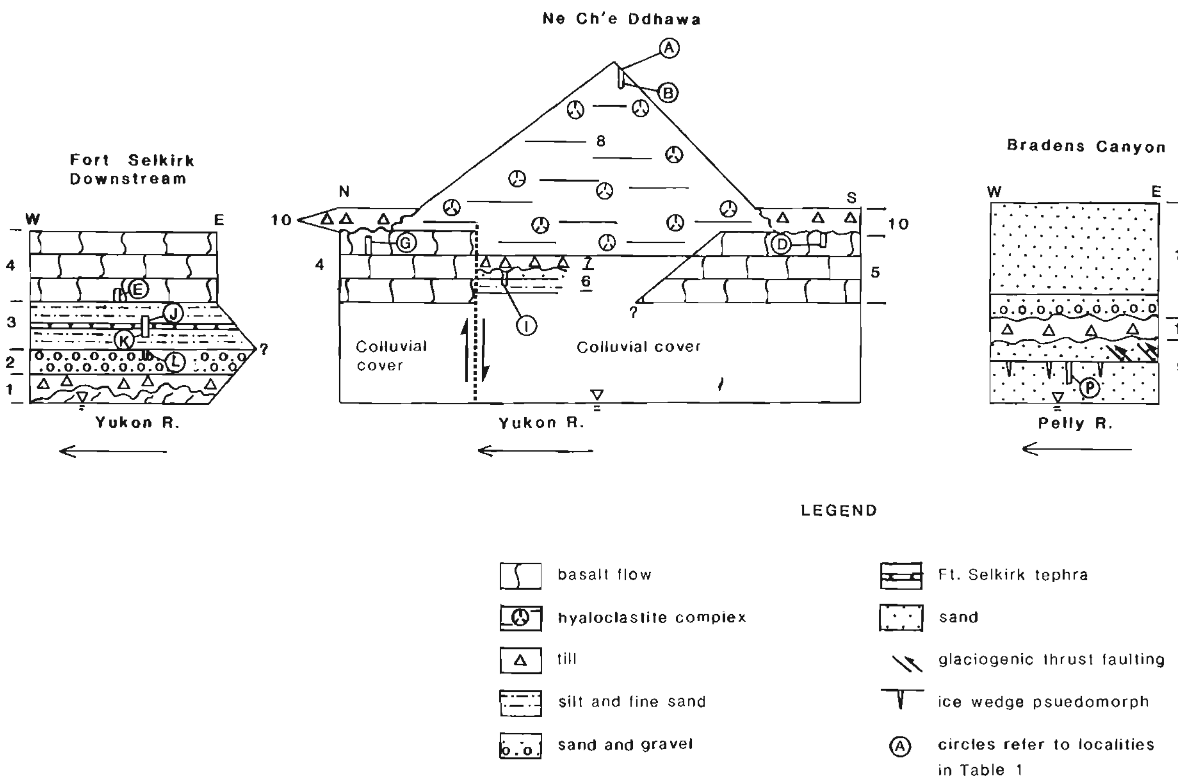

Figure 2. Schematic cross-section (not to scale; unit thicknesses given in Table 1). Numbers refer to units described in Table 1 and 1.

stream, Ne Ch'e Ddhäwa, and Bradens Canyon (Fig. 1). Eleven generalized stratigraphic units or packages of units are defined. These were numbered sequentially from oldest to youngest and are described in Table 1. The problems investigated are now described.

Ne Ch'e Ddhäwa (sampling localities A,B,D,G,I)

Jackson (1989) showed the till (unit 10) ringing the base of Ne Ch'e Ddhäwa (unit 8) to be of Reid age and the subglacial eruption of Ne Ch'e Ddhäwa to have occurred during the most recent pre-Reid Glaciation. The object of sampling in the Ne Ch'e Ddhäwa area was to date the latest pre-Reid Glaciation with reference to the reversal time scale by determining the polarity of suitable rocks near the summit of the mountain. It was also of interest to determine the age relationships between unit 8 and the underlying basalts and sediments.

Fort Selkirk downstream (sampling localities E,J,K,L)

Paleomagnetic sampling was carried out in order to clarify the age relationships: (1) between lava flows in unit 4 at this locality and units 4 and 5 at Ne Ch'e Ddhäwa, and (2) between sedimentary units 1-3 and 6 and 7.

Bradens Canyon (sampling locality P)

Field observations showed these sediments to be filling a valley excavated in unit 4 basalt flows. Paleomagnetic sampling was carried out at this site in order to determine if units 9 and 10 were pre-Reid or younger and to test the correlation of unit 10 here with unit 10 at Ne Ch'e Ddhäwa.

SAMPLING AND MEASUREMENT

Oriented samples were obtained from basalts and sediments. Basalts were cored using a small gasoline powered drill. Two cylindrical specimens were cut from each core. The sediment samples were obtained by pushing plastic cubes into a vertical face which had been previously cleaned and shaped by shovel and knife; 69 cores and 116 cubes were collected in June 1989.

Measurements were made on Schonstedt spinner magnetometers to which are attached devices (developed by Robotic Systems International of Sidney, British Columbia) that automate specimen manipulation and data acquisition. Batches of 15 are measured unattended. The system is controlled by IBM PC. Stepwise alternating field (AF) demagnetization was carried out using a Schonstedt GSC-5 for fields up to 100 mT, and Sapphire Instrument's SI-4 for fields up to 180 mT. Demagnetization was carried out by treatment along three axes successively. With the exception of the basalts, whose magnetization is simple, and of some C-type specimens (explained below), all have been treated in at least 6 sequential demagnetization steps.

MAGNETIZATION TYPES

The basalts are excellent recorders of the paleofield. A typical example is shown in Figure 3, which is a single component magnetization of very high stability. Paleodirections have been obtained by AF cleaning at 40 mT. This simple blanket cleaning is justified because the basalts produce excellent linear decay lines that pass through the origin on orthogonal plots and end points are well grouped. All basalts have reversed polarity. Such "ideal" magnetizations are referred to as A-type. All basalt samples have A-type magnetization.

Most of the sediments collected have A-type magnetization. Examples from different lithologies are shown in Figures 4 to 7. The eolian silts (sample locality J in unit 3) and the underlying fine sands (sample locality K in unit 3; Fig. 4 and 5) have A-type magnetizations, very similar in character to that of the basalt (Fig. 3), but with intensities two, or more commonly, three orders of magnitude less. They yield excellent linear decay lines, and end points. The eolian silts have normal polarity whereas the underlying sands have reversed polarity. Paleodirections in A-type sediments have been determined by least-squared fitting to decay lines within the limits 10 to 150 mT. Paleodirections and polarities are clearly defined.

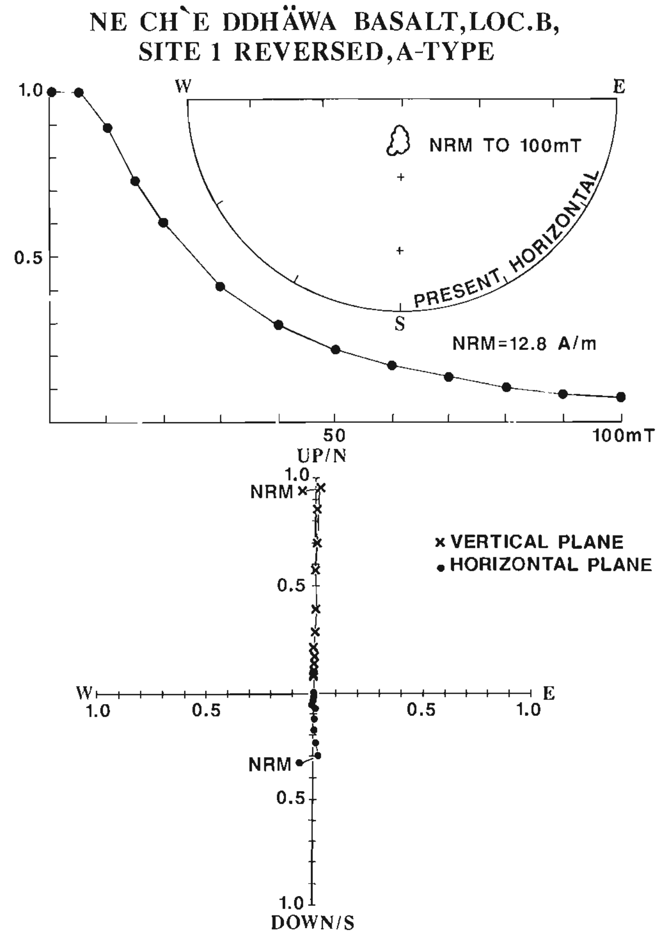

Figure 3. Reversed polarity, A-type magnetization. Ne Ch'e Ddhäwa basalt, locality B, site 1. Direction (top) and normalized intensities (middle) during alternating field demagnetization. Below is an orthogonal plot on the north-south vertical plane (crosses) and the horizontal plane (dots). Inclinations upward. NRM = normal remanent magnetization; Up/N

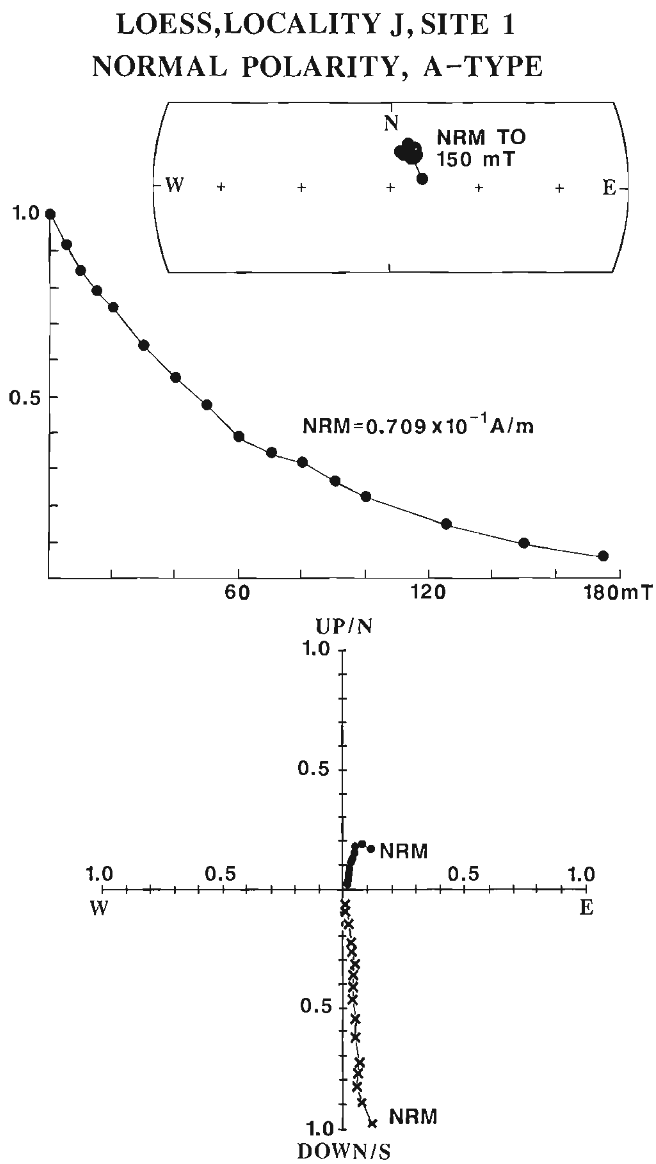

Figure 4. Normal polarity, A-type magnetization. Surface loess, Fort Selkirk tephra site, locality J, site 1. Inclinations downward.

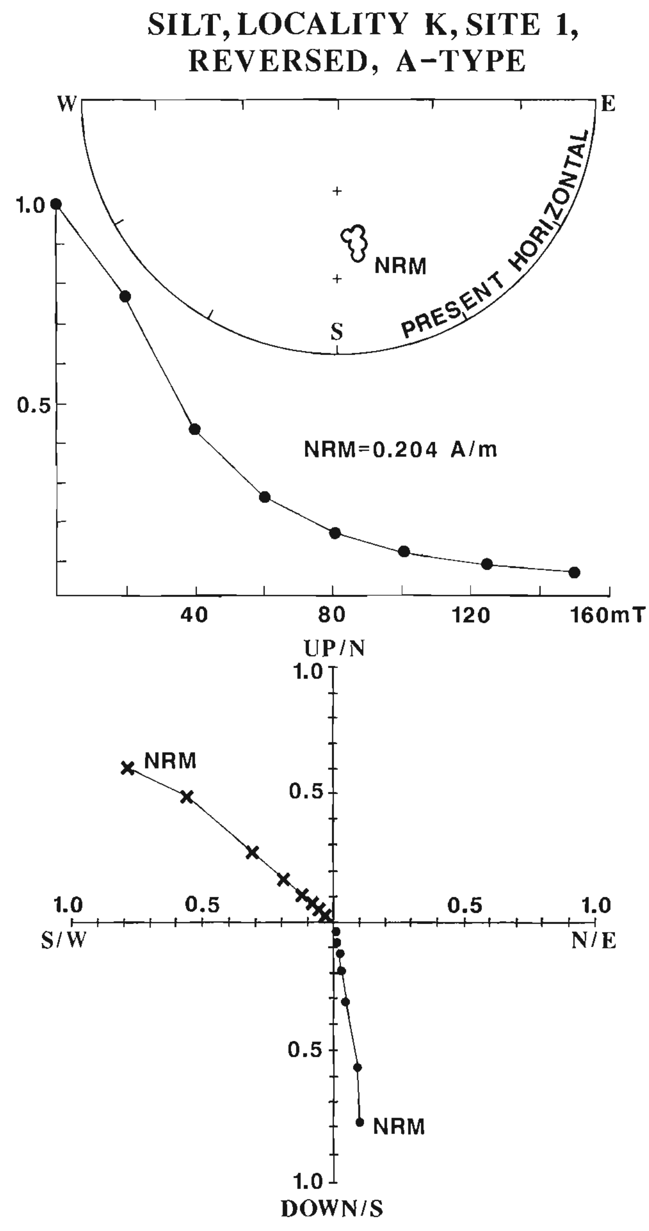

Figure 5. Reversed polarity, A-type magnetization. Silts and sands directly above Fort Selkirk tephra site, locality K, site 1. Inclinations upward.

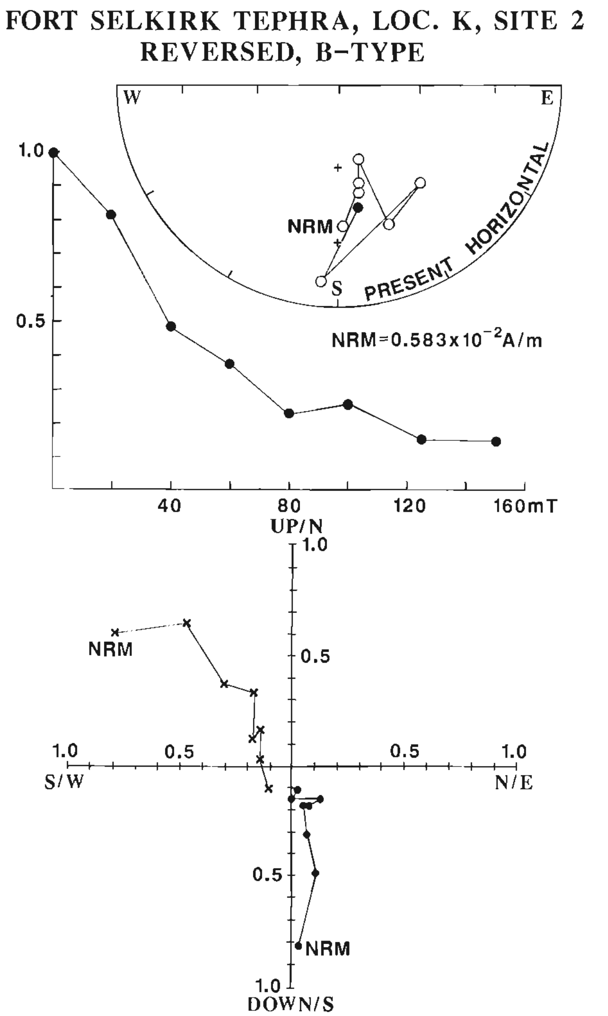

Figure 6. Reversed polarity, B-type magnetization. Fort Selkirk tephra, locality K, site 2. Inclinations upward except 175 mT point.

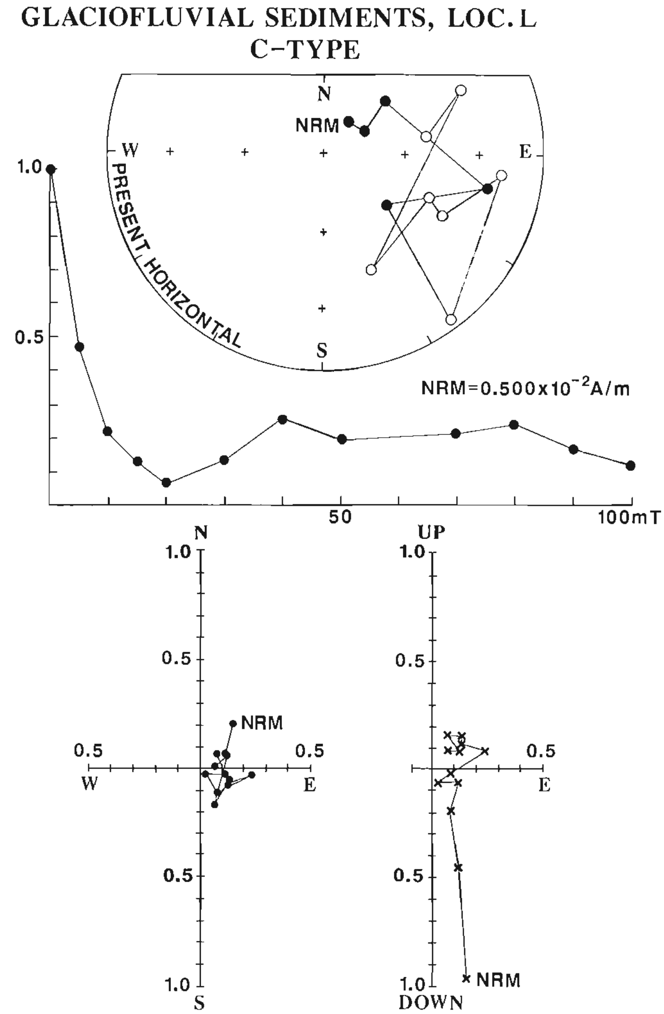

Figure 7. C-type magnetization. Glaciofluvial sediments directly above older pre-Reid till, Fort Selkirk tephra site, locality K, site 4. Magnetizations uninterpretable.

Some sediment samples (13 %), yield less precise data, referred to as B-type; an example is given in Figure 6. The decay lines are poor and, although a rough grouping of directions is retained during demagnetization, there is no "end point" and directions in high fields are commonly scattered. In this example polarity is evidently reversed, but the paleodirection cannot be accurately estimated. Hence B-type magnetizations have been used to determine polarity but they have not been used to estimate paleodirections.

A third category of data, C-type, was found in 24 % of the samples. An example from glaciofluvial sediments (sample locality L in unit 2) is illustrated in Figure 7. The directions are scattered during demagnetization, and no decay lines can be identified. Such data define neither paleodirection nor polarity and are discarded.

Most of the C-type magnetization was likely caused by extreme disturbance of sediment samples during emplacement of sampling cubes around them. Sampling disturbance was particularly a problem at locality L where medium to coarse sands were sampled; all fifteen samples yielded C-type magnetization. B-type magnetization may have a small sampling disturbance component but circumstantial evidence links it to geological processes. For example, sediments at locality I (10-A's, 6-B's, 3-C's) have been deformed by glacial tectonism; sediments at locality P (15-A's, 5-B's) have experienced glacial tectonism and periglacial disturbance.

SUMMARY OF PALEOMAGNETIC DATA

The paleomagnetic results are compiled by sites and by stratigraphic unit in Table 2. The directional results in Table 2 have been obtained using A-type data only. The mean paleodirections of rock units are plotted in Figure 8 and their polarities noted on the stratigraphic section shown in Figure 9.

The means for unit 8 (localities A and B), unit 5 (locality G) and unit 4 (locality E) at Fort Selkirk downstream are essentially in perfect agreement, differing (2°), by far less than statistical error. It is therefore, highly probable, based on paleomagnetic data alone, that these three units were coeval, because the chances of sampling exactly the same paleofield at three separate times are remote. The mean of unit 4 at Ne Ch'e Ddhäwa (locality G) is 9° from the mean of paleodirections at localities A, B, D,and E (entry F in Table 2) which exceeds the sum of the 95 % error, indicating that they are of significantly different age. As expected from stratigraphic relationships, the paleodirections at other localities (M,J,I) differ significantly from one another, indicating they have differing ages.

The data, although from only a few localities, display normal and reversed groups 180° apart and the overall mean, irrespective of sign (N), agrees very well with the geocentric axial dipole (GAD) field direction.

The paleodirections from locality I in unit 6 are exceptional; relative to present horizontal, they are to the northeast, somewhat outside the range that might be expected for paleosecular variation. When corrected to bedding, the paleodirections become widely different. There is no unequivocal explanation at present for this unusual result, but one possible sequence of events is that the sediments were postdepositionally remagnetized (presumably by grain rotation) and after the magnetization became stabilized. Thus the sediments here were rotated some 30° or so clockwise.

This sequence of events is compatible with the depositional history of unit 6. The sands, silts, and clays of this unit were loaded by glacial ice. The loading pressed boulders into these fine sediments causing diapiric deformation. This deformation indicates rates of loading and water contents were sufficient to cause conditions of very low effective stress within these sediments and thus there was an opportunity for grain reorientation and remagnetization. Subsequent 30° rotation of the sampled interval could have been accomplished by thrust faulting under the impetus of the overlying and flowing glacial ice. This subsequent rotation cannot be demonstrated at present because the present exposure is confined to a 1 m wide trench; a much larger exposure would be required in order to detail larger scale structural features within this unit.

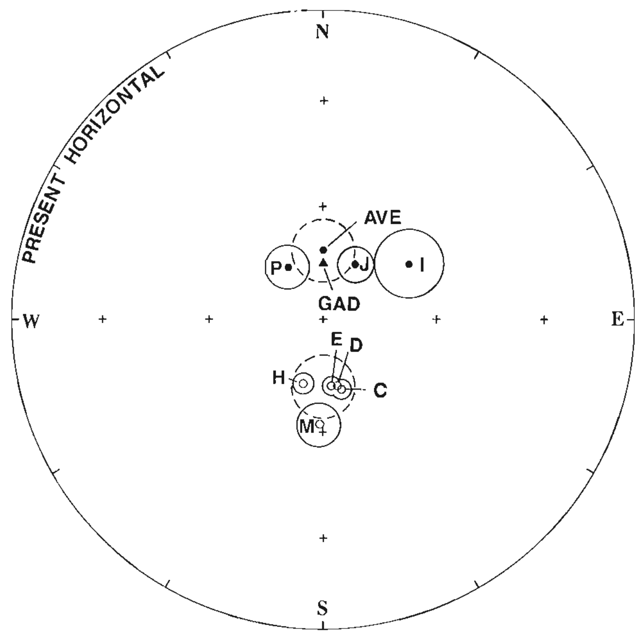

Figure 8. Summary of directions of magnetization by locality plotted on a Wulff stereonet. Open (closed) symbols, negative (positive) inclination. Letters refer to localities in Table 2. GAD is the direction of the geocentric axial dipole field. AVE is the summary direction of Table 2.

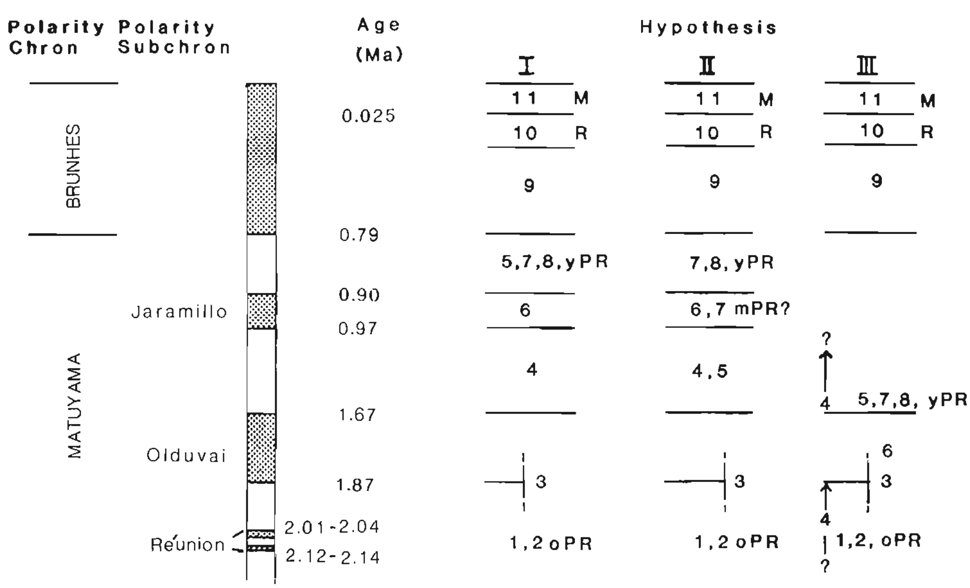

Figure 9. Polarity timescale after Mankinen and Dalrymple (1979). Hypotheses I, II, and III refer to the possible interpretations of Fort Selkirk polarity data.

Table 1. Stratigraphic units shown in Table 2

Unit number	Thickness (m)	Sample locality	Sites sampled	Description
1	1.5-2	—	—	Indurated stony sandy diamicton. Stones 25-30% fresh vesicular amygdaloidal basalt. Basalt clasts reach 1 m in maximum dimension; non-basalt clasts reach 10 cm diameter. The contact with underlying greenstone is erosional. Unit capped by a coarse erosional lag of basalt boulders.
2	12	L	1	Glaciofluvial gravels grading upward to interstratified sands and gravels with fine sandy stringers near the top. Samples were obtained from these stringers for paleomagnetic analysis. Gravels have a lower content of basalt and a higher content of plutonic and metamorphic clasts than unit 1.
3	4	J,K	4	Upper 1.3 m of this unit is composed of massive eolian silt. The silt contains scattered rodent teeth and rodent and large vertebrate bones. This silt sequence was called site J for paleomagnetic sampling purposes. This grades downward into 60 cm of thinly bedded fine to medium sands which locally contain coarse sand stringers. These overlie 25 cm of thinly bedded fine sand with the upper 2 cm rich in organic silt particles. The underlying 23 cm contains lamellae of reworked Fort Selkirk Tephra. The 20 cm thick underlying Fort Selkirk Tephra is bounded by abrupt upper and lower contacts and is underlain by 225 cm of massive fine to medium sand. The sediments below site J were refered to as site K for paleomagnetic sampling.
4	70-150	E,G	2	Flows of vesicular amygdaloidal basalt downstream of Ne Ch'e Ddhäwa. At the Fort Selkirk downstream locality (Fig. 2) the base of the flows rest on about 1 m of basaltic lapilli containing charred wood which in turn overlies unit 3. Individual flow thicknesses range from 6 m to 60 m (Bostock, 1936). At Ne Ch'e Ddhäwa, the bases of the flows are buried by colluvium. Exposed flows are as thin as 3 or 4 m with columns ranging from horizontal to vertical in orientation. Flows are separated by hyaloclastite breccia and pillows. Paleomagnetic samples at locality E were taken from massive basalt immedialely above the basal lapilli. Samples at locality G were taken from a columnar basalt 2 m below the top of the uppermost flow
5	>50	D	1	Columnar vesicular amygdaloidal basalt consisting of more than three cooling units (lower half of the flows covered by colluvium). This unit thins in a downstream direction and apparently had an eruptive source to the southwest in Wolverine Creek valley. Sampling locality N was 1 m below the top of the uppermost flow.
6[1]	>3	I	1	More than 2 m of semilithified rusty sand overlain by alternating beds of silty sand and clayey silt. The beds have been disturbed by glacial tectonism. The upper 75 cm of the unit has been highly deformed by the penetration from above of 20 to 50 cm blocks of basalt, some of which are intensely striated. Deformation in unit 6 increases toward the base of overlying unit 7. Sampling locality I is situated 2 m below the top of unit 6 in banded clay and fine sands with intraclasts. Unit 6 sediments yield a pollen spectra reflecting a vegetation community similar to extant vegetation.
7[1]	0.5	—	—	Dark grey, stony diamicton. Matrix is sandy loam in texture. Stones are commonly flat-iron shape and striated. Its contact with underlying unit 6 is highly sheared. The contact with the overlying hyaloclastites is abrupt and apparently depositional.
8	600	A,B,	5	Hyaloclastite complex forming Ne Ch'e Ddhäwa (see Jackson, 1989, for a detailed description). Sampling localities A and B are zones of fused lapilli/tuff breccia beds and pillow basalts, respectively, near the summit (Jackson, 1989, p. 254-255).
9	>5m	P	1	Unit 9 corresponds to units 1 and 2 described by Ward (1989) and are the lowest sediments exposed along Pelly River at the base of the Bradens Canyon section. These stratified sands contain peaty organic beds and ice-wedge casts. The upper part of this unit contains eastward dipping thrust faults caused by glacial overriding. Sample locality N was located in relatively undisturbed sandy silts about 2 m below the top of the unit.
10	<0.5-12	—	—	Stony diamicton. Matrix is sandy loam in texture. Stones are commonly flat-iron shaped and polished. Erosional contact with underlying rock. The underlying sediments are sheared or thrust-faulted.
11	30	—	—	Glaciofluvial valley fill: sands and gravels.

[1]Units 6 and 7 were uncovered through trenching. The thickness of unit 6 and the lateral extent of both units are unknown.

STRATIGRAPHIC INTERPRETATIONS

Based on paleomagnetic data and observed field relationships, three stratigraphic columns are suggested with respect to the paleomagnetic time scale (Fig. 9). Testing of these hypotheses must await further K-Ar dating and paleomagnetic investigations. However, all three have the following features and these should be regarded as conclusions:

1) The eruption of Ne Ch'e Ddhäwa beneath an ice sheet of the most recent pre-Reid glaciation occurred during the Matuyama Reversed Polarity Chron.

2) The sediments filling Pelly River valley at Bradens Canyon were deposited during the Brunhes Normal Polarity Chron. Consequently, unit 10, a glacial diamicton, was deposited during the Reid Glaciation because this locality lies within the Reid limit and all other glacial deposits within this area are pre-Reid.

3) K-Ar dates in the 1-1.5 Ma range for the base of unit 4 at the Fort Selkirk downstream site place the magnetic reversal recorded in unit 3 either near the beginning of the Olduvai Normal Polarity Subchron or within an even older normally magnetized interval. This means that units 1 and 2, till and outwash deposited during the oldest pre-Reid Glaciation, are older than 1.87 Ma.

Hypothesis I

This interpretation assumes:

1) Units 5 and 8 essentially are coeval as suggested by paleomagnetic evidence and supported by the conformable contact between them.

2) Unit 4 was essentially erupted over the same time period at Fort Selkirk downstream and Ne Ch'e Ddhäwa, as indicated by stratigraphic continuity. The coincidence of paleodirections from locality E with those from A,B, and D is assumed to be a chance occurrence.

3) The eruption of units, 5 and 8 and the deposition of unit 7 occurred during the same glaciation.

This being the case, the eruptions of units 5 and 8 and the younger pre-Reid Glaciation took place in the interval between the Brunhes/Matuyama boundary and the end of the Jaramillo Normal Polarity Subchron. Unit 6 was deposited during the Jaramillo Normal Polarity Subchron.

Table 2. Paleomagnetic results by collection Site and by geological unit.

Locality	Site	*n*	*M*	*D, I*	*k*	α95	Polarity	*n/Mag-Type*
A. (Unit 8 tuff breccia)	1	5	19.0	179,-72	1162	2	R	5-A
B. (Unit 8 pillow basalt)	1	3	19.0	174,-71	66	15	R	3-A
	2	2	4.0	155,-77	276	15	R	2-A
	3	12	13.0	166,-70	64	10	R	12-A
	4	6	6.0	168,-72	419	6	R	6-A
C. (Unit 8 combined)	-	28	12.7	168,-72	109	3	R	
D. (Unit 5)	1	11	1.3	166,-71	512	2	R	11-A
E. (Unit 4: Fort Selkirk downstream)	1	14	2.0	172,-72	186	3	R	14-A
F. (C, D, E, combined)	-	3	5.3	169,-72	5220	1.7	R	
G. (Unit 4: Ne Ch'e Ddhäwa)	1	8	1.0	197,-73	175	4	R	8-A
	2	8	0.7	196,-71	269	3	R	8-A
H. (Unit 4: Ne Ch'e Ddhäwa, combined sites)	-	16	1.3	197,-72	217	3	R	
I. (Unit 6)	1	10	0.004	57, 63	29	9	N	10-A /6-B/3-C
J. (Unit 3: eolian silts)	1	11	0.047	30, 73	74	5	N	11-A
K. (Unit 3: sand/tephra sequence)								
— sands above tephra	1	12	0.059	173,-56	75	5	R	12-A/1-B
— tephra	2	5	0.006	192,-67	34	13	R	5-A/1-B
— oxidized sands below tephra	3	7	0.017	198,-66	20	14	R	7-A/5-C
L. Fort Selkirk glaciofluvial silts and sands	1	15	0.004	-	-	-	unresolved	15-C
M. (Unit 3: sand/tephra sequence combined)	-	24	0.036	182,-62	30	6	R	
P. (Unit 9: subtill fine sands)	5	0.12		325, 73	44	6	N	15-A/5-B
SUMMARY (F,H,J, and K combined)	-	5	1.1	1, 72	80	8.6		

n the number of samples
M the mean intensity of natural remanent magnetization (A/m)
D,I the declination and inclination of mean direction
k precision (Fisher, 1953)
α95 radius of circle of 95° confidence (p= 0.05)
N,R normal and reversed polarity
n/Mag-type number of specimens and magnetic type as explained in text

Hypothesis II

The same as for hypothesis I except that units 6 and 7 are assumed to be coeval and not related to the younger pre-Reid Glaciation. This being the case, these units could have been deposited during a middle pre-Reid Glaciation.

Hypothesis III

This interpretation assumes:
1) Units 5 and 8 at Ne Ch'e Ddhäwa are coeval with the base of unit 4 at Fort Selkirk downstream as suggested by the paleomagnetic directional data.
2) The eruption of unit 8 and the deposition of unit 7 occurred during the same glaciation.

This being the case, unit 4 was erupted over at least 0.2 Ma beginning prior to the Olduvai Normal Polarity Subchron: lava flows at Ne Ch'e Ddhäwa were erupted prior to the Olduvai Subchron whereas those at Fort Selkirk downstream postdate the Olduvai. Normally magnetized unit 6 was deposited during the Olduvai Subchron. Eruption of Ne Ch'e Ddhäwa and the younger pre-Reid Glaciation occurred between 1.5 Ma and the end of the Olduvai Subchron at 1.67 Ma.

A fourth hypothesis could be constructed in the manner of Hypothesis II in order to illustrate the possiblity that a middle pre-Reid occurred.

CONCLUSIONS

The Selkirk Volcanics, with their interstratified glacial and nonglacial sediments, span much of the Matuyama Reversed Polarity Chron. The youngest pre-Reid Glaciation occurred between the end of the Matuyama Chron and the end of the Olduvai Normal Polarity Subchron; precise radiometric dating will be required to resolve age more accurately. The older pre-Reid glaciation occurred prior to the Olduvai Normal Polarity Subchron and there is a possibility that a third pre-Reid glaciation of intermediate age occurred. The sedimentary fill within Pelly River valley in the area of Bradens Canyon is normally magnetized indicating Brunhes age. The glacial diamicton and thrust faults exposed in this section are products of the Reid Glaciation.

ACKNOWLEDGMENTS

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Paleomagnetism of the Old Crow Batholith, northern Yukon Territory

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Park, J.K. Paleomagnetism of the Old Crow Batholith, northern Yukon Territory; in Current Research, Part E, Geological Survey of Canada, Paper 90-1E, p. 287-290, 1990.

Abstract

A paleomagnetic remanence that corresponds roughly to the Cretaceous-early Tertiary reference field for cratonic North America has been obtained from the Old Crow granite batholith in the northern Yukon. This component is interpreted as an overprint acquired during orogeny, and hence could not be used to establish whether the region rotated relative to the craton between Late Triassic and Early Cretaceous time.

Résumé

Une rémanence paléomagnétique, qui correspond grossièrement au champ de référence du Crétacé-Tertiaire inférieur pour l'Amérique du Nord cratonique, a été obtenue à partir du batholite granitique d'Old Crow situé dans la partie nord du Yukon. On interprète cette composante comme une surimpression acquise au cours d'une orogenèse et on ne pourrait pas, par conséquent, l'utiliser pour établir si la région a tourné par rapport au craton entre le Trias supérieur et le Crétacé inférieur.

INTRODUCTION

A paleomagnetic study of the Old Crow granite batholith was undertaken to determine whether the northern Yukon rotated counterclockwise between Late Triassic and Early Cretaceous time according to a scenario invoked for northern Alaska (Carey, 1958; Tailleur and Brosgé, 1970; Tailleur, 1973; Irving and Sweeney, 1982; Norris, 1984). Several earlier paleomagnetic studies in the Brooks Range of northern Alaska (Newman et al., 1977; Hillhouse and Grommé, 1983) failed to prove the rotation hypothesis, probably due to the presence of a post-deformational magnetic overprint of probable Cretaceous age, but another study, from the North Slope, revealed a stable magnetization consistent with the hypothesis (Halgedahl and Jarrard, 1986).

METHODS

Twelve sites were drilled in the field: eight from the Old Crow batholith and four from outlying stocks (1, Dave Lord; 12, Mt. Fitton; 11, Mt. Sedgwick; 10, Ammerman Mt.; Fig. 1). Sun and magnetic compasses were used to record core orientations. In the laboratory cores (generally 5/site) were sliced into 2 or 3 specimens of standard size. Specimens were treated thermally or by alternating fields (AF). Thermal treatment in steps up to 600°C was done in commercial (Schonstedt TSD-1) or "in-house" instruments (Roy et al., 1972); AF treatment up to 200 mT, in an "in-house" apparatus (Roy et al., 1973). Specimens were measured on a commercial spinner magnetometer (Geofyzika JR4).

GEOLOGY

The Old Crow batholith is centred about 30 km west of the town of Old Crow on the Alaska/Yukon border (Fig. 1). Several other granite plutons are located generally north of Old Crow up to 100 km distant. It is not known whether any of these bodies have been tilted or rotated since intrusion. The Old Crow batholith varies widely in composition, but it is generally of granitic composition. The Dave Lord pluton is a syenodiorite. Little else has been published about these bodies and their tectonic history. The Old Crow and Mt. Sedgwick bodies contain two generations of micas, and they and the Dave Lord pluton show sericitization of feldspars and chloritization of the biotite and amphibole (Bell, 1985). Microscopic deformation is evident in the Old Crow pluton from the kinking of biotite and the bending of twin lamellae in plagioclase (Judge, 1985).

Opaque minerals, very rare in the Old Crow body, consist of microscopic grains of magnetite, ilmenite, hematite, and possibly sulphides. In contrast, opaque minerals, especially magnetite, are abundant in the outlying stocks (sites 1, 10-12).

Radiometric K-Ar ages from the Old Crow and outlying plutons, ranging from 97-376 Ma (Bell, 1985; Judge, 1985), are equivocal. Rb-Sr studies, both mineral (Judge, 1985) and whole rock (Bell, 1985), of the Old Crow failed to produce a date but indicated that the body may be older than 364 Ma (Judge, 1985).

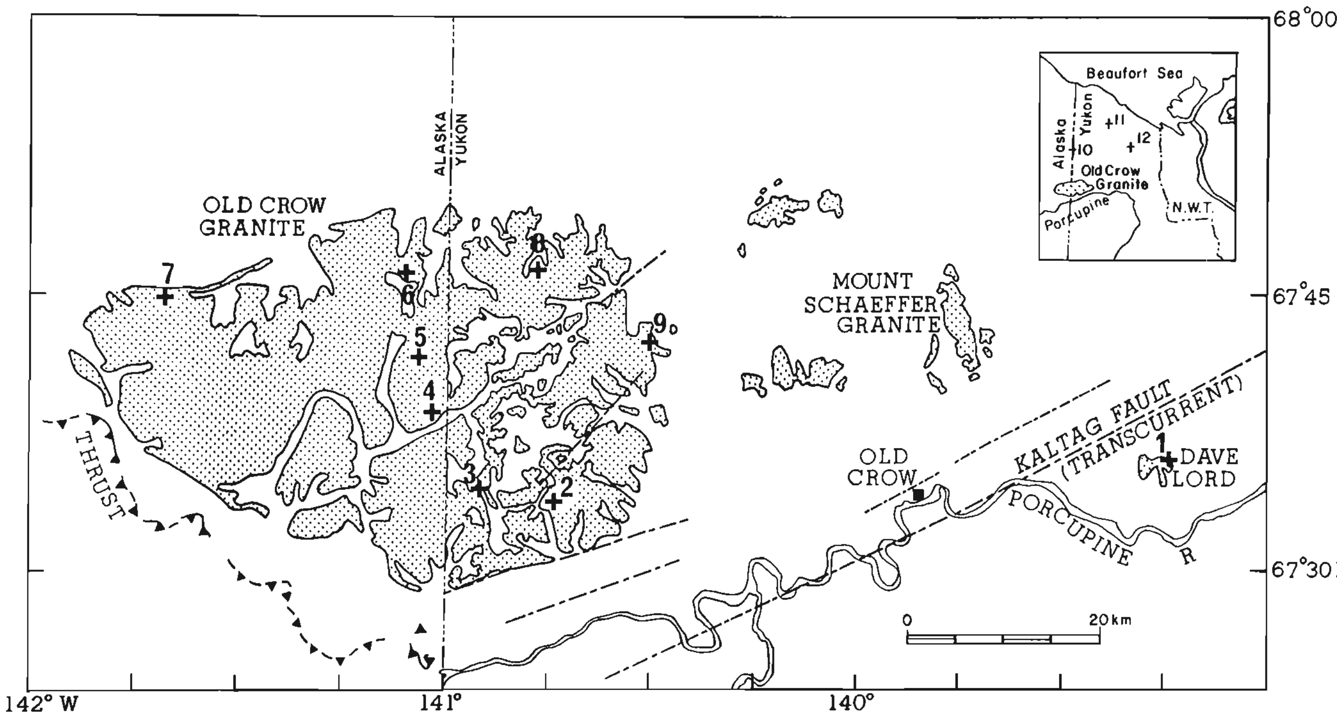

Figure 1. Geological map of the Old Crow batholith and outlying stocks (after Norris, 1981; Brosgé and Reiser, 1969). Sites are denoted by crosses.

PALEOMAGNETIC RESULTS

All specimens from the Old Crow batholith, except those from sites 3 and 9, have NRMs (natural remanent magnetizations) with low intensity (0.2 - 0.7 mA/m) and apparently random directions. AF and thermal treatment isolated two components from the NRM as well as other, usually less-stable components with scattered directions. One component, O_B, has steeply inclined negative and positive directions, unblocking temperatures (T_{UB}'s) below 450°C, and resistive coercive forces (rcfs) below 80 mT (Table 1). The other component O_A is the most stable, with T_{UB}'s up to 600°C and rcfs greater than 200 mT (Fig. 2b,2c; Table 1). It was only isolated from two cores at site 3, which also gave evidence of other less stable components (Fig. 2a), including O_B. Some sites (4, 6, 7, 9) yielded magnetic components with T_{UB}'s up to 520°C but with no coherent direction (Table 1).

Magnetizations in the outlying granite stocks are somewhat different from those in the Old Crow batholith. NRM directions are more coherent, tending to be steeply-inclined with positive polarity. Also, NRM intensities are two or three orders of magnitude higher (0.13 - 0.29 A/m), consistent with the greater abundance of magnetite. AF or thermal cleaning of sites 1, 10, and 11 removed a steeply-inclined component with T_{UB}'s below 250°C that may be related to either the present earth's field direction (PEF) or the field direction at the time of the Cretaceous-early Tertiary orogeny. The removed component at site 12 is somewhat different, with a direction in the southwest quadrant and higher T_{UB}'s (Table 1). A great circle passing through the directional trend of the site 12 distribution also passes through the PEF, indicating that the directions may be influenced by the PEF.

DISCUSSION

Thermal or chemical alteration associated with the Cretaceous-early Tertiary orogeny and with recent events has no doubt removed or masked much of the primary magnetization of the granite bodies. It has produced the steeply-directed O_B magnetization, especially evident in the stocks, that appears to be similar to the post-deformational magnetic overprint found in the previous paleomagnetic studies in the Brooks Range of Alaska (Newman et al., 1977; Hillhouse and Grommé, 1983). This direction from both the Old Crow and outlying bodies is consistent with a recent origin, but could equally well relate to the Cretaceous-early Tertiary orogeny (Table 1). Site 3 yielded a steep, negative O_B direction with a Cretaceous-like pole, significantly different from the PEF in the region. It also gave the very stable O_A magnetization, which yields a low-latitude pole. Because O_A is only found in five specimens from two cores its significance is unknown.

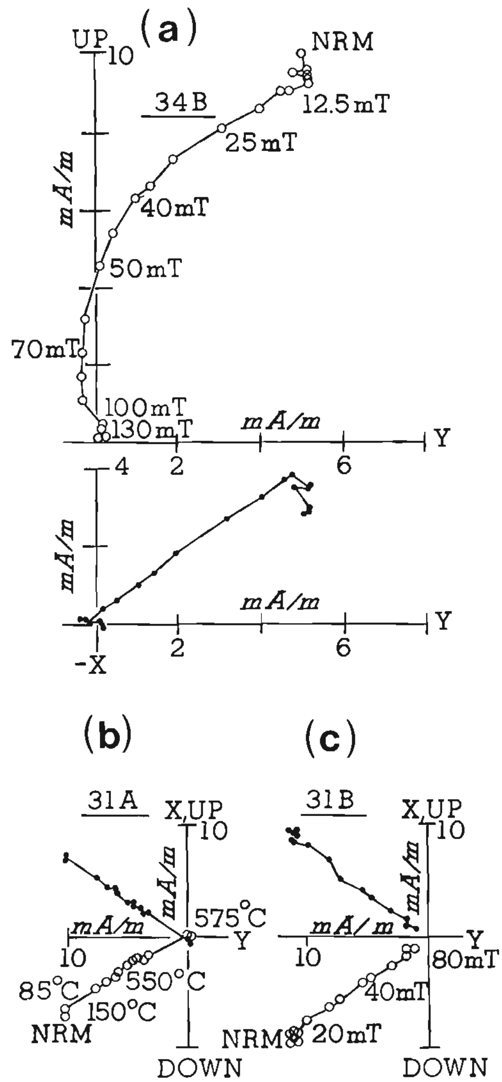

Figure 2. Vector diagrams showing examples of the demagnetization of specimens from site 3 under AF or thermal treatment. Solid symbols denote data points plotted on the horizontal X-Y plane; open symbols, data points plotted on the vertical UP-Y plane. The specimens of (b) and (c) are from the same core.

CONCLUSION

No magnetizations resolved in this study have been proven to predate the period of possible relative rotation of the northern Yukon. Factors that have precluded a positive result include the presence of a pervasive magnetic overprint of probable Cretaceous-early Tertiary age, the lack of age control on the Old Crow batholith, and the lack of structural controls.

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Gravity surveys in support of Lithoprobe Southern Cordillera Transect, Kamloops region, British Columbia

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Thomas, M.D., Halliday, D.W., Moore, J.M., and Grover, B. Gravity surveys in support of Lithoprobe Southern Cordillera Transect, Kamloops region, British Columbia; in Current Research, Part E, Geological Survey of Canada, Paper 90-1E, p. 291-295, 1990.

Abstract

LITHOPROBE-related gravity measurements 1 to 3 km apart were made along traverses in the Intermontane and Omineca Crystalline belts in summer 1989. A traverse along seismic reflection line 19 within the Omineca Crystalline Belt realized the objective of completing closely spaced measurements along all seismic lines. Other traverses investigated specific geological targets within the Intermontane Belt near Kamloops, in an area between line 19 and more southerly seismic lines. The new gravity data together with older regional data define a prominent -25 mGal amplitude anomaly over the largely granitic Central Nicola Horst. In contrast, traverses across the NE margin of the granitic Wild Horse Batholith reveal it to be associated with a weak gravity signature. Considering that nearby rocks of the Nicola Group may be thrust eastward over the Chapperon Group the gravity data suggest that the Wild Horse Batholith is a thin thrust wedge.

Résumé

Des mesures de la pesanteur prises 1 à 3 km l'une de l'autre, ont été effectuées suivant des cheminements dans les zones cristallines intramontagneuse dans le cadre du programme Lithoprobe et d'Omineca pendant l'été de 1989. Un cheminement situé le long du profil 19, dans la zone cristalline d'Omineca, a permis de réaliser l'objectif qui était de procéder à des mesures détaillées, c'est-à-dire très rapprochées, le long de tous les profils sismiques. D'autres cheminements ont permis d'étudier des cibles géologiques spécifiques au sein de la zone intramontagneuse près de Kamloops, dans une région située entre le profil 19 et des profils sismiques plus au sud. Les nouvelles données gravimétriques ainsi que des données régionales plus anciennes définissent une anomalie importante d'une amplitude de −25 mGal au-dessus du horst central de Nicola de nature surtout granitique. Par contre, des cheminements qui traversent la marge nord-est du batholite granitique de Wild Horse révèlent que ce dernier est associé à une faible signature gravimétrique. Si l'on tient compte du fait que les roches avoisinantes du groupe de Nicola peuvent constituer un chevauchement vers l'est au-dessus du groupe de Chapperon, les données gravimétriques semblent indiquer que le batholite de Wild Horse est constitué d'un mince prisme de charriage.

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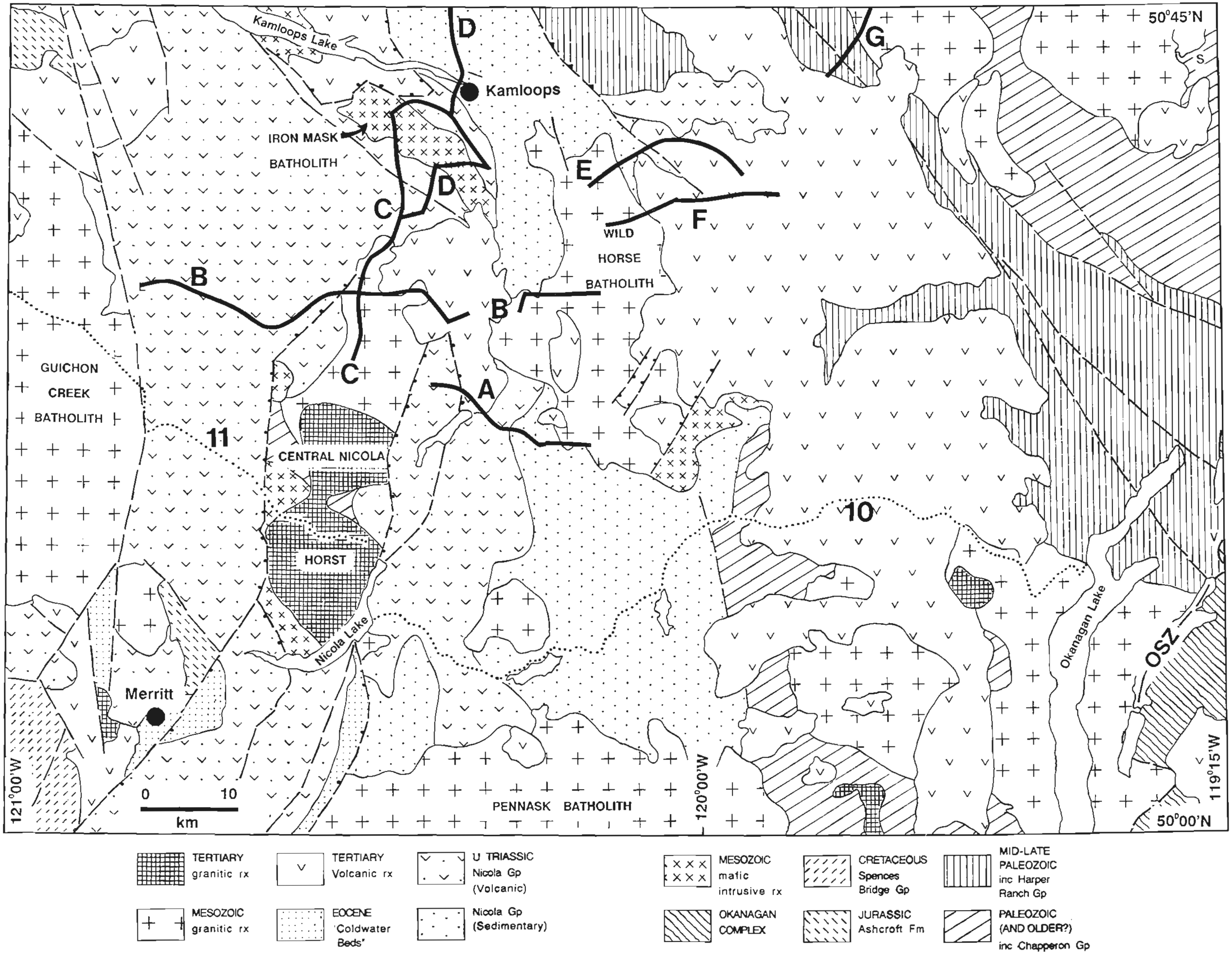

Figure 1. Geological map of region south of Kamloops (after Moore, 1989b); OSZ, Okanagan Shear Zone. 1989 gravity profiles are indicated by heavy solid lines labelled A through G. Lithoprobe lines 10 and 11 are indicated by dotted lines.

INTRODUCTION

The Lithoprobe southern Cordillera transect comprises 19 separate line segments. Detailed gravity surveys along lines 1 to 18 were carried out in the summers of 1987 and 1988. Preliminary results of these surveys are reported in Thomas et al. (1988 and 1989, respectively). Line 19 was completed in the summer of 1989 as an adjunct to regional gravity surveys undertaken in other parts of the Cordillera.

Moore (1989a,b) has described the results of recent geological investigations in a region of the Intermontane Belt south of Kamloops, lying between the Guichon Creek Batholith and Okanagan Lake (Fig. 1). The southern part of this region is traversed by Lithoprobe lines 10 and 11. One of the structures investigated is the Central Nicola Horst, a complex of "at least Mesozoic and early Tertiary plutonic rocks as well as metamorphosed supracrustal rocks of several ages" (Moore, 1989b). Thomas et al. (1989) had noted that the southern part of the horst, crossed by line 11, correlated with only a small negative gravity anomaly. This suggested that the 'granitic' elements of the horst in this area do not extend to any great depth. However, regional gravity data outline a sizable negative anomaly corresponding to the northern part of the horst, indicating thicker 'granites' in that area. It was concluded that detailed gravity investigations of the horst would be a worthwhile objective. Given the proximity of the region to line 19, additional gravity profiling could be carried out conveniently at the time that line 19 was surveyed. In addition to various profiles across the Central Nicola Horst, profiles were positioned to cross:

1 - the eastern margin of the Wild Horse Batholith,
2 - the Iron Mask Batholith, a NW-trending body of mafic igneous rocks immediately south of Kamloops,
3 - a belt of steep gravity gradients along Highway 1, east of Kamloops, coinciding with a strip of mid-upper Paleozoic rocks located between Tertiary volcanic rocks and Mesozoic granitic rocks.

GEOLOGY

The geology of the region south of Kamloops has been most recently described by Okulitch (1979), Monger and McMillan (1984) and Moore (1989a,b; Fig. 1). The area lies essentially within the Intermontane Belt, although a small segment of the Omineca Crystalline Belt appears in the extreme southeast corner, to the east of the Okanagan Shear Zone. The western part is dominated by Upper Triassic, arc-type volcanic and volcanogenic rocks of the Nicola Group. These rocks are intruded by Triassic and Jurassic calc-alkaline plutons. The eastern half includes large areas of Paleozoic rocks of oceanic affinity that are in both faulted and unconformable contact with rocks of the Nicola Group. This eastern sector is invaded by granitic plutons ranging in age from Triassic to Cretaceous. A large part of the eastern region is underlain by relatively flat-lying Tertiary volcanic rocks.

From a structural viewpoint the Central Nicola Horst is one of the more interesting features. It is a complex including at least four distinct Mesozoic and early Tertiary plutonic units, as well as supracrustal rocks of the Nicola and an older (?) group, metamorphosed to amphibolite facies. It is separated from surrounding sub-greenschist to greenschist grade metavolcanic rocks of the Nicola Group by steep, brittle Tertiary faults. Moore (1989a,b) considered the horst to be essentially uplifted basement that affords an insight into the probable roots of the Nicola arc.

GRAVITY SURVEYS

A total of 212 gravity stations were occupied along several profiles at a spacing varying generally from about 1 to 3 km. All measurements were made with Lacoste and Romberg gravity meter No. 172 and these were tied to base stations of the National Gravity Net. Elevation control along seismic line 19 was provided by temporary benchmarks established specifically for the seismic work using trigonometric methods. These benchmarks are tied to elevation base stations derived from satellite data; they are considered to have an accuracy of ± 5 m relative to the national geodetic datum. Individual benchmarks are probably accurate to within ± 1 m relative to the base elevation. Horizontal positions, also determined using trigonometric surveys, are believed to be accurate to within ± 20 m.

Elevation control for all other profiles was provided variously by provincial and federal geodetic benchmarks, altimetry tied to the benchmarks and data obtained from a global positioning system (GPS) satellite receiver. The accuracies of gravity station elevations derived using benchmarks and altimetry are estimated to be ± 5 cm and ± 3 m, respectively; the accuracy for elevations computed using GPS data is also estimated to be $\pm$ 3 m. Most horizontal positions were obtained using GPS; some were scaled from 1:50 000 topographic maps. The estimated accuracy is ± 20 m.

Bouguer anomalies were computed using the standard uniform density of 2.67 g/cm^3 and sea-level datum, and are considered to be accurate to within ± 1 mGal. This latter estimate does not consider terrain corrections, which have yet to be computed for the new data. Regional gravity data in the area south of Kamloops indicate that terrain corrections probably lie in the range 2 to 5 mGal, while data to the northeast along the Trans-Canada Highway and near line 19 indicate corrections as high as 10 to 20 mGal.

PRELIMINARY RESULTS

A Bouguer gravity anomaly map of the region, contoured at an interval of 5 mGal is shown in Figure 2. Because the map is based both on regional data that incorporate terrain corrections and on uncorrected profile data it must be remembered that slight adjustments to the contour patterns will result after application of corrections to profile data. However, for the purpose of the discussions in this report the map is of adequate precision. Gravity profiles along Lithoprobe lines (numbered) and supplementary profiles (lettered) are indicated by heavy solid lines; individual stations are not shown. Gravity stations established in earlier, mainly regional surveys, are indicated by dots; these are generally spaced about 10 to 20 km apart.

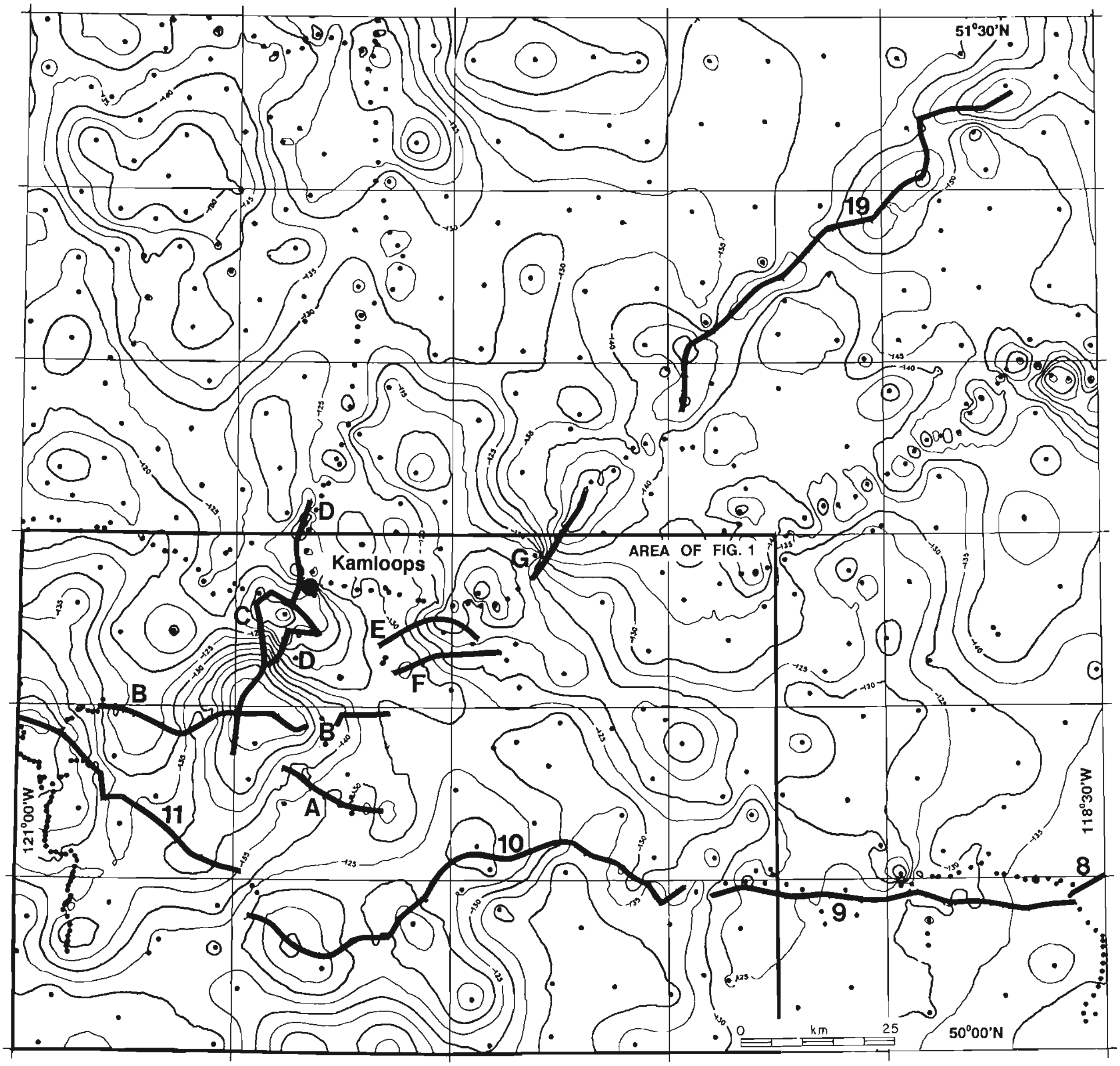

Figure 2. Bouguer gravity anomaly map of region embracing the area of the geological map (Fig.1) and an area to the north and east containing Lithoprobe line 19. Gravity profiles surveyed in 1989 are indicated by heavy solid lines labelled A through G and 19. Lithoprobe lines 8, 9, 10 and 11, along which detailed gravity observations were made in 1988, are also indicated. Contour interval is 5 mGal. Regional gravity stations and stations along pre-1989 profiles are indicated by dots; Bouguer anomalies computed for these stations include terrain corrections, those along the 1989 profiles do not.

Profiles A, B and C and the southern part of D, together with regional stations, now better define a large (−25 mGal amplitude) negative gravity anomaly associated with the northern part of the Central Nicola Horst. The horst has been interpreted by Moore to represent an exhumed deformed terrane that elsewhere is obscured by overlying mainly volcanic and volcanogenic rocks of the Upper Triassic Nicola Group. The presence of the gravity anomaly allows the deep structure to be investigated. Certainly where the anomaly reaches its peak amplitude near the intersection of profiles A and B, the structure should be traceable for several kilometres into the crust. Farther south, where the horst is crossed by seismic line 11, the amplitude of the anomaly is considerably reduced. This does not necessarily imply that the horst itself has a smaller vertical extent, but it does indicate that the thickness of the granitic component, which must be the main source of the anomaly, is greatly reduced.

The northern part of Profile C and Profile D cross a large mafic intrusive body, the Iron Mask Batholith. The southwestern and northeastern boundaries of this body are paralleled by NW-trending faults (not at the boundaries themselves). The association of a large positive gravity anomaly, roughly 25 mGal amplitude, with this body allows modelling of the body and possibly of the faults to be undertaken.

Profiles E and F were positioned across the northeastern margin of the Wild Horse Batholith to investigate the geometry of this boundary. The regional gravity data had indicated that there was little gravity expression associated with this intrusion, thereby implying that it exists only as a thin sheet. The detailed gravity data reinforce this conclusion. Southward along the strike of the eastern boundary of the batholith, roughly along longitude 120°W, the eastern contact of the Nicola Group with underlying, presumably upper Paleozoic rocks of the Chapperon Group is identified as an unconformity in at least one locality, but is believed to be a probable thrust elsewhere (Moore, 1989a,b). This raises the possibility that thrusting along this line of strike has affected the eastern margin of the Wild Horse intrusion and that the latter has the form of a thin thrust wedge.

Profile G along the Trans-Canada Highway results in better definition of a belt of steep gradients that probably is related to a contact between a large body of Mesozoic granitic rocks to the northeast and Paleozoic rocks to the southwest. Bouguer gravity anomaly values decrease across the gradient, in the direction of the granitic rocks, by about 35 mGal.

Gravity variations along Lithoprobe line 19 are generally fairly gentle. Terrain near this line is more rugged than in the region south of Kamloops, consequently until terrain corrections become available no comment is made on geological relationships.

FUTURE PLANS

The next task in the gravity program is to complete terrain corrections in order to arrive at a uniform data set. Density and magnetic susceptibility measurements will be made on rock samples collected from the area during the gravity surveys and ongoing geological studies of one of the authors (JM). Finally, modelling of the gravity anomalies will proceed, constrained by the rock densities, geological contacts and structural information, and features appearing on seismic reflection sections (available for Lithoprobe lines). Aeromagnetic data obtained along lines spaced roughly 800 m apart and covering the area of interest will also be examined and modelled, and may further help to constrain gravity models.

ACKNOWLEDGMENTS

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Volcanic stratigraphy and some aspects of alteration zonation in the western part of the Mount Skukum volcanic complex, southwestern Yukon[1]

David A. Love[2]

Love, D.A., Volcanic stratigraphy and some aspects of alteration zonation in the western part of the Mount Skukum volcanic complex, southwestern Yukon; in Current Research, Part E, Geological Survey of Canada, Paper 90-1E, p. 297-308, 1990.

Abstract

Eocene volcanic rocks of the Mount Skukum volcanic complex host the Mount Skukum epithermal gold deposit, situated approximately 65 km southwest of Whitehorse. The revised stratigraphy of the complex consists of four formations: the Mount Reid Formation (the lowest), monomictic conglomerates; the Butte Creek Formation, felsic pyroclastic rocks and epiclastic rocks; the Watson River Formation, andesitic flows and volcaniclastic rocks; and the Mount Kopje — Vesuvius Formation, felsic volcaniclastics and flows. Acid-sulphate — type alteration zoning occurs in a rhyolite stock that intruded along the intersection of two synvolcanic faults within the Mount Kopje — Vesuvius Formation in the Main Cirque area. The texture and mineralogy of the altered rocks and their occurrence in an intrusive body suggest that alteration formed by reaction with magmatic volatiles. Gold-mineralized veins are hosted in faults and fractures that are not directly related to volcanic activity, and may therefore be later than formation of the alunite.

Résumé

Des roches volcaniques éocènes du complexe volcanique du mont Skukum renferment le gisement aurifère épithermique du mont Skukum, situé à environ 65 km au sud-ouest de Whitehorse. La stratigraphie révisée du complexe comprend les quatres formations suivantes: la formation de Mount Reid (la plus basse), composée de conglomérats monogéniques; la formation de Butte Creek, composée de roches pyroclastiques felsiques et de roches épiclastiques; la formation de Watson River, composée de coulées andésitiques et de roches volcanoclastiques; et la formation de Mount Kopje-Vesuvius, composée de roches volcanoclastiques et de coulées felsiques. Une zonation due à l'altération de type acide-sulfate se manifeste dans un petit massif intensif de nature rhyolitique qui a pénétré la long de l'intersection de deux failles contemporaines du volcanisme, dans la formation de Mount Kopje-Vesuvius, dans la région de Main Cirque. La texture et la minéralogie des roches altérées ainsi que leur présence dans un massif intrusif semblent indiquer que l'altération s'est produite à la suite de la réaction avec des éléments volatiles magmatiques. Des filons aurifères remplissent des failles et des fractures qui ne sont pas directement associées à une activité volcanique et peuvent par conséquent être plus récentes que la formation de l'alunite.

[1] Contribution to Canada — Yukon Mineral Development Agreement 1985-1989. Project carried by Geological Survey of Canada, Mineral Resources Division.
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PURPOSE

This report presents a preliminary re-analysis of the stratigraphy and volcanic evolution of the Mount Skukum volcanic complex, based on approximately three and one-half months of mapping on surface, one month mapping underground, and two months logging drill core. The primary objectives of this research were: to refine present knowledge of the geological setting of the deposit, and to characterize, in detail, the mineralized rocks.

PREVIOUS WORK

M.J. Pride (née Smith) is engaged in a study of the overall geology of the Mount Skukum volcanic complex as part of a Ph.D. program (Smith, 1982, 1983; Pride, 1984, 1986). B.W.R. McDonald described the geology of the Mount Skukum deposit and studied the fluid inclusions and stable isotope geochemistry of the Cirque Zone (McDonald, 1987; McDonald et al., 1986; McDonald and Godwin, 1986). Doherty and Hart (1988) mapped all but the westernmost part of the Mount Skukum volcanic complex as part of a 1:50 000 scale mapping contract under the Canada — Yukon Economic Development Agreement.

GENERAL GEOLOGY

The Mount Skukum volcanic complex underlies part of the Tagish Highlands physiographic region, which constitutes the transition between the Intermontane and Coast Plutonic Belts (Holland, 1964). The complex is part of the Eocene Sloko Volcanic Province of southern Yukon and northern British Columbia (Fig. 1; Souther, 1977; Armstrong, 1988), representing the last major magmatic episode in the Canadian Cordillera. The magmatic front in Eocene time was about 250 km inland from the present location of the Aleutian trench, and the Mount Skukum complex itself, approximately 300 km inland from this trench.

The complex is preserved as a down-faulted block of intermediate — to — felsic volcanic rocks and derived sediments that lie unconformably on Cretaceous granitic rocks of the Coast Plutonic Complex, folded Jurassic sedimentary rocks of the Whitehorse Trough, and Precambrian (?) Yukon Group metasedimentary rocks of the Yukon Crystalline Terrane, which is part of the Stikine Terrane (Fig. 2). The volcanic complex is elliptical in plan, and covers an area of about 140 km^2. The volcanic and intrusive rocks in the complex have medium- to high-K calc-alkaline affinities (McDonald, 1987). The western and southern parts of the complex contain the remnants of an andesitic stratovolcano. Felsic volcaniclastic rocks dominate the eastern part of the complex, where the existence of a cauldron-subsidence structure has been postulated (Pride, 1985, 1986). A quartz-feldspar — porphyritic rhyolite stock intrudes the centre of the complex.

Quartz-carbonate-sericite veins located in the west-central part of the complex host the Mount Skukum gold deposit. The veins underlie Main Cirque, at the head of Butte Creek (Fig. 2 and 3). Mineralized veins cut most of the volcanic and hypabyssal rocks exposed in the deposit area. Both adularia-sericite and quartz-alunite alteration mineral assemblages are present in the Mount Skukum area, but gold mineralization was associated only with the former. The quartz-alunite — bearing mineral assemblage, locally named the "Alunite Cap" area (Fig. 3 and 4), occurs topographically higher than the gold-mineralized part of the system, and is barren.

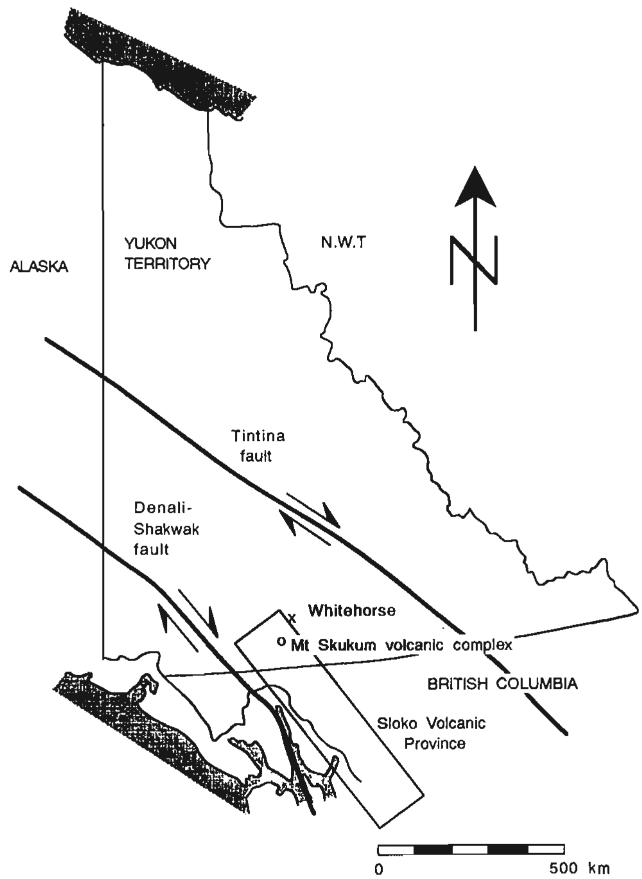


Figure 1. Location map of the Mount Skukum volcanic complex, Yukon Territory.

The structural geology of the vein systems in the Main Cirque area is described in the accompanying paper (Love, 1990). Underground mapping of the veins and associated fractures in the Brandy and Lake zones showed that the veins formed in a Riedel shear fracture system (Love, 1989, 1990), not a normal fault system as was previously thought (McDonald et al., 1986). Therefore their formation may not have been directly related to the volcanic evolution of the complex, but rather, may have been related to tectonism in the area. Structures in the Cirque Zone could not be mapped because mining there was completed. Reconstruction of the geology of the Zone, based on core-logging and company records, indicates that it opened by strike-slip movement on a fault, and that higher grades and thicknesses occurred where the fault refracted as it crossed a rhyolite dyke.

GEOLOGY OF THE WEST-CENTREL PART OF THE COMPLEX

STRATIGRAPHY OF THE SKUKUM GROUP

The geology of the area within several kilometres radius from the mine has been mapped by the writer (Fig. 3, 4).

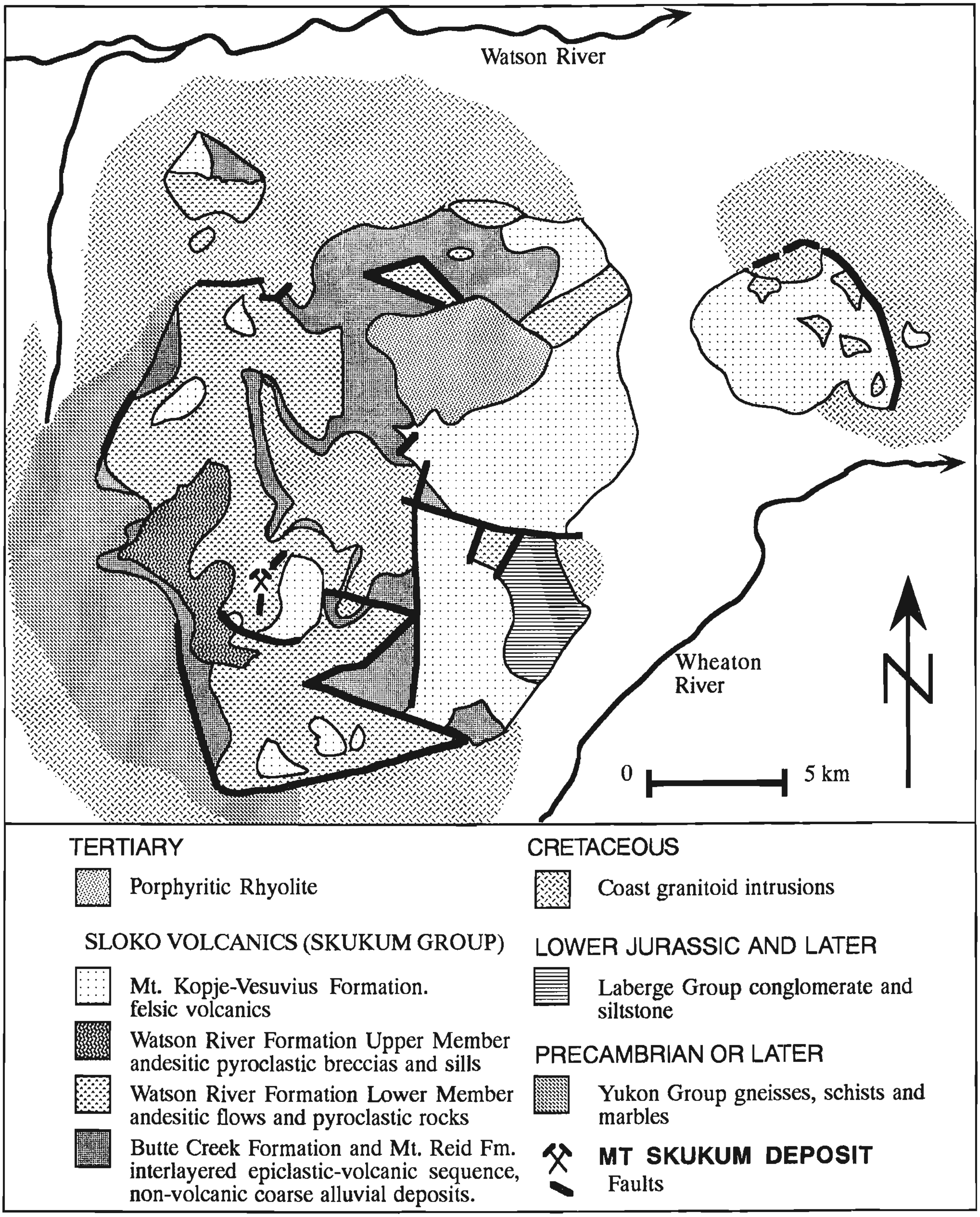
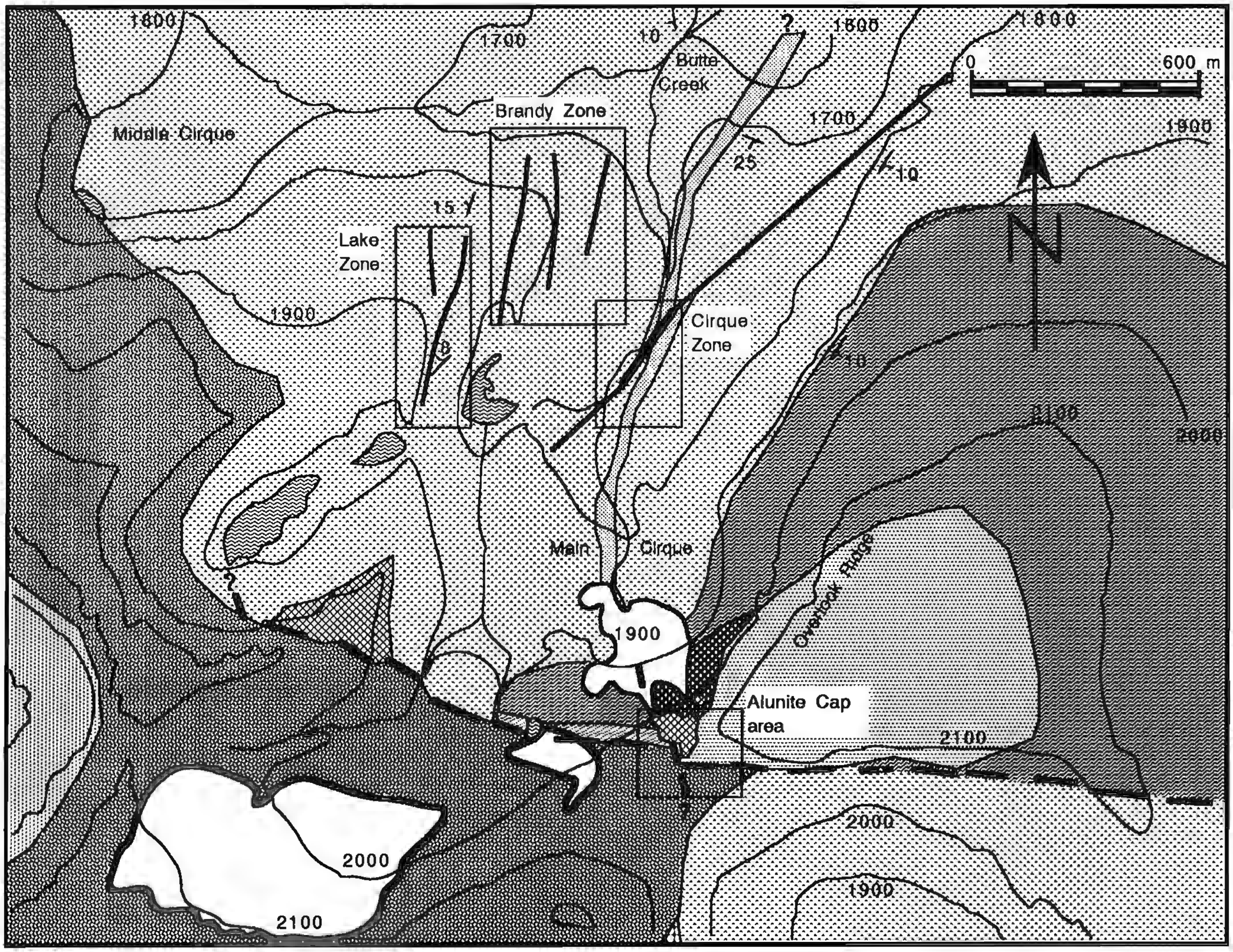


Figure 2. General geological map of the Mount Skukum volcanic complex (modified after Pride, 1986).



LEGEND

- Massive, flow-banded, or brecciated high-level rhyolite intrusions

Mount Kopje - Vesuvius Formation:
- Rhyolitic ignimbrite
- Mega-breccia / meso-breccia
- Massive rhyolite dyke and breccia dyke
- Rhyolite breccia and lapilli tuff
- Flow-banded rhyolite flows

Watson River Formation Upper Member:
- Monolithic andesite breccia
- Heterolithic andesite breccia

Lower Member:
- Andesite flows , tuffs, block tuffs, lapilli tuffs, and associated volcaniclastic sediments

- Creeks
- Snow fields and glaciers
- Faults
- Veins
- 1700 Contours in metres (contour interval 100 m)

Figure 3. Geological map of the Main Cirque area.

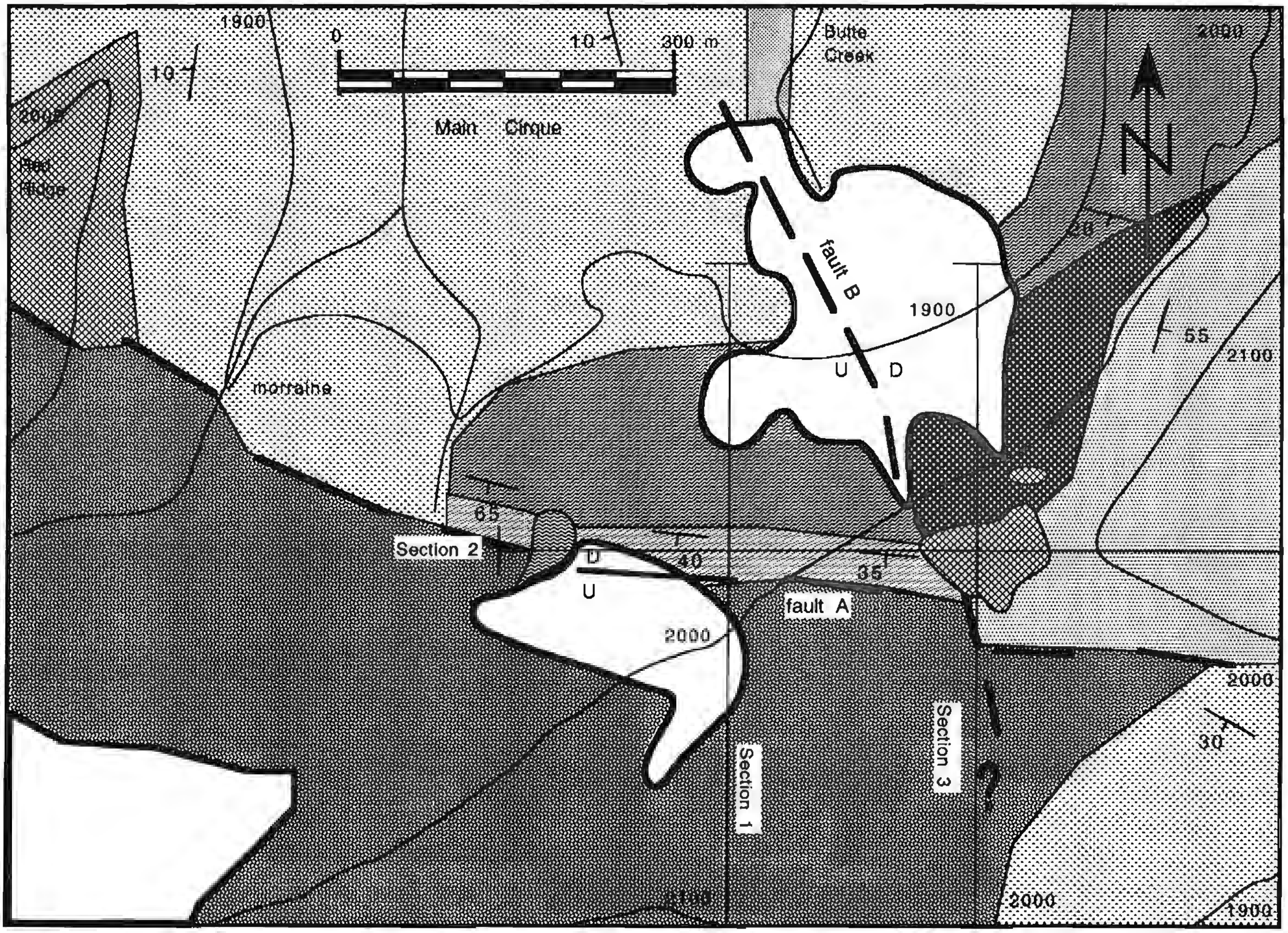


Figure 4. Geological map of the Alunite Cap area, with the alteration removed, showing the location cross-sections 1 to 3 (same legend as Fig. 3).

The geological history of the Main Cirque area interpreted from this mapping is generally similar to that proposed by McDonald (1987) and Doherty and Hart (1988), but differs from that of Pride (1985, 1986) (Table 1).

Recently-proposed geographic names for the formations (M.J. Pride, pers. comm.) are adopted in the following revised stratigraphy. These differ from the earlier usage of Pride (1985, 1986). Pride (*op. cit*) divided the volcanic stratigraphy into five lithostratigraphic formations numbered in apparent depositional order (Table 1). To maintain some nomenclatural stability Pride's numbered formations are also referred to. Present and previous stratigraphic nomenclatures of the Mount Skukum volcanic complex are compared and correlated in Table 1.

The lowest formation, the Mount Reid Formation (formerly Formation 1) (see Table 1), consists of monomictic clast- and matrix-supported pebble-boulder conglomerates and sandstones, lying unconformably on granodiorite basement. The Butte Creek Formation (formerly Formation 2) comprises felsic lapilli tuffs, block tuffs, and clast- and matrix-supported coarse conglomerates, and is locally gradational with the Mount Reid Formation. The boundary between the Mount Reid Formation and the Butte Creek Formation is marked by the first occurrence of volcanic fragments in the Butte Creek Formation (Pride, 1986). The Mount Reid Formation and the Butte Creek Formation are included in Doherty and Hart's (1988) Cycle 2 (Table 1).

The Butte Creek Formation is divided into six members (Table 2), referred to here as members 2 to 7, to maintain consistency with published descriptions (Pride, 1984). Pride (1985, 1986) redefined member 1 as Formation 1 (the Mount Reid Formation). Because the individual members of the Butte Creek Formation were not mappable on the geological map of the Mount Skukum volcanic complex, Pride (1985) divided that Formation into three map units. Although Pride did not document it, members 2 and 3 comprise the lower unit, and in the western part of the complex the upper map unit contains members 4, 5, 6, and 7. The third map unit outcrops only in the eastern and southern parts of the volcanic complex, and is thought to be the lateral equivalent of members 4, 5, 6, and 7, in which member 5, welded pumice lapilli-tuffs, is considerably thicker than in the west (Pride, 1985, 1986).

Table 1. Correlation of present and previous stratigraphic nomenclature of the Skukum Group in the Mount Skukum volcanic complex. *1 : redefined herein as the Upper Member of the Watson River Formation. *2: considered herein to be in the Lower Member of the Watson River Group. *3; coincides with the Mount Reid Formation.

Pride 1984 -based primarily on west-central part of the complex	Pride 1985 - 1:25,000 map, parts of 105 D/3, 4, 5 & 6	Pride 1986 -revised stratigraphy, more representative of the entire complex. [new names, Pride, in prep.]
-porphyritic rhyolite intrusions	-porphyritic rhyolite intrusion	-porphyritic rhyolite intrusion
	FORMATION 5 -intermediate monolithic pyroclastic breccia -heterolithic pyroclastic breccia -intermediate lava flows and tuffs	FORMATION 5 [Mt. Skukum Fm.] *1 -intermediate monolithic pyroclastic breccia -heterolithic pyroclastic breccia -intermediate lava flows and associated lapilli tuffs
EASTERN PART -brecciated, spherulitic, and flow-layered rhyolite lava flows and intrusions -altered felsic pyroclastic and epiclastic rocks	FORMATION 4 -altered felsic volcanic rocks -felsic lapilli tuff and tuff, interpreted as mostly welded pyroclastic flow -felsic flow banded spherulitic and brecciated flows and intrusions(?) -layered felsic volcaniclastic rocks	FORMATION 4 [Mt. Kopje - Vesuvius Fm.] -altered felsic volcanic rocks -felsic lapilli tuff and tuff, interpreted as mostly welded pyroclastic flow -felsic flow banded spherulitic and brecciated flows and intrusions(?) -layered felsic volcaniclastic rocks
WESTERN PART UPPER UNIT -andesite lava flows, pyroclastic and sedimentary rock	FORMATION 3 -interlayered epiclastic rocks and intermediate lava flows (6)	FORMATION 3 [Watson River Fm.] -interlayered epiclastic rocks and intermediate lava flows
LOWER INTERLAYERED SEDIMENTARY-VOLCANIC SEQUENCE -MEMBER 7 -planar bedded ash and conglomerate, reversely graded in part, thickens to the north -MEMBER 6 -interbedded siltstone and sandstone -MEMBER 5 -densely to moderately welded felsic pyroclastic flow -MEMBER 4 -clast-supported conglomerate -MEMBER 3-heterolithic debris and/or pyroclastic flows -MEMBER 2-interbedded sandstone and siltstone	FORMATION 2 -interlayered tuff and lapilli tuff, epiclastic rocks and lava flows, includes a felsic welded pyroclastic marker horizon, = Members 4,5,6 &7 -lapilli tuff, tuff and epiclastic rocks, comprises mostly pyroclastic rocks that may be equivalent to the welded felsic pyroclastic flow marker horizon, =Members 4, 5, 6 & 7 in eastern and southern parts of the complex -interlayered primary felsic volcanics, epiclastic rocks and minor lava flows, =Members 2 & 3	FORMATION 2 [Butte Creek Fm.] -planar bedded tuff and conglomerate, reversely graded in part, and interbedded siltstone and sandstone, only found locally in the central part of complex, =Members 6 & 7, becomes thicker in the north -densely to moderately welded felsic pyroclastic flow and clast supported conglomerate, =Members 4 & 5, becomes much thicker in the east -felsic volcanic and associated epiclastic rocks, heterolithic debris flows, felsic pyroclastic flows, tuffs, sandstone, siltstone, and conglomerates, = Members 2 & 3, becomes much thicker in the south
-MEMBER 1-monolithic granitic boulder to pebble debris flows	FORMATION 1 -nonvolcanic conglomerate, minor sandstone and siltstone	FORMATION 1 [Mt. Reid Fm.] -clast-supported conglomerate, muddy matrix-supported conglomerate, and minor sandstone and siltstone, with exclusively basement fragments

McDonald 1987 -Main Cirque area	Doherty et al. 1988 - 1:50,000 maps 105 D/3 & 104 D/6	this study - 1:5000 105 D/3 NW corner, west central part of the complex
-porphyritic rhyolite	Cycle 7 Erfp -rhyolite feldspar porphyry, high level, felsic domes, plugs and laccoliths Er -rhyolite dykes, fine-grained rhyolite, dykes, dyke swarms, composite dykes, flow domes	-fine-grained, sometimes porphyritic intermediate dykes -high level rhyolite porphyry intrusion -flow-banded, brecciated rhyolite dykes -intrusive breccia containing brecciated, massive and aphanitic, to flow-banded rhyolite
FORMATION 4 -rhyolite flows and welded to non-welded pyroclastic rocks	Cycle 6 Elt -felsic pyroclastic rocks, rhyolite to intermediate tuff, lithic tuff and welded tuff Ewt -welded felsic tuff, columnar jointed, densely welded, includes unwelded tuff on Vesuvius Hill Ert -rhyolite lithic tuff, probably equivalent to Erf or Ewt Embx -felsic megabreccia, intracaldera collapse breccia, =no separate unit on other maps Erf -rhyolite flows Es3 -epiclastic tuffs and sediments, sandstone, siltstone, occurs locally at the base of Erf Cycle 5 Esk -Skukum Group, undifferentiated	MOUNT KOPJE - VESUVIUS FORMATION (in the Main Cirque area, thicker and more variable to the east) -rhyolitic densely welded, pumice lapilli tuff -heterolithic megabreccia to mesobreccia containing blocks of porphyritic andesite, andesite lapilli tuff, andesitic volcaniclastic rocks, conglomerate and flow-banded rhyolite -well bedded rhyolite pyroclastic rocks -flow-banded, brecciated and spherulitic rhyolite flows
FORMATION 3 -thick, poorly-bedded, monolithic andesite breccia -interbedded andesitic pyroclastic and porphyritic andesite flow rocks -porphyritic andesite flow rocks -interbedded porphyritic andesite flow rocks, andesitic debris flows, and derived epiclastic rocks	Cycle 4 Eab -andesite breccia, massive monolithic breccia of porphyritic andesite fragments Cycle 3 Ean -andesite flows and tuffs, massive to poorly bedded tuffs and porphyritic flows	WATSON RIVER FORMATION UPPER MEMBER -monolithologic andesitic block tuff and bomb tuff -heterolithic pyroclastic breccia LOWER MEMBER -andesitic or dacitic lapilli tuffs and ash tuffs -andesitic volcaniclastic rocks, including pyroclastic breccia, lapilli-tuff, and conglomerate -interbedded andesitic or dacitic ash tuff and lapilli tuff and plagioclase porphyritic andesite flows -plagioclase porphyritic andesite flows -interbedded plagioclase porphyritic andesite flows, andesitic debris flow rocks, and derived epiclastic rocks
FORMATION 2 -andesitic lapilli tuff -andesite ash tuff -volcaniclastic sedimentary rocks -debris flow/lahar rocks -rhyolite ash tuff -rhyolite lapilli tuff	Cycle 2 Edt -dacite to andesite lithic tuff *2 Es2 -interlayered epiclastic epiclastic sediments, bedded tuff, lithic tuff, and a densely welded ignimbrite marker bed, = Members 4, 5, 6 & 7 Es1 -tuffs and epiclastic rocks, moderate to well bedded, felsic, fresh or altered pyroclastic rocks, commonly interbedded with epiclastic rocks, = Members 2 & 3 Es -tuffs and epiclastic rocks, Es1 and Es2 unsubdivided Ecg -conglomerate, massive, clast supported, cobble and boulder conglomerate, contains locally derived basement clasts *3	BUTTE CREEK FORMATION -planar bedded ash and conglomerate, reversely graded in part, =Member 7 -interbedded siltstone and sandstone, =Member 6 -densely to moderately welded felsic pyroclastic flow, =Member 5 -clast-supported conglomerate, =Member 4 -heterolithic pyroclastic flows and minor debris flows, thickens to the south, =Member 3 -interbedded sandstone and siltstone, =Member 2
FORMATION 1 -interbedded siltstone , sandstone and debris flows -granitic-boulder conglomerate		MOUNT REID FORMATION -clast-supported conglomerate, and minor sandstone and siltstone, with exclusively granitic basement fragments
	Cycle 1 Eva -felsic volcanics, dark vitreous rhyolite and dacite flows and welded tuffs, containing granite fragments, hornfelsed at contact with Erfp, maybe late Cretaceous age	

The Watson River Formation (formerly Formation 3) consists of andesitic lava flows, andesitic breccia tuffs, and minor andesitic or dacitic tuff and lapilli tuff. The boundary between the Butte Creek and the Watson River Formation is at the base of the first lava flow in the Watson River Formation.

In the Main Cirque area the Mount Kopje — Vesuvius Formation (formerly Formation 4) in only a few hundred metres thick and comprises flow-banded, spherulitic and brecciated felsic flows, felsic lapilli tuff, megabreccia and mesobreccia, and ignimbrite. However, in the northeastern part of the complex it is thicker and more variable in rock type. Correlation of the Mount Kopje — Vesuvius Formation between the eastern and western parts of the complex is based on the thick, flow-banded, spherulitic and brecciated felsic flows.

Cycle 5 of Doherty and Hart (1988) outcrops only in the eastern part of the complex, and is absent in the Main Cirque area; but because of its overall felsic composition, and its place in the stratigraphic sequence it must constitute the lower part of the Mount Kopje — Vesuvius Formation. That portion of the Mount Kopje — Vesuvius Formation that occurs in the Main Cirque area coincides with Cycle 6 of Doherty and Hart (1988) (Table 1).

The intermediate heterolithic and monolithologic pyroclastic breccias that overlie the intermediate flows and tuffs of the Watson River Formation are redefined as the Upper Member of that Formation, and the underlying intermediate flows and tuffs, as the Lower Member. Formation 5 was originally defined as consisting mainly of these heterolithic and monolithologic andesitic breccias, but also included interlayered andesitic lava flows and pyroclastic rocks west of Main Cirque and south of Middle Cirque (Table 1; Pride, 1986). Pride (1985, 1986) placed the lower contact of this formation at an arbitrary level in a continuous sequence of andesite flows and tuffs, at a point where the composition changes somewhat (MgO content increases slightly upwards, (McDonald, 1987)). Outside of the Main Cirque area, Pride placed the lower contact at the base of the heterolithic breccia, or, where the heterolithic breccia is absent, at the base of the monolithologic breccia.

In Pride's (1985, 1986) interpretation of the stratigraphy, the intermediate pyroclastic breccias unconformably overlay the Mount Kopje — Vesuvius Formation (Table 1). However, the author's mapping in the area around Main Cirque (see Map 1 in Love, 1989; and Fig. 4) shows that, although felsic flows, lapilli tuff and ignimbrite of the Mount Kopje — Vesuvius Formation are lower in elevation than some of the intermediate pyroclastic breccias, they are delimited by caldera-margin — type mesobreccias, and were down-dropped on caldera- or crater-bounding faults. It is concluded, therefore, that they were deposited after the intermediate pyroclastic breccias.

In Middle Cirque and on the south and east sides of Mount Skukum, the intermediate pyroclastic breccia directly overlies intermediate flows and tuffs of the Watson River Formation. In most exposures their contact is conformable. On the south side of Middle Cirque their contact is obviously an angular unconformity; however, flows and tuffs of the Watson River Formation there are nearly horizontal, and the steeply-dipping contact probably represents the wall of a volcanic crater. At the southwest end of Main Cirque, flows of the Watson River Formation are overlain by heterolithic breccia and then monolithologic breccia. At the southeastern end of Main Cirque the contact relationship between the flows and tuffs of the Watson River Formation and the overlying intermediate pyroclastic breccias is hidden by down-faulted felsic volcanic rocks of the Mount Kopje — Vesuvius Formation.

Because the definitions of all the formations in the Mount Skukum volcanic complex are based on their lithic characteristics and stratigraphic positions, and the pyroclastic breccias that immediately overlie the intermediate flows and tuffs of the Watson River Formation are lithologically similar to that formation, they should not have separate formation status, but should be included in the Watson River Formation. Use of the term Mount Skukum Formation or Formation 5 of Pride (1985, 1986) should be abandoned from the lithostratigraphy of the complex because it was improperly defined. The use of the name Mount Skukum for this redefined lithostratigraphic unit should also be abandoned because that name is already used in the Mount Skukum volcanic complex and the Skukum Group. The intermediate pyroclastic breccias shall be referred to as the Upper Member of the Watson River Formation, and the underlying intermediate flows and tuffs, as the Lower Member (Table 1).

The Lower Member of the Watson River Formation coincides with Doherty and Hart's (1988) Cycle 3, and the Upper Member, with their Cycle 4.

Doherty and Hart (1988) included their map unit E_{dt}, an assemblage of dacitic to andesitic lithic tuff that covers much of the northwestern part of the complex, in their Cycle 2 (Table 1). However, in the Camp Cirque, Frigid Cirque and Tuning Fork Creek areas, north and west of Butte Creek, plagioclase-porphyritic andesite flows directly overlie the Butte Creek Formation, and are in turn overlain by andesite to dacite lithic tuff. In the Camp Cirque area, interbedded sandstone and siltstone of Member 6 of the Butte Creek Formation is overlain by andesite flows. In the Frigid Cirque area andesite flows overlie planar-bedded ash and conglomerate of Member 7 of the Butte Creek Formation. Andesite flows overlie densely-welded ignimbrite of Member 5 of the Butte Creek Formation on the west side of Tuning Fork Creek. These flows, by definition, mark the base of the Watson River Formation, and therefore the overlying andesite to dacite tuffs must be contained in, or lie above, the Watson River Formation. Because the andesitic to dacitic lithic tuffs are overlain by intermediate pyroclastic breccia of the Upper Member of the Watson River Formation on the south side of Camp Cirque, they must belong in the Lower Member of that Formation (Cycle 3 of Doherty and Hart, 1988). In the northwest part of the complex, there are significantly fewer flows and more tuffs in the Lower Member of the Watson River Formation, suggesting the occurrence of a more distal facies in that area.

Cycles in the Skukum Group

A cycle is a rhythmic sequence of related processed and conditions repeated in the same order and preserved in the rock record. The use of the term cycle to name the formations (as in McDonald et al., 1986; and Doherty and Hart, 1988) should be discontinued because none of these units constitutes a cycle; rather, several of them make up a cycle. The Butte Creek Formation and the Watson River Formation exhibit an overall compositional trend, from felsic ignimbrite to andesitic tuff upwards in the Butte Creek Fromation, continuing to mafic andesite in the Watson River Formation, as described above. The felsic rocks of the Mount Kopje — Vesuvius Formation could represent the lower part of a second cycle, which was either aborted before it reached intermediate composition or partly eroded so that the more mafic upper part of the cycle was not preserved. A clear distinction must be maintained between the division of a stratigraphic column into groups, formations and members, and its division into cyclic or rhythmic repetitive sequences (North American Commission on Stratigraphic Nomenclature, 1983).

Synvolcanic faults in the Main Cirque area

The Mount Kopje — Vesuvius Formation contains evidence of two synvolcanic faults. This formation, in Main Cirque, can be divided into upper and lower parts: the lower comprises flow-banded felsic flows, and felsic lapilli and breccia tuffs; and the upper, a caldera-collapse breccia and a welded ingnimbrite. The felsic flows unconformably overlie andesitic flows and tuffs of the Watson River Formation (Fig. 5). They are conformably overlain by felsic breccia and lapilli tuffs. The southern contact of felsic flows and tuffs against the Watson River Formation is a steep, east-west — trending, undulose surface that is interpreted as a fault (fault A, see Fig. 4 and 5). A westward thickening wedge of mesobreccia to megabreccia, containing blocks of porphyritic andesite, andesite tuffs, conglomerate and flow-banded rhyolite, separates the rhyolitic flows and tuffs from a welded pumice ingnimbrite (Fig. 4, 6). The abrupt vertical western contact of the mesobreccia against rhyolitic flows and tuff (see Fig. 6) is interpreted as a fault scarp (fault B), and the mesobreccia is interpreted as a caldera-collapse breccia, as described in Colorado by Lipman (1976).

The undulose nature of fault A and its offset by fault B (Fig. 4), which is clearly synvolcanic, indicate that it was synvolcanic and may be a caldera margin fault, or fault scarp. The exact age of fault A is not clearly defined because the Mount Kopje — Vesuvius Formation is not exposed south of this fault: it could either predate or postdate the rhyolite flows and tuffs on its north side. If it predated them, then the exposed part would be a fault scarp, and the rhyolite flows and tuffs could have ponded against the scarp. If it postdated the rhyolite flows and tuffs, then they could have been more aerially extensive south of the fault trace and have been removed by erosion.

The episode of faulting represented by the fault scarp (fault B), and the deposition of the breccias, were followed by, and probably related to, eruption of the ignimbrite, which overlies both the mesobreccia and the felsic flows of

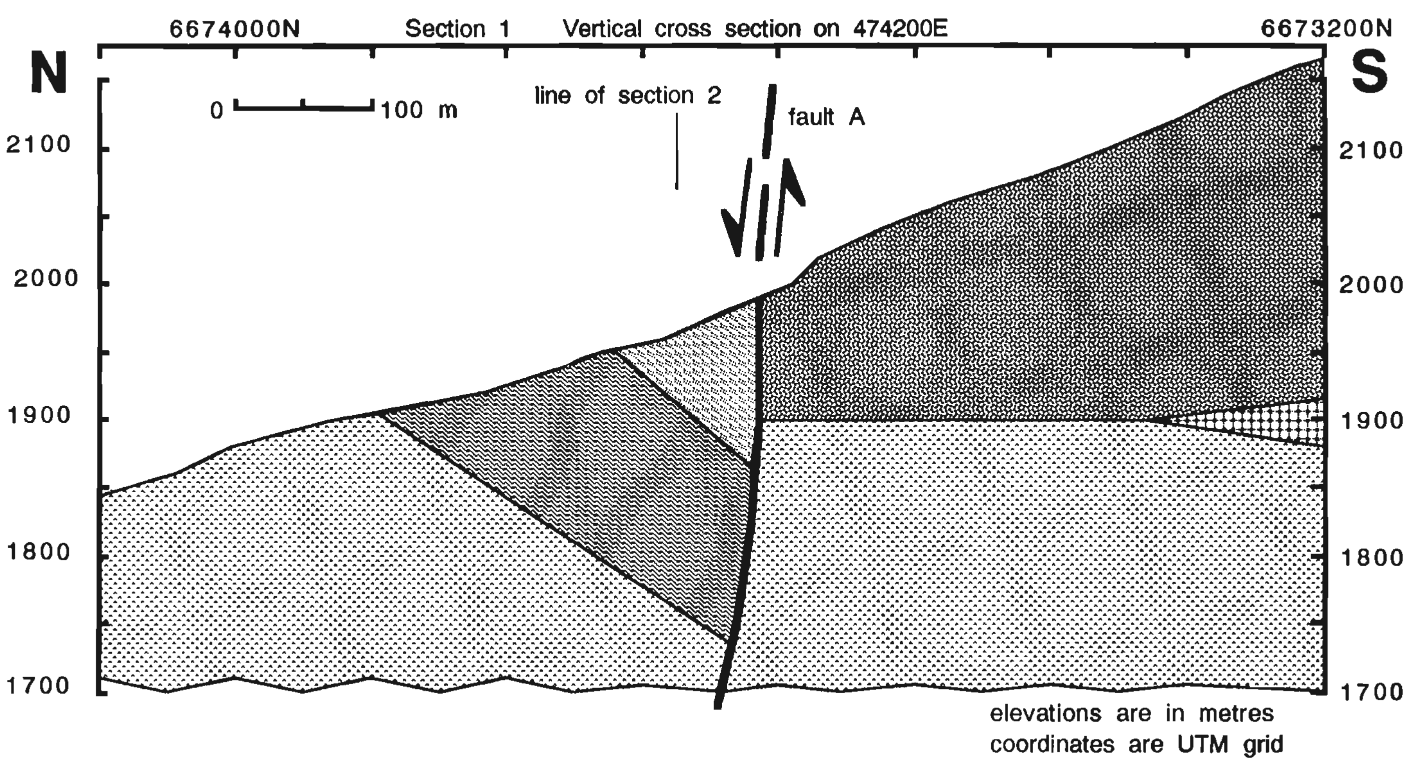


Figure 5. North – south cross-section (No. 1) through the Alunite Cap area (looking east), showing the stratigraphy and an east-west trending synvolcanic fault (fault A), outside the acid-sulphate alteration zone (same legend as Fig. 3).

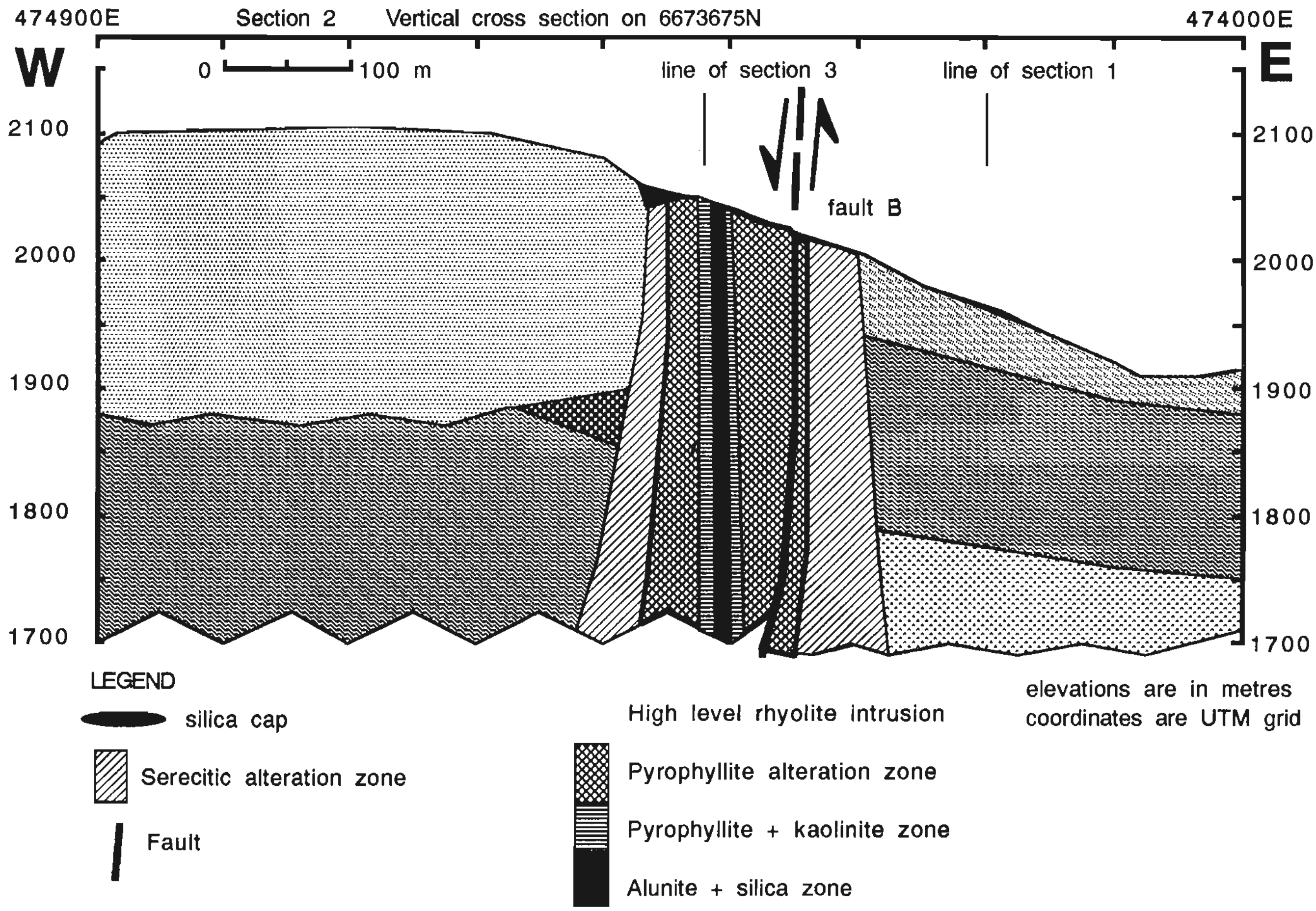


Figure 6. East – west cross-section (No. 2) through the Alunite Cap area (looking south), showing the acid-sulphate alteration zone in a rhyolite stock that intrudes a north-south – trending synvolcanic fault (fault B) (same legend as Fig. 3).

the lower part of the Mount Kopje — Vesuvius Formation. Felsic breccia and lapilli tuffs of the Mount Kopje — Vesuvius Formation are not present beneath the ignimbrite (Fig. 4, 6), probably because they were eroded during its eruption.

The massive to flow-banded rhyolite exposed on Red Ridge, about 900 m west-northwest of the Alunite Cap area, has a gently southwest-dipping lower contact against andesitic flows and lapilli tuffs of the Lower Member of the Watson River Formation (Fig. 3). The southern contact of this rhyolite body, against monolithological andesite breccia of the Upper Member of the Watson River Formation, is sharp and nearly vertical (Fig. 3). This vertical southern contact may be the western extension of fault A, described above. No rhyolitic volcaniclastic rocks are exposed on Red Ridge. The rhyolite of Red Ridge could correlate with the flows of the Mount Kopje — Vesuvius Formation in the Alunite Cap area, or it could be a subvolcanic sill or laccolith.

An irregularly-shaped, flow-banded and autobrecciated rhyolite stock, approximately 100 × 150 m in plan, intrudes the intersection of the two synvolcanic faults (Fig 6, 7). The barren quartz-alunite — bearing alteration assemblage, after which the "Alunite Cap" area was named, occurs in this intrusion. The rhyolite was rebrecciated by hydrothermal activity. Silicified breccia, containing rounded, probably hydraulically milled, fragments of rhyolite, occurs in the Alunite Cap area.

Alteration in the Alunite Cap Area

In the Alunite Cap area, alteration mineralogy grades outward from massive alunite + silica, through zones of pyrophyllite ± kaolinite, pyrophyllite ± silica, and sericite, to a peripheral zone of very weak chloritically altered rocks (see Map 1, Love, 1989; and Fig. 6, 7). The alteration zones shown in Figures 6 and 7 are projected vertically from those mapped on surface. It was earlier thought that the ignimbrite of the Mount Kopje — Vesuvius Formation was altered to such an extent that it lost its original texture (Love, 1989). However, detailed examination of drill-core from this zone indicated that a previously unrecognized rhyolite stock hosts the inner three alteration zones. A more detailed mineralogical study of samples from the drillholes will further clarify the nature of the alteration zones at depth.

Advanced argillic alteration minerals occur in both adularia-sericite — type and acid-sulphate — type epithermal deposits (Heald et al., 1987). The argillic alteration

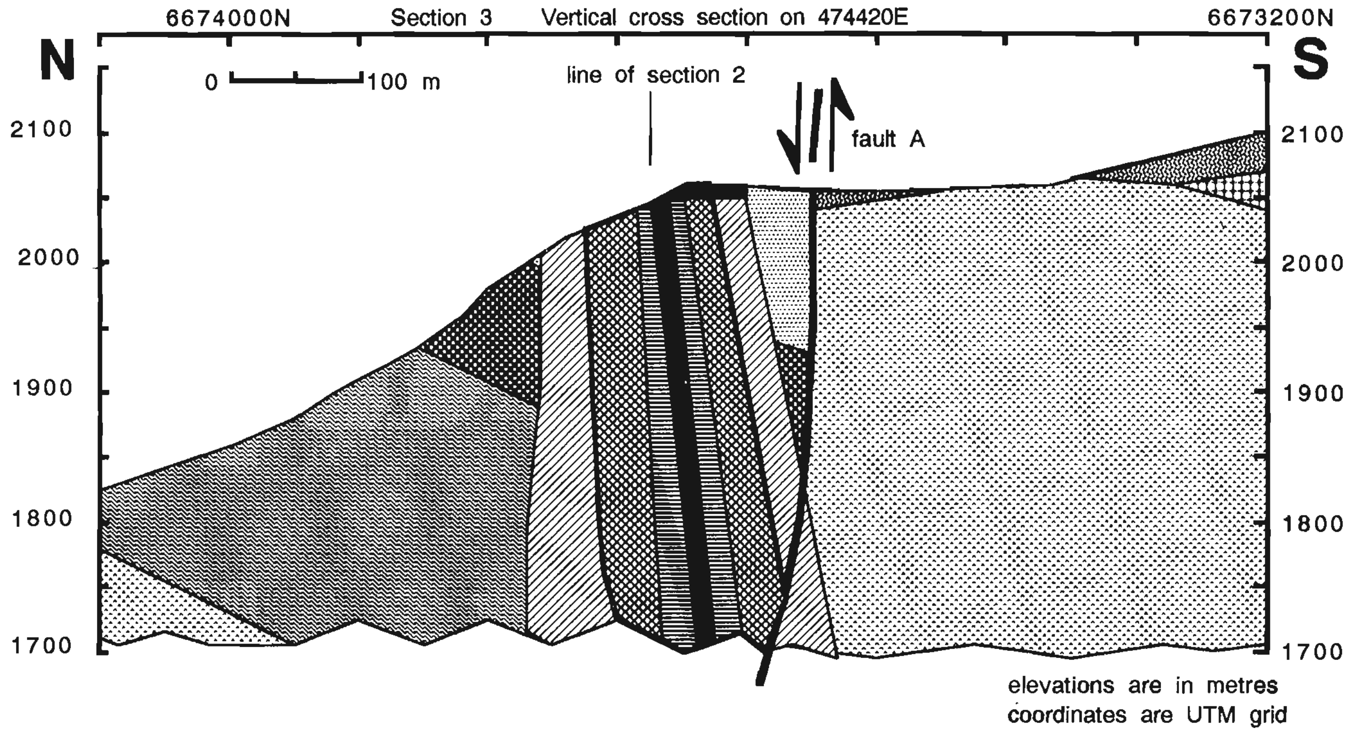


Figure 7. North – south-cross section (No. 3) through the Alunite Cap area (looking east), showing the acid-sulphate alteration zone and the east-west – trending synvolcanic fault (fault A) same legends as Figs. 3 and 6).

associated with acid-sulphate — type gold mineralization is thought to have formed by disproportionation of magmatic SO_2, whereas the barren caps found above adularia-sericite vein deposits formed by oxidation of hydrogen sulphide separated from deep near-neutral brines during boiling (Heald et al., 1987). The presence of pyrophyllite and the relatively coarse grain-size and massive texture of the alunite-silica alteration minerals at Mount Skukum, as well as their occurrence in a rhyolite stock suggest that this alteration was similar to that formed by disproportionation of magmatic SO_2 (R.H. Sillitoe, pers. comm., 1989).

The Environment of Veining

McDonald et al. (1987) suggested that the morphology of Main Cirque, in which the ore deposit outcrops, was controlled by high-angle normal faults that produced a steplike topography comprising three down-dropped blocks. Mapping, by the author, in the Alunite Cap area has confirmed the presence of one of these faults, the north-south — trending synvolcanic fault (fault A) through the Alunite Cap, described above. This fault extends north to north-northwest from the Alunite Cap, towards the approximate area of the Brandy Zone, and has about 80 m vertical displacement in the Alunite Cap area, with the east side down. However, there is no indication of a fault with this amount of displacement in the Brandy Zone area. This fault could have developed into a set of many smaller-displacement normal faults there, or it could be a rotational hinge fault on which displacement decreases to the north.

A three-dimensional strain analysis of the Brandy Zone (in Love, 1990) demonstrated that the vein-hosting fractures are shear fractures formed by horizontal compression, and that further strike-slip movement as well as dilation resulted in vein emplacement. Analysis of the structure of the Cirque Zone indicated that the maximum stress axes at the time of its formation must have been oriented approximately north-south, and movement on the fault that hosts the Zone was horizontal sinistral (Love, 1990). McDonald et al. (1987) suggested that the veins formed in the bounding faults between the down-dropped blocks, when they were reactivated as a result of resurgent doming. The structural analyses of the Brandy and Cirque zones preclude this possibility.

In as much as the alunite alteration at Mount Skukum, hosted in a rhyolite stock localized by synvolcanic faults, is barren, but the gold mineralization is hosted in fracture-filling veins, not directly related to volcanic activity, the gold may have been introduced later than the intrusion.

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Barium carbonate bodies associated with the Walt (Cathy) stratiform barium deposit, Selwyn Basin, Yukon: a possible vent complex associated with a Middle Devonian sedimentary exhalative barite deposit

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Turner, R.J.W. and Goodfellow, W.D., Barium carbonate bodies associated with the Walt (Cathy) stratiform barium deposit, Selwyn Basin, Yukon: a possible vent complex associated with a Middle Devonian sedimentary exhalative barite deposit; in Current Research, Part E, Geological Survey of Canada, Paper 90-1E, p. 309-319, 1990.

Abstract

A stratabound sheet-like body up to 60m thick and 2 km in strike length occurs within Middle Devonian chert and siliceous shale of the Lower Earn Group, Selwyn Basin. This baritic body is zoned from a brecciated barium carbonate core outwards to massive barium carbonate, massive barite and laminated barite. Textural relationships indicate barium carbonate replacement of sedimentary laminated barite; the latter is altered to massive, recrystallized barite at the replacement front. The barite is interpreted to have precipitated in the water column and settled to the seafloor following the mixing of hydrothermal barium discharged from vents with seawater sulphate. The barium carbonate core may represent a vent complex formed by the hydrothermal alteration and replacement of sedimentary barite above and laterally adjacent to the vent. Such a barium carbonate vent complex is analogous to ferroan carbonate vent complexes associated with nearby Zn-Pb SEDEX deposits.

Résumé

Un corps stratiforme, atteignant 60 m d'épaisseur et 2 km de long suivant la direction, se manifeste au sein de couches de cherts et de shales silicieux du Dévonien moyen du groupe de Lower Earn, dans le bassin de Selwyn. Ce massif de barytine montre une zonation qui va d'un carbonate de barium bréchoïde au centre à un carbonate de baryum massif, une barytine massive et une barytine feuilletée à l'extérieur. Des relations structurales indiquent que le carbonate de baryum remplace la barytine sédimentaire feuilletée; cette dernière se présente sous forme altéré à massif de barytine recristalisée d'aspect au front de remplacement. D'après les auteurs, la barytine s'est précipitée dans la colonne d'eau et s'est déposée au fond de la mer à la suite du mélange de baryum hydrothermal rejeté des cheminées et de sulfate marin. Le noyau de carbonate de baryum peut représenter un complexe de cheminées formé par l'altération hydrothermale et le remplacement de la barytine sédimentaire au-dessus des cheminées et adjacente latéralement à ces dernières. Ce complexe de cheminées de carbonate de baryum est analogue aux complexes de cheminées de carbonate de fer associées aux gisements zinco-plombifères de type SEDEX avoisinants.

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INTRODUCTION

The genesis of stratiform sediment-hosted barite deposits is poorly understood. Most workers have interpreted them to be sedimentary in origin, related to the submarine discharge of barium rich hydrothermal fluids (e.g. Lydon et al., 1985; Poole, 1988), although a diagenetic origin of some barite deposits has been advocated (Rye et al., 1978). Models for a sedimentary exhalative origin for barite include sedimentation adjacent to a submarine exhalative vent (Lydon et al., 1985), sedimentation at a redox interface within a stratified watermass ('bathtub-ring') distant from an exhalative centre, or sedimentation related to overturn and oxygenation of a reduced water mass such as Middle to Upper Silurian bioturbated mudstone, Selwyn Basin (Goodfellow and Jonasson, 1986).

Hydrothermal barite is also associated with laminated sulphides in stratiform sediment-hosted Zn-Pb deposits (e.g. Jason, Tom), or as massive bodies lateral to sulphide ores (e.g. Meggen, West Germany; Silvermines, Ireland). Stratiform barite deposits in Middle Devonian to lower Mississippian outer miogeoclinal strata throughout Alaska, Yukon and British Columbia are spatially and temporally related to stratiform zinc-lead deposits (Abbott et al., 1986; Turner, 1988; Turner et al., 1989b). Stratiform barite deposits have been interpreted to form from lower temperature hydrothermal fluids than stratiform Zn-Pb-barite deposits (Lydon et al., 1985). Alternatively, stratiform barite deposits may represent discharge of zinc-, lead- and barium-rich fluids into an oxygen-rich seawater causing deposition of barite but dispersal of base metals into seawater (Lydon et al., 1979).

Our understanding of stratiform sediment-hosted barite deposits is hampered by a lack of well described deposits. This paper presents the preliminary results of a field based study of one such deposit.

LOCATION AND HISTORY OF WORK

The Walt property is located near MacMillan Pass, Yukon, 400 km northest of Whitehorse, Yukon (63°17′N, 130°33′) (Fig. 1) and includes the Cathy and Row claims. The Walt deposit outcrops along the crest of a southwest-trending ridge above tree line at about 1700 m elevation. In 1980, Baroid of Canada Ltd drilled 10 holes (899m) that outlined a zone up to 30 m thick and 150 m long containing 450 000 tonnes of baritic rock with a specific gravity above 4.25 (Abbott, 1981).

OBJECTIVES OF THE STUDY

The objective of the study was to establish the internal stratigraphy of the Walt deposit to further our understanding of the genesis of barite deposits. The relationship of the bedded barite to adjacent prominant, cliff-forming carbonate units was a major focus of the study. Dawson and Orchard (1982) interpreted these breccia-rich carbonate units as sedimentary olististromes; however they also reported the presence of barium carbonates, unusual for sedimentary carbonates. Abbott (1981) noted irregular lenses of barite within these carbonate breccias.

Four days were spent by the authors during July 1989 on the Walt property and vicinity. The deposit was mapped at a scale of 1: 1500, 8 stratigraphic sections were measured, and a suite of samples collected for petrography and chemical analysis.

REGIONAL SETTING

The Walt barite deposit occurs within the MacMillan Fold Belt (Abbott, 1982, 1983; Cecile and Abbott, 1989), a 60 km by 30 km west-trending structural domain anomalous within the northwest-trending structural grain of the MacKenzie Mountains. The MacMillan Fold Belt (MFB) lies

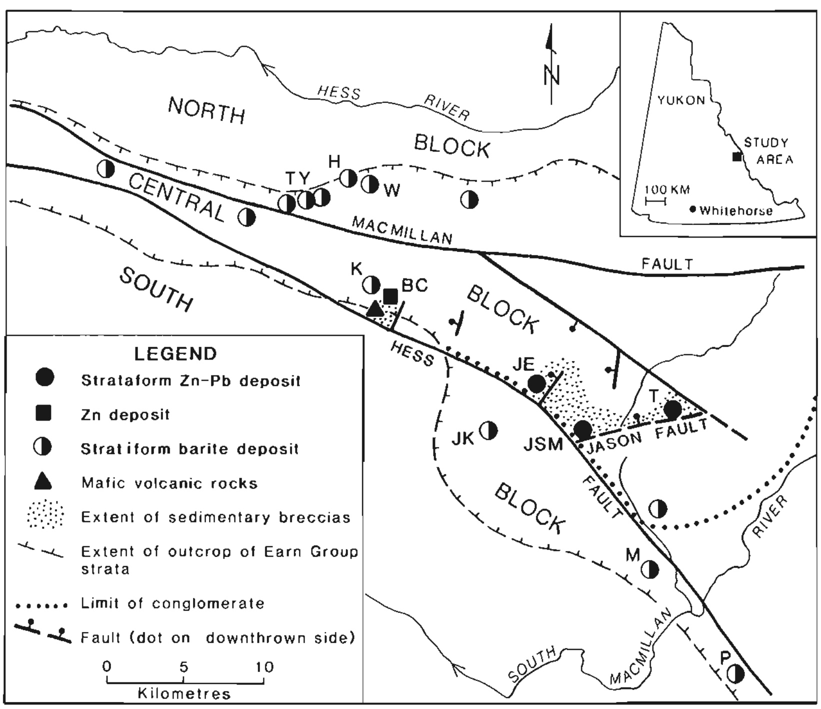

Figure 1. Distribution of Devonian age faults, sedimentary breccias, stratiform Zn-Pb- barite, stratiform barite deposits and volcanic rocks in the MacMillan Pass area. The North, Central and South blocks are separated by the MacMillan and Hess faults respectively. Zinc-lead prospects noted on the map are Boundary Creek (BC), Jason End zone (JE), Jason South/Main (JSM), and Tom (T). Barite prospects noted on the map are Tyrala (TY), Hess (H), Walt (W), K (Kobuk), JK (JK), Gary (G), Moose (M), and Pete (P). Inset map shows location of MacMillan Fold Belt.

within the Selwyn Basin about 100 km west of the present location of the Ordovician to Silurian carbonate shelf edge. Strata exposed within the MFB include Upper Proterozoic, Cambrian and Ordovician sandstone, shale and limestone; Ordovician to Devonian shale, chert, limestone and minor mafic volcanic rocks of the Road River Group; and Devonian shale, chert, chert conglomerate and sandstone of the Lower Earn Group. These miogeoclinal strata are weakly metamorphosed to prehnite-pumpellyite grade (Read, 1988) and cut by granitic stocks of Late Cretaceous age (Abbott, 1982).

The unusual Mesozoic structural trends of the MFB reflect older Devonian age faults according to Abbott (1982) who divided the MFB into three tectonostratigraphic domains (North, Central and South blocks) (Fig. 1) based on style of Mesozoic structure and Silurian-Devonian stratigraphy. The Central Block is the most structurally complex and dominated by west-trending tight to isoclinal folds and steep contractional faults. The Walt barite deposit occurs along the southern margin of the North block, a domain cut by an array of south-directed thrust faults (Fig. 2). The South Block has a simple structure of west-trending open folds. A thick conglomeratic middle turbidite member occurs only in the Central Block and southern margin of the North Block and reflects the extent of a Devonian age graben (Abbott, 1982). Also restricted to this graben are Silurian and Devonian volcanic rocks, Upper Devonian faults, and Late Devonian stratiform zinc-lead deposits (Fig. 1).

GEOLOGY OF THE EASTERN NORTH BLOCK

Closely spaced southerly directed imbricate thrust faults in the North Block repeat Proterozoic or Lower Cambrian to Upper Devonian basinal strata. These thrust faults trend northeast, dip moderately to steeply northwest and intersect at a shallow angle a west-trending belt of Upper Devonian chert conglomerate along the southern boundary of the block. This zone of transition from a northeast-trending structure to a west-trending structure in the southernmost North Block and Central Block is interpreted to represent the position of a fault, active during the Devonian and reactivated during Mesozoic deformation (Abbott, 1982). Abbott refered to this reactivated old fault zone as the MacMillan fault zone. The depositional edge of coarse clastic sediments along this graben margin is defined by a pinch-out of Upper Devonian conglomerate north of the southernmost North Block.

Proterozoic to Silurian strata in the northern and western portions of the North Block are prevalent at higher structural levels. Chert and siliceous shale of the lower member of the Lower Earn Group occur generally at lower structural levels along the southern margin of the western North Block and in the eastern North Block. Conglomerate, siltstone and siliceous shale of the middle and upper members of the Lower Earn Group are exposed within tight west-trending folds along the southern margin of the North Block.

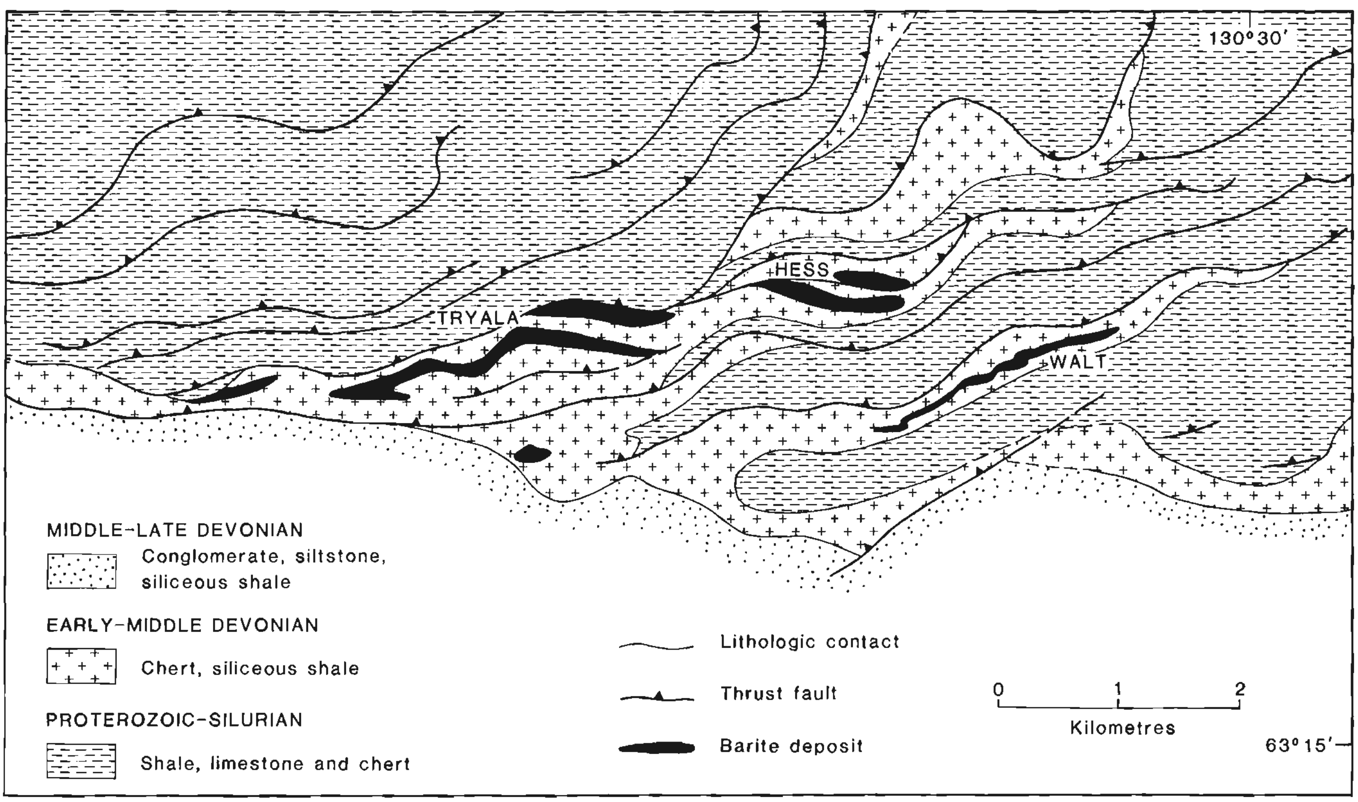

Figure 2. Geological map of a southern portion of the North Block showing distribution of stratiform barite deposits.

BARITE DEPOSITS

Thirteen stratiform barite deposits of Middle to Late Devonian age occur within the MacMillan Fold Belt (Fig. 1). Within the South and Central Blocks, stratiform barite deposits typically occur within siliceous shale of the upper member of the Lower Earn Group (unit uDpt of Abbott, 1983; Unit 3B of Carne, 1979). Barite deposits in the North Block and northernmost western Central Block, however, occur only within siliceous shale and chert of the lower member of the Lower Earn Group (unit emDpt of Abbott, 1983). Within the North Block, eight barite deposits occur within four separate thrust panels over a strike length of 13 km (Fig. 1, 2). Whether or not these barite deposits were part of a single barite body prior to faulting is unclear. The Walt barite deposit is on the eastern side and at the lowest structural level of the major cluster of barite deposits in the western part of this belt of barite deposits (Fig. 1).

GEOLOGY OF WALT BARITE DEPOSIT

The Walt deposit occurs near the top of a northwest-dipping thrust panel consisting of Cambrian to Devonian strata, near the thrust fault contact with overlying Cambo-Ordovician strata (Fig. 3, 4). Strata hosting the Walt deposit form a moderately northwest-dipping, generally homoclinal sequence. Locally, these strata are deformed into steep west-northwest plunging folds overturned to the southwest causing an en echelon map distribution of baritic occurrences (Abbott, 1983) (Fig. 3).

The Walt deposit occurs within interbedded chert and siliceous shale of the lower member of the Earn Group. Chert is black, occurs as black weathering beds 2 to 10 cm thick, and is interbedded with medium grey weathering black siliceous shale. This chert sequence overlies grey to tan weathering platy silty limestone, grey weathering calcareous black shale and grey weathering massive limestone beds with chert granules that occur downslope to the south of the Walt deposit (Fig. 3, 5). These lithologies are recessive and form scree-covered slopes except for low outcrops of limestone beds.

The Walt deposit is a concordant body, up to 60 m thick and continuous over at least 2 km, of cliff-forming carbonate bodies that are flanked and overlain by more recessive barite strata (Fig. 3). The deposit forms an east-northeast-trending ridge, the north slope of which is a dip-slope of barite beds (Fig. 4). Carbonate units within the barite body yielded conodonts of Eifelian to Givetian age (Middle Devonian) (Dawson and Orchard, 1982). XRD analysis of 4 samples from the Walt deposit reported by Dawson and Orchard (1982) identified barite, barytocalcite, witherite, calcite, and gypsum.

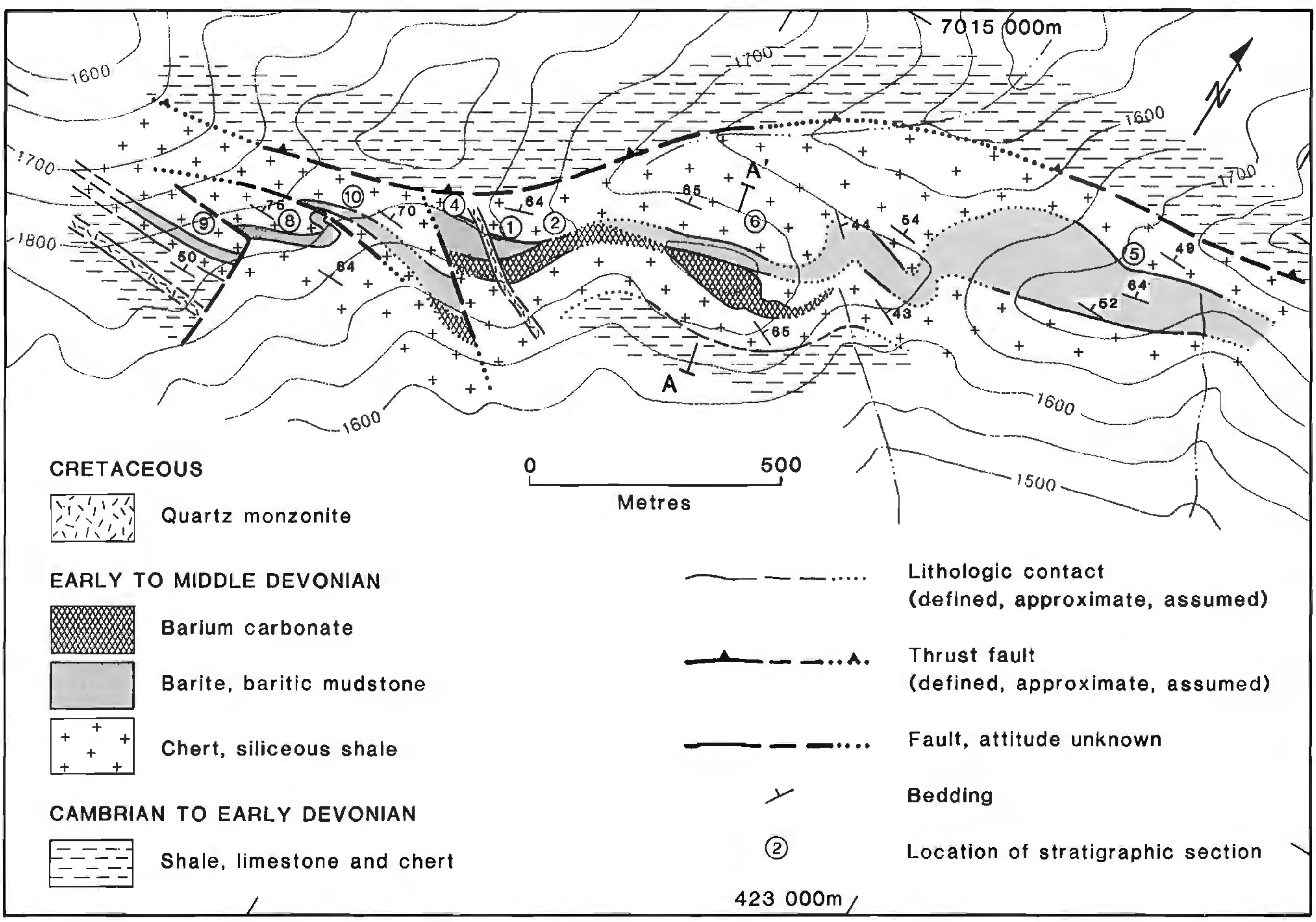

Figure 3. Geological map of the Walt property. Location of the thrust fault based on Abbott (1983).

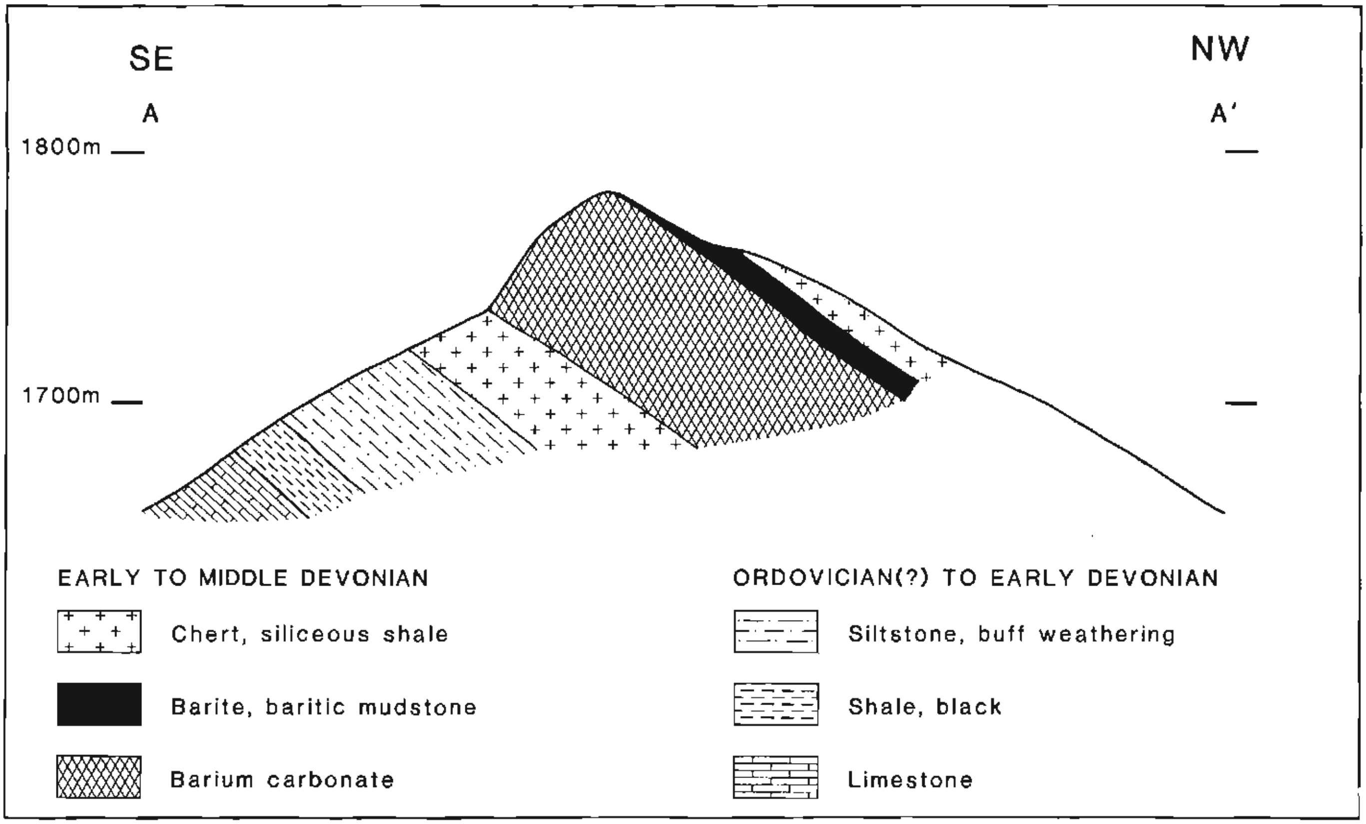

Figure 4. Structural cross-section across the Walt deposit. Location of section indicated in Figure 3.

BARITE AND BARIUM CARBONATE LITHOLOGIES

The barite and barium carbonate strata of the Walt deposit are divided into barite and barium carbonate facies on the basis of composition, and into subfacies using textural criteria (Fig. 5).

Barite Facies

The barite facies is much more extensive than the barium carbonate facies within the Walt deposit, and is the dominant facies within barite deposits throughout the MacMillan Fold Belt.

Laminated barite subfacies

Sequences of finely laminated dark grey weathering barite up to 25 m thick form recessive slopes covered with platey scree and occasional small low outcrops. Barite laminae are 0.5 to 2mm thick, dark grey in colour, and interbedded with minor black carbonaceous chert laminae. Laminated barite is organic-rich, has a fetid smell when broken, and, along with massive barite subfacies rock, is the economically important portion of the Walt deposit. Laminated barite occurs with massive barite, baritic mudstone, tan-altered laminated barite and massive barium carbonate (Fig. 6d, 7c).

Massive barite subfacies

Massive to faintly banded pale grey to black barite occurs as scree-covered slopes and small low outcrops. Barite is medium-grained and sugary in texture. Massive barite up to 10 m thick occurs interbedded with chert, laminated barite, tan-altered barite and baritic mudstone and is an important and widespread facies. Contacts between massive barite and chert are sharp. Massive barite is the common form of barite at the contact between barite and barium carbonate facies. In this transition zone, crudely interbanded massive barite and massive carbonate, irregular or elongate pods of barium carbonate aligned along bedding occur within massive barite, and irregular pods, irregular vein-like networks and discontinuous beds of massive barite occur within massive barium carbonate (Fig. 7b,d). Textural relationships suggest replacement of massive barite by massive barium carbonate.

Baritic mudstone subfacies

Baritic mudstone subfacies consists of cleaved baritic rocks which include fine- to medium-grained platey barite (argillaceous barite), fine grained compact baritic rock with planar fracture (baritic mudstone), to fissile baritic rock (baritic shale). These strata have a distinctly lower specific gravity than laminated or massive barite, presumably due to the presence of detrital quartz and clays. These medium

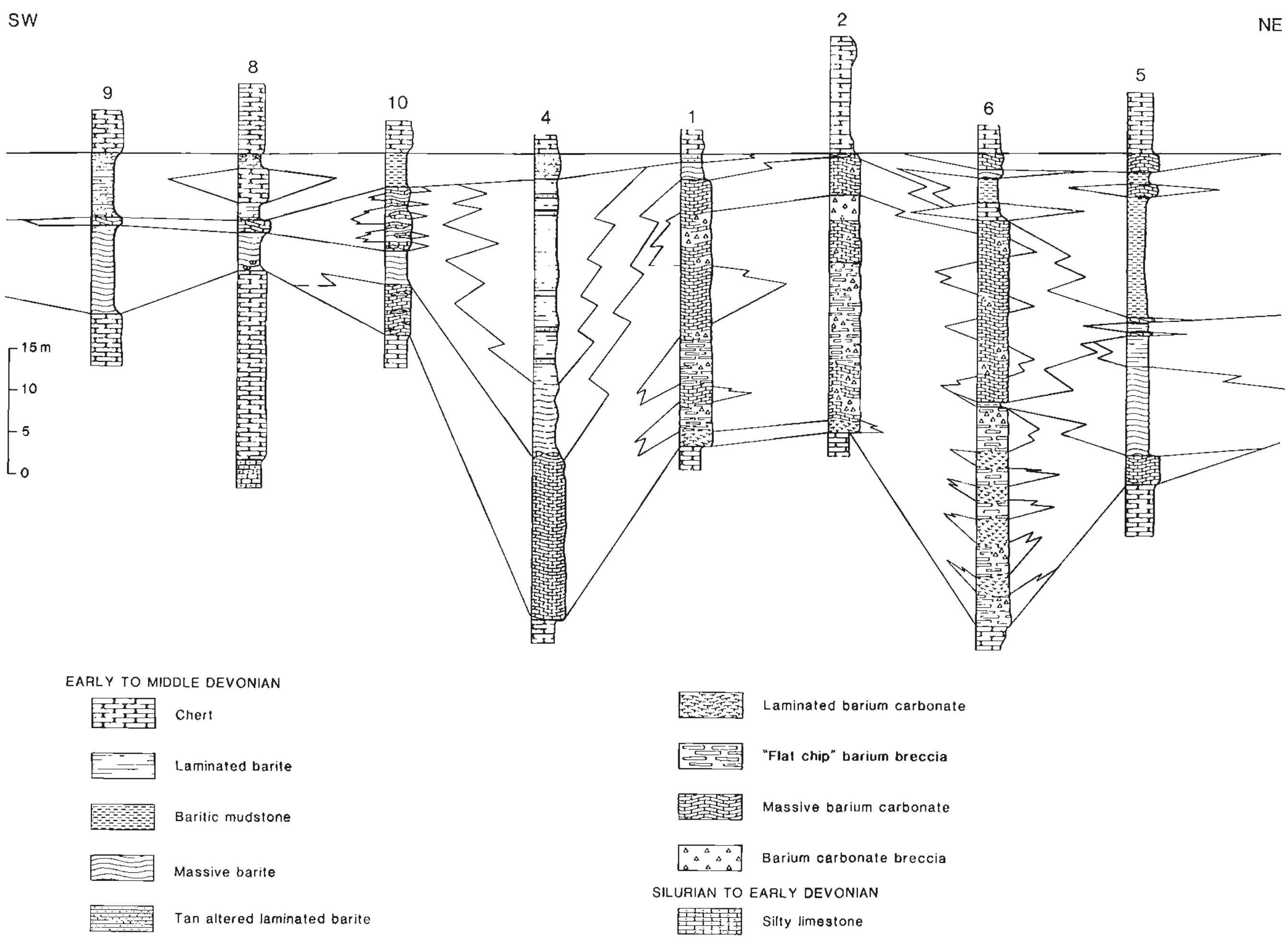

Figure 5. Stratigraphic fence diagram of the Walt deposit. Location of stratigraphic sections indicated in Figure 3.

to dark grey weathering baritic rocks are recessive and form scree-covered slopes. Baritic mudstone subfacies is interbedded most commonly with laminated and massive barite near the top of the barite horizon. At the nearby Hess property (Fig. 2), cross-laminated baritic sandstone beds up to 1 cm thick are interbedded with baritic mudstone.

Tan-altered laminated barite subfacies

Tan to buff weathering, laminated baritic rock caps the southwestern portion of the barite body on the Walt property. The mineralogy of this rock is not known at present but likely includes barite, quartz and an iron bearing phase such as a ferroan carbonate. Locally, tan-altered laminated barite is noted to be transitional between laminated barite and massive carbonate. This tan-altered rock forms scree slopes and low outcrops.

Barium Carbonate Facies

A barium carbonate-rich unit, henceforth referred to as the barium carbonate facies, crops out over 800 m of strike length and is more limited in extent than the barite facies. A barium carbonate composition is indicated by high specific gravity and effervescence in dilute HCl. This field identification is supported by XRD analyses of samples from the Walt property (Dawson and Orchard, 1982). Elsewhere in the MacMillan Fold belt, barium carbonate facies is not common but is well exposed on the Hess property a kilometre northwest of the Walt within a higher structural panel (Fig. 2); it is possible that both the Hess and Walt deposits represent faulted portions of the same Middle Devonian barite deposit. Barium carbonate have also been recognized at the Tea barite deposit (Lydon et al., 1979) and the Gary North barite deposit (I.R. Jonasson, pers. comm.).

Massive barium carbonate subfacies

Massive, fine to medium grained barium carbonate is the principal component of cliff-forming carbonate bodies and forms units up to 20m thick (Fig. 6a,b). This grey massive carbonate weathers a pale grey to cream, rarely dark grey, pitted surface with an irregular pattern of low ridges. Massive barium carbonate is associated with all other barium carbonate types but is most abundant at the upper and lateral margins of the barium carbonate body, in a transitional position between carbonate breccias and barite facies rock. The contact between massive barium carbonate and barite facies,

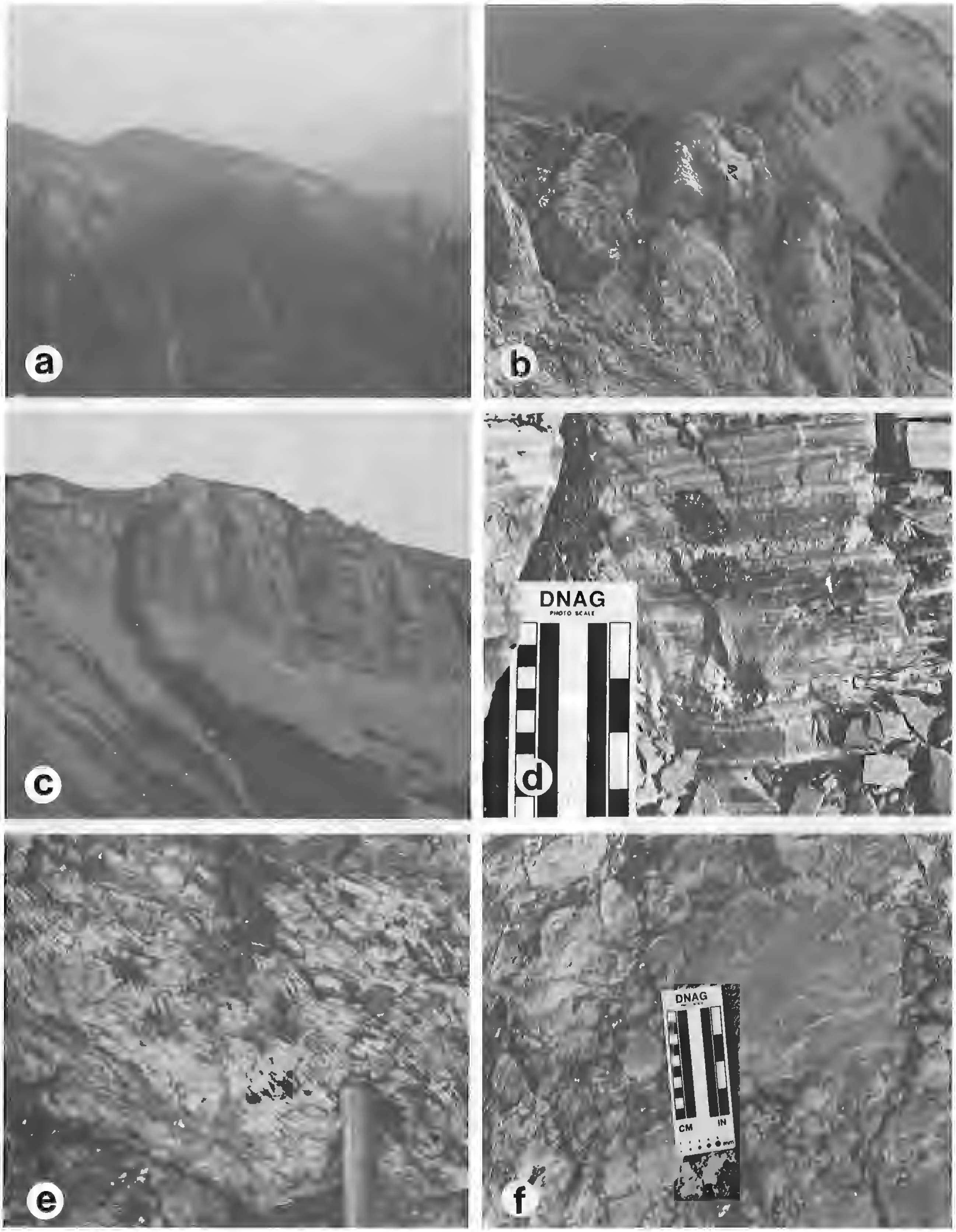

Figure 6. Photographs of the Walt (Cathy) barite deposit. (a) Aerial photograph looking north at pale grey barium carbonate and barite outcropping along ridge; (b) ridge-forming resistant and blocky outcrops of pale grey barium carbonate, looking eastward along ridge; (c) blocky outcrop of massive barium carbonate replacement within the vent complex; (d) pale to medium grey laminated barite partly replaced along bedding planes by pale grey to white bands of barium carbonate; (e) "flat chip" barium carbonate breccia; (f) barium carbonate breccia composed of pale grey angular to subrounded barium carbonate clasts relatively recessive with respect to a matrix of brown-grey weakly silicified barium carbonate.

typically massive barite, is interdigitating and complex (Fig. 6d, 7b, 7c, 7d). Irregular pods or vein-like networks of black massive barite commonly occur within massive barium carbonate near the contact with barite facies, whereas adjacent massive barite hosts irregular pods or elongate bodies of barium carbonate aligned with bedding. Replacement of barite facies by massive barium carbonate is suggested by (1) contacts between massive carbonate and barite facies that cut bedding, (2) carbonate mound-like forms that have 'eaten' into overlying baritic strata, (3) the lack of clear bedding textures in the massive carbonate, and (4) association of carbonate subfacies with brecciation textures in the core of the complex .

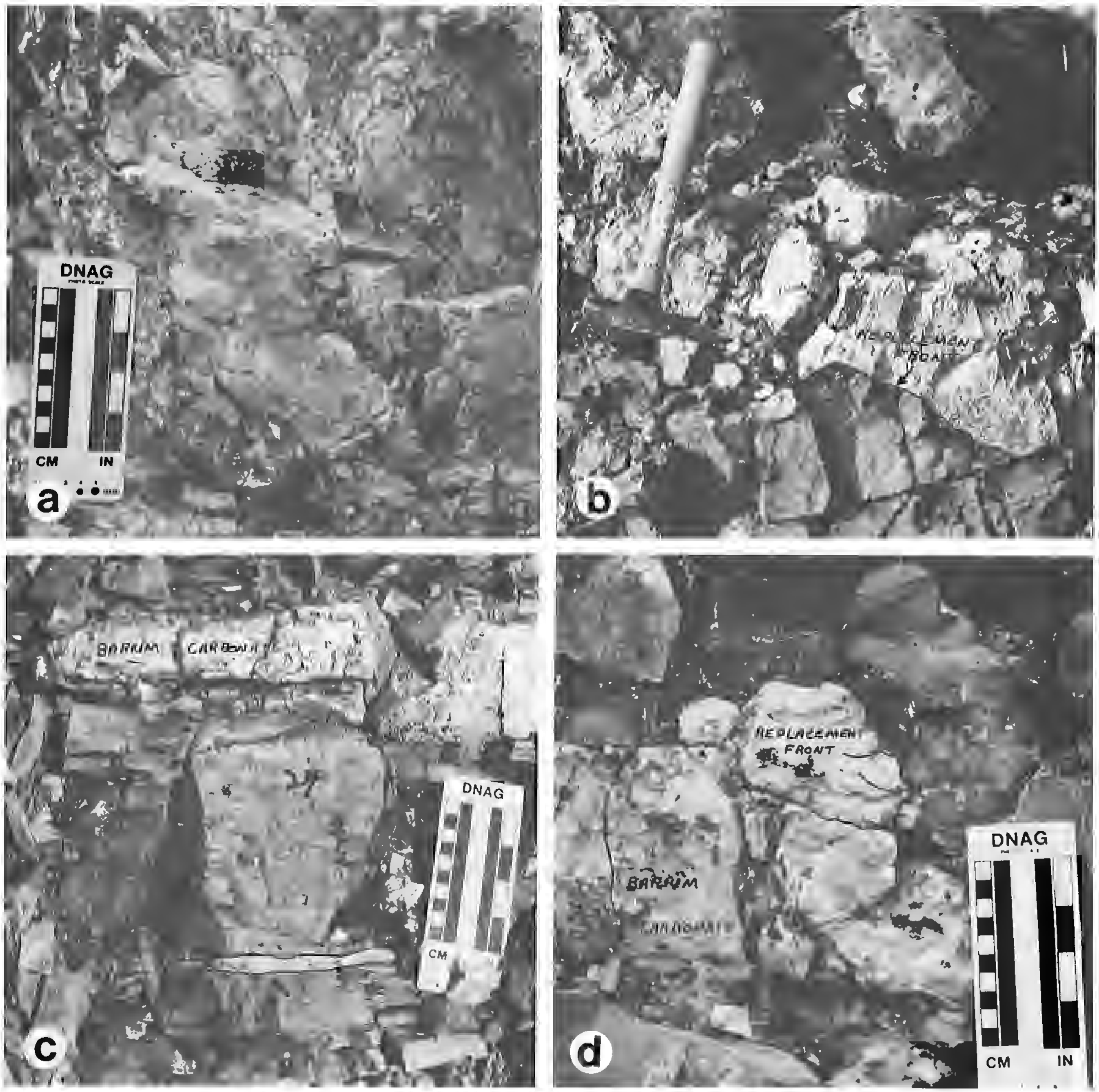

Figure 7. Photographs of the Walt deposit. (a) Barium carbonate breccia composed of pale grey subrounded to angular barium carbonate clasts set in a medium grey matrix of barium carbonate; (b) irregular replacement contact between medium grey massive to diffusely banded recrystallized barite (bottom half) and pale grey massive barium carbonate (top half); (c) diffusely laminated medium grey barite with bands of pale grey barium carbonate that has replaced barite along bedding plances; (d) medium grey massive crystalline barite replaced by pale grey barium carbonate along an irregular replacement front.

Laminated barium carbonate subfacies

This subfacies is medium to dark grey, weathers a pale grey to cream colour, and is composed of fine- to medium-grained laminae of barium carbonate, barite and quartz up to 2 mm thick (Fig. 7d). Laminated barium carbonate forms beds up to 3 m thick that are cut by or interbanded with carbonate breccias near the base of the large cliff-forming carbonate bodies (Fig. 1). In surface outcrops, this subfacies weathers to a finely ribbed texture reflecting compositional differences between bands. Resistant bands appear more barite- and quartz-rich whereas recessive bands are more barium carbonate-rich. Locally, banded barium carbonate rock is transitional to laminated barite and laminae are noted to pass though a discordant reaction front between laminated barite and barium carbonate.

Flat chip barium carbonate breccia subfacies

Flat chip breccia is an important subfacies within the lower part of the cliff-forming carbonate bodies where units up to 5m thick are interbanded with laminated barium carbonate and locally cut by discordant zones of barium carbonate breccia. This distinctive form of breccia is composed of angular to subrounded rectangular clasts or 'flat chips' 1 to 2 cm thick and 3 to 8 cm long aligned with bedding (Fig. 6e). Rotation of clasts is uncommon. Both clast and matrix are composed of fine- to medium-grained grey barium carbonate that weathers pale grey to cream. Clast shape distinguishes this breccia type from the sub-equant clasts of the barium carbonate breccia subfacies . Because cherts underlying flat chip barium carbonate breccia subfacies at the base of the barium carbonate facies weather recessively, it is unclear whether this unit is also brecciated.

Barium carbonate breccia subfacies

Irregular discordant, locally concordant breccias up to 3m thick occur within the core of the cliff-forming barium carbonate body. Clasts are angular to subrounded, sub-equant in shape, and typically 2-3 cm in diameter (Fig. 6f, 7a) although clasts up to 10 cm across were noted. The composition of matrix and clasts is variable as clasts can be darker or lighter, and more or less resistant than adjacent matrix. With decreasing contrast between clast and matrix, this breccia is gradational into massive barium carbonate.

DISCUSSION

Formation of barite and barium carbonate

The laterally extensive, concordant nature of the barite facies, locally preserved bedded character, and sharp contacts with interbedded chert and shale suggest a sedimentary origin for the barite (Fig. 8). The black chert and siliceous shale host to the barite body indicate hemipelagic and pelagic deposition within a basin starved of clastic input. The black colour combined with the absence of bioturbation indicate sedimentation below a reduced water column. Argillaceous barite and baritic mudstone may represent resedimented baritic material based on the association of similar strata with thin cross-bedded baritic sandstone within barite facies at the nearby Hess barite prospect.

Field evidence suggests the barium carbonate facies formed by the replacement of portions of the stratiform sedimentary barite facies such as laminated barite or baritic mudstone (Fig. 8). Subfacies zonation documented earlier can be related to progressive replacement of laminated barite by carbonate-rich hydrothermal fluids. Laminated barite, because it is the dominant facies type in deposits lacking barium carbonate facies, is considered the protolith. Transitional contacts between laminated barite and massive barite suggest recrystallization and locally partial oxidation of organic matter. Massive barite was completely replaced by massive barium carbonate along irregular reaction fronts. Laminated barium carbonate indicates textural preservation and a different replacement process that may involve tan-altered laminated barite as incipient carbonatization followed by complete barium carbonate replacement of barite.

The replacement of barite by barium carbonate may be related to vent processes during Middle Devonian deposition of the barite, or to metasomatism adjacent to younger faults or intrusions. There is no evidence for association of replacement with younger faulting at the Walt deposit. It is unlikely that replacement is related to Cretaceous intrusions as a felsic dyke of probable Cretaceous age cuts both barite facies and barium carbonate facies in the central part of the Walt deposit and there is neither barium carbonate alteration of barite adjacent to the dyke nor carbonate within the dyke. However, the nearby Tom and Jason Zn-Pb SEDEX deposits include carbonate zones interpreted to be vent complexes or upflow zones of hydrothermal fluids that formed the stratiform barite and sulphide lenses; these ferroan carbonate bodies formed by replacement of the stratiform barite and sulphides (Turner et. al, 1989a; Goodfellow et al., 1989). Therefore, we speculate that barium carbonate replacement of barite at the Walt deposit was related to vent processes during Middle Devonian deposition of the stratiform barite (Fig. 8).

There are several possible origins of the barium carbonate breccias such as (1) sedimentary olististromes (Dawson and Orchard, 1982), (2) brecciation associated with younger tectonic faulting, (3) brecciation associated with younger intrusions, or (4) brecciation during formation of the barite deposit. The barium carbonate composition, gradational contacts with massive barium carbonates, and the absence of similar carbonates elsewhere within the lower member of the Lower Earn Group except associated with barite deposits argue against a sedimentary origin for the breccias. There is no field evidence that the breccias are related to faults, nor is brecciation associated with a felsic dyke that intrudes the central part of the Walt deposit (Fig. 3). However, like the Walt deposit, the core of the carbonate vent complex at the Jason SEDEX Zn-Pb deposit is brecciated (Turner, 1986); at the Jason deposit, breccias are adjacent to a syndepositional fault. We therefore believe that brecciation was associated with barium carbonate replacement of the barite during the Middle Devonian. Brecciation may have resulted from CO_2 efforvescence in the hydrothermal fluids, hydraulic fracturing during hydrothermal fluid discharge, or tectonic movement on a syndepositional fault, though such a fault has not been recognized.

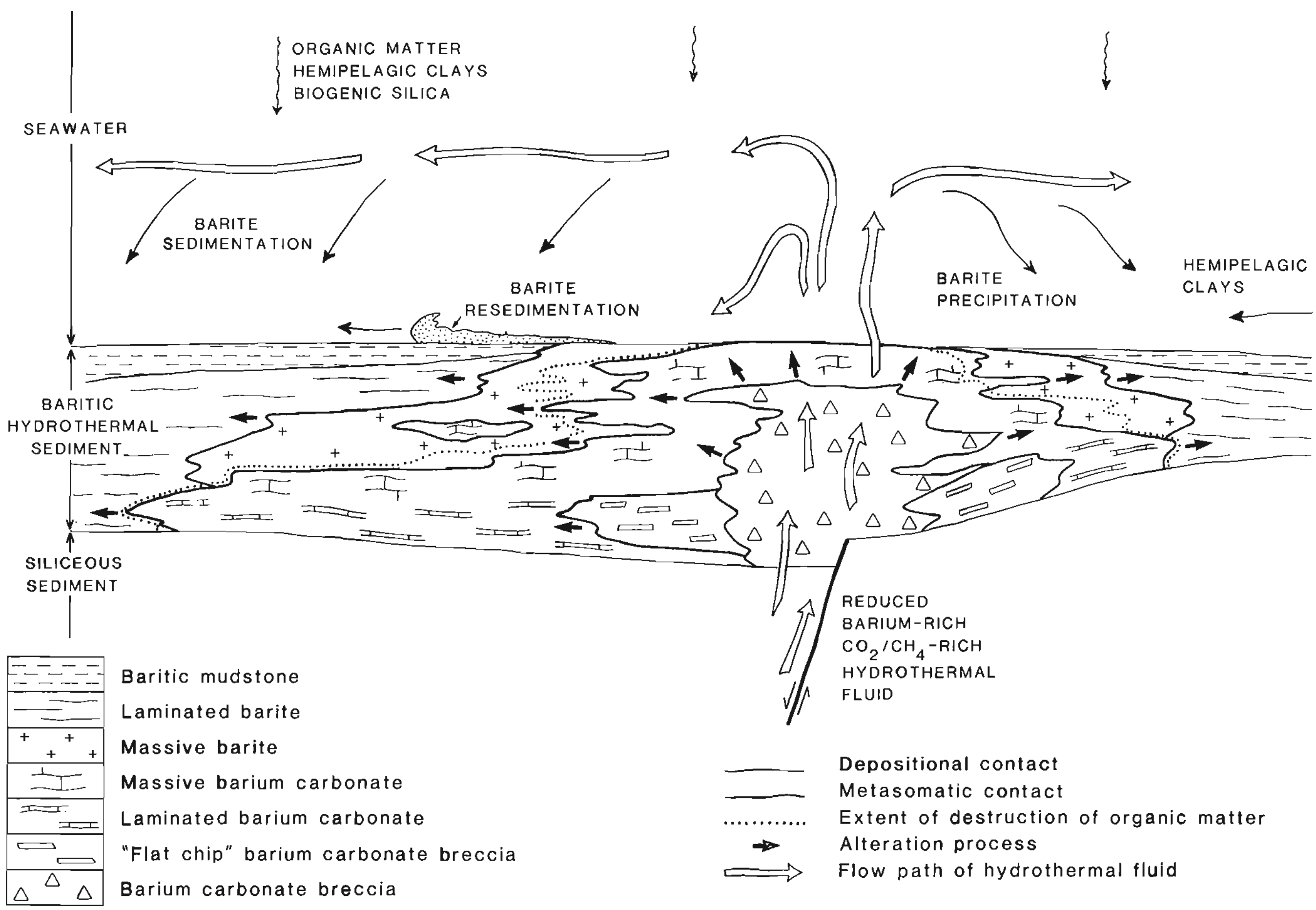

Figure 8. Schematic representation of processes during formation of Walt deposit.

GEOCHEMISTRY OF REPLACEMENT PROCESSES

Although detailed mineralogical and isotopic studies have not been completed, field textural relationships indicate clearly that sedimentary barite bas been replaced by barium carbonate. The following are speculations on possible processes by which this replacement may have taken place. The replacement of barite by barium carbonates probably involves initial dissolution of barite and reprecipitation of barium as a carbonate. Although barite has a prograde solubility up to 300 decrees C in greater than 1 molal NaCl solutions (Blount, 1977), the wholesale dissolution of barite most likely involved sulphate reduction during reaction with hydrothermal fluids buffered at an $^{f}O_2$ below th SO_4/H_2S redox boundary. This reation has been shown experimentally to be important at temperatures greater than 250 degrees C in the presence of organic matter (Kiyosu 1980). Since basinally derived hydrothermal fluids probably equilibrated with a carbonaceous sedimentary sequence, the fluid $^{f}O_2$ was likely controlled by the carbonate-methane buffer. For this reaction to be important, there must be sufficient amounts of methane (or other organic compounds) to reduce the barium sulphate. Furthermore, the products of this reaction are isotopically heavy sulphide and light carbon. It is expected therefore that the barium carbonates in the core of the complex will have very negative delta ^{13}C values. The lack of sulphides at the Walt deposit precludes an evalution of the isotopic composition of sulphur.

COMPARISON TO SEDEX Zn-Pb-BARITE DEPOSITS

A similar relationship of massive and brecciated carbonate replacing bedded baritic strata occurs at the nearby Jason and Tom stratiform Zn-Pb-barite deposits (Turner, 1986; Turner et al., 1989a; Goodfellow et al., 1989). In these cases, ferroan carbonate associated with pyrrhotite, galena and pyrite cut earlier-formed sedimentary barite, chert, sphalerite, galena and pyrite. Barium carbonates are common near the replacement front of carbonate replacing baritic strata (Garder and Hutcheon, 1985; Turner, unpublished data). At Jason, this carbonate body is cored by a breccia body along a syndepositional fault. At both Tom and Jason, the carbonate body is interpreted to represent a paleovent complex or hydrothermal upflow zone and the laminated baritic strata to be hydrothermal sediments lateral to the vent.

The major difference between stratiform Zn-Pb-barite deposits in the MacMillan Fold Belt and the Walt deposit is the lack of base metals in the case of the latter. Unlike the Tom and Jason deposits, the sedimentary barite facies at the Walt deposit do not contain laminated sphalerite, galena or pyrite. Furthermore, at Tom and Jason, ferroan carbonate is the dominant replacement mineral along with base metal sulphides. The lack of base metals at the Walt in both sedimentary and replacement facies may be due to

(1) formation during a period of oxygenated and therefore low H_2S botton water conditions (e.g. Goodfellow, 1987), or (2) a low base metal content of the hydrothermal fluids. The carbonaceous, non-bioturbated chert and mudstone that is host to the Walt deposit were deposited under reduced botton water conditions and does not support the former hypothesis. The latter possibility is supported by the predominance of barium carbonate over ferroan carbonate, and the lack of sulphides in the replacement vent complex (despite the possible formation of H_2S during sulphate reduction). The formation of the Walt deposit north of the MacMillan graben and remote from mafic volcanic rocks and presumed areas of maximum geothermal gradients, and the absence of pervasive silicification which characterizes vent complexes associated with stratiform Zn-Pb deposits, indicates further that the Walt deposit formed from low-temperature, probably low-f_{O_2}, Ba-rich, non-metalliferous hydrothermal fluids. In addition to being situated distal to the center of hydrothermal activity, the Walt deposit of Middle Devonian age also predates the major episode of hydrothermal activity in the Late Devonian, and therefore formed during the early stages of geothermal activity in the district.

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Boundary Creek zinc deposit (Nidd property), MacMillan Pass, Yukon: sub-seafloor sediment-hosted mineralization associated with volcanism along a late Devonian syndepositional fault

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Turner, R.J.W. and Rhodes, D. Boundary Creek zinc deposit (Nidd property), MacMillan Pass, Yukon: sub-seaflow sediment-hosted mineralization associated with volcanism along a late Devonian syndepositional fault; in Current Research, Part E, Geological Survey of Canada, Paper 90-1E, p. 321-335, 1990.

Abstract

At the Boundary Creek zinc deposit, quartz, ferroan carbonate, pyrite, minor sphalerite and trace galena occur within conglomerate, muddy conglomerate, diamictite and volcanic rocks adjacent to a late Devonian syndepositional fault. Hydrothermal minerals occur as cement within conglomerate and as replacement of diamictite at shallower depths below the paleo-seafloor, as veins and replacement of all rock types at greater depths, and as large banded sphalerite-carbonate veins and carbonate breccias locally along the syndepositional fault. This large tonnage, very low grade zinc deposit is interpreted to represent an epigenetic variant of sedimentary exhalative (SEDEX) deposits that formed synchronously with the nearby Zn-Pb SEDEX deposits. Boundary Creek mineralization occurred during or following the waning stages of basaltic volcanism. A greater role of magmatism than previously recognized is indicated in the formation of SEDEX deposits in the MacMillan Fold Belt.

Résumé

Dans le gisement de zinc de Boundary Creek on trouve du quartz, du carbonate de fer, de la pyrite, un peu de sphalérite et des traces de galène à l'intérieur d'un conglomérat, d'un conglomérat boueux, d'une diamictite et des roches volcaniques adjacentes à une faille synsédimentaire du Dévonien supérieur. Des minéraux hydrothermiques se manifestent sous forme de ciment au sein de conglomérats et de remplacement de la diamictite à faibles profondeurs au-dessous du paléofond marin, sous forme de veines et de remplacement de tous les types de roches à des plus grandes profondeurs ainsi que sous forme de grandes veines zonées de sphalérite et carbonate et de brèches carbonatées par endroits le long de la faille synsédimentaire. D'après les auteurs, ce gisement à fort tonnage et à très faible teneur en zinc représente une variation épigénique de gisements sédimentaires exhalatifs (SEDEX) qui se sont formés en même temps que des gisements de Zn-Pb de type SEDEX avoisinants. La minéralisation de Boundary Creek s'est produite au cours des phases de diminution de l'activité volcanique de nature basaltique ou après ces phases. La formation des gisements de type SEDEX dans la zone de plissement de MacMillan indique que le magmatisme a joué un rôle plus grand que celui reconnu antérieurement.

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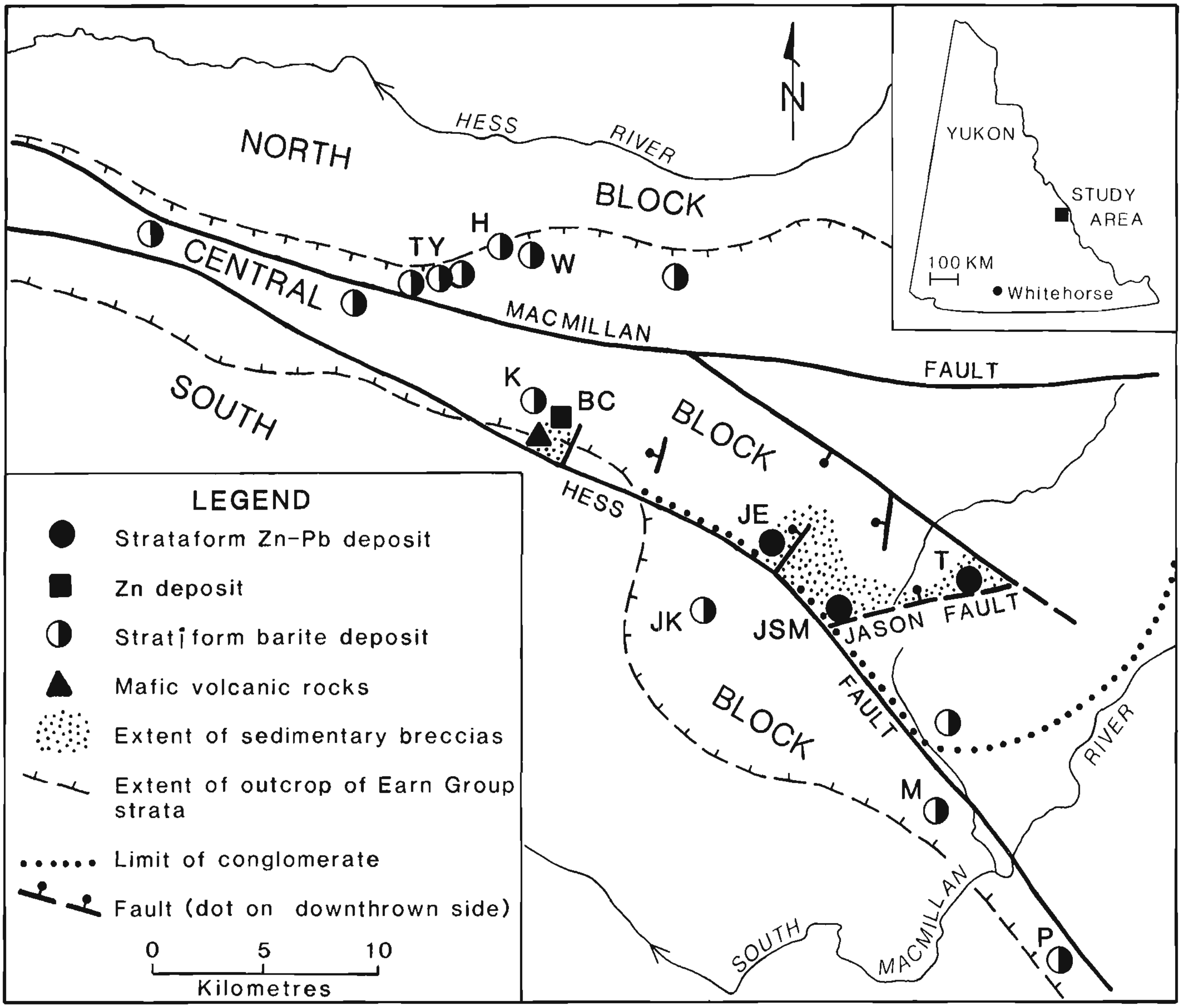

Figure 1. MacMillan Fold Belt showing distribution of Devonian age features such as syndepositional faults, sedimentary breccias, volcanic rocks, base metal deposits, and stratiform barite deposits. The extent of ferroan carbonate alteration within sedimentary strata of the Earn Group and Road River Group is approximately coextensive with the distribution of sedimentary breccias. The Central Block of Abbott (1982) is bound on the north by the Macmillan fault and on the south by the Hess fault. Zinc-lead prospects noted on the map are Boundary Creek (BC), Jason End zone (JE), Jason South/Main (JSM), and Tom (T). Barite prospects noted on the map are Tyrala (TY), Hess (H), Walt (W), K (Kobuk), JK (JK), Gary (G), Moose (M), and Pete (P). Inset map shows location of MacMillan Fold Belt.

LOCATION AND SETTING

The Boundary Creek deposit on the Nidd property near MacMillan Pass (105 0/1), Yukon is located approximately 390 km northeast of Whitehorse (63°11′ N; 1050/1, 130°21′ W) (Fig. 1). The Boundary Creek deposit underlies valley bottom drift near the junction of the south-flowing Boundary Creek and an east-flowing tributary of the Hess River. The Nidd property was staked in 1976 by Cominco Ltd., and between 1982 and 1989 fourteen holes totalling 4665m have been drilled at the Boundary Creek prospect. Studies of the Boundary Creek deposit by Cominco staff include Hodson (1984), Rhodes and Murrel (1983), and Rhodes (1985, 1986). Access to the property is by an unsurfaced road which joins the North Canol Road at the MacMillan Pass airstrip and runs 25 km northwest to Boundary Creek.

REGIONAL SETTING

The Boundary Creek deposit is within the MacMillan Fold Belt (Abbott, 1982, 1983), a 60 km by 30 km structural domain with an anomalous westerly trend within the northwest-trending structural grain of the MacKenzie Mountains. The MacMillan Fold Belt (MFB) lies within Selwyn Basin about 100 km west of the present location of the Ordovician to Silurian carbonate shelf edge. Strata exposed within the MFB include Upper Proterozoic, Cambrian and Ordovician sandstone, shale and limestone; Ordovician to Devonian shale, chert, limestone and minor mafic volcanic rock of the Road River Group; and Devonian carbonaceous shale, chert conglomerate and sandstone of the Lower Earn Group. These miogeoclinal strata are weakly metamorphosed to prehnite-pumpellyite grade (Read, 1988) and are cut by granitic stocks of Late Cretaceous age (Abbott, 1982).

The unusual Mesozoic structural trends of the MFB reflect older Devonian age faults according to Abbott (1982) who divided the MFB into three tectonostratigraphic domains (North, Central and South blocks) based on style of Mesozoic structure and Silurian-Devonian stratigraphy. The Boundary Creek prospect, along with the Jason and Tom stratiform zinc-lead deposits occur within the Central Block (Fig. 1). The structure of the Central Block is dominated by west-trending tight to isoclinal folds and steep contractional faults, a more complex structure than the North or South blocks. Within the Central block, Lower Earn Group strata overlie basaltic flows and volcaniclastic rocks as young as late Middle Devonian. Abbott (1986) recognized three informal units within the Lower Earn Group: a lower member of carbonaceous chert, a middle turbidite member, and an upper member of carbonaceous siliceous shale. A thick conglomerate unit within the middle turbidite member occurs only in the Central Block and southern margin of the North Block and reflects the extent of a Devonian graben (Fig. 1) (Abbott, 1982). Also restricted to this graben are Silurian and Devonian volcanic rocks, Late Devonian faults, and Upper Devonian stratiform zinc-lead deposits.

STRUCTURE OF THE BOUNDARY CREEK AREA

The Boundary Creek prospect occurs a kilometre north of the west-northwest- trending Hess fault zone, a major structure that forms the southern boundary of the Central Block (Fig. 1). Nearby strata of the Central Block are deformed into west-trending tight folds cut by steep south-dipping reverse faults (Abbott, 1983). Late north-northeast-trending faults cut all other structures.

Poor surface exposure, lack of good stratigraphic marker units, and wide drill hole spacing to date cause some elements of our structural interpretation of the Boundary Creek prospect to be conjectural. Drill hole data indicate that strata trend west, dip steeply north, are deformed by a weak west-trending cleavage, and are cut by a series of faults. Based on stratigraphic top indicators such as graded bedding, cross laminations and bed assymmetries, altered and mineralized strata occur within a tight syncline overturned to the south (Fig. 2, 3,4,5). Lower Earn Group strata in the core of this fold are in fault contact to the north and south with Road River Group strata. This west-trending fold lies on trend with a syncline mapped a kilometre to the east (Abbott, 1983). The Boundary Creek syncline is complicated by a number of faults; correlation of faults between drillholes suggests a set of major west-trending, steeply north-dipping faults subparallel to the fold limbs (Fig. 2,5). Stratigraphic offset indicates reverse movement on several of these faults. North of the syncline, a southwest-trending anticline exposes rocks of the Road River Group (Fig. 2).

Syndepositional fault

A cataclastite, vein and breccia zone dipping steeply north is interpreted as a syndepositional fault zone (Fig. 4). A 9 m interval (260-268.5m) in drill 14 is composed of two cataclastic zones 1 and 3.5m wide that bracket and are overlain by highly altered rock with banded veins up to 50 cm thick. The cataclastic zones consist of foliated fine-grained quartz, clay, and ferroan carbonate that contain rounded to angular fragments of ankerite veins, fine-grained siderite and siliceous mudstone (Fig. 6a, 6b). Veins associated with the fault zone are rich in chalcopyrite and galena compared to the rest of the Boundary Creek deposit. Volcanic rocks adjacent to the fault zone are altered to a black chlorite, ferroan carbonate and quartz assemblage unique within the deposit. The fault zone juxtaposes volcanic rocks of the Earn Group against mudstone of the Road River Formation (Fig. 4); this structural elevation is reflected in the map pattern by a northward bulge of the upper contact with the Road River Group (Fig. 2). Reactivation of this fault, perhaps during the Mesozoic or Tertiary also accounts for some of this offset. In drillhole 13, a ferroan carbonate breccia zone (Fig. 6c) is interpreted as a transition zone between the fault and underlying Road River Group strata. The abundance of resedimented and locally derived coarse clastic rocks in the Lower Earn Group suggests major faulting during the Middle to Late Devonian. The northeasterly distribution of alteration zones and sulphides in the Boundary Creek area possibly reflects the trend of this syndepositional fault.

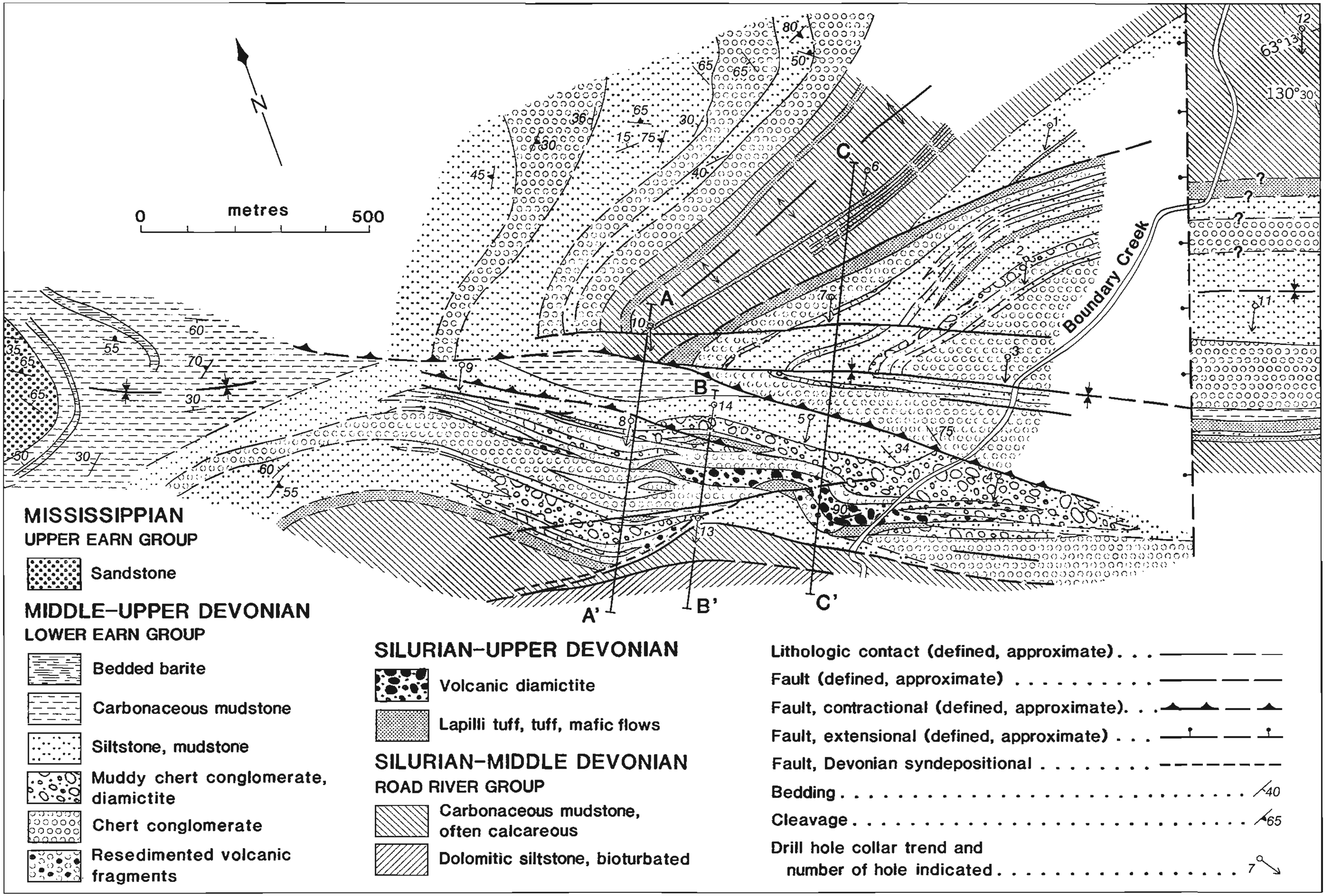

Figure 2. Generalized geological map of the Boundary Creek area. Location of cross sections (Fig. 4,5 and 6) are indicated.

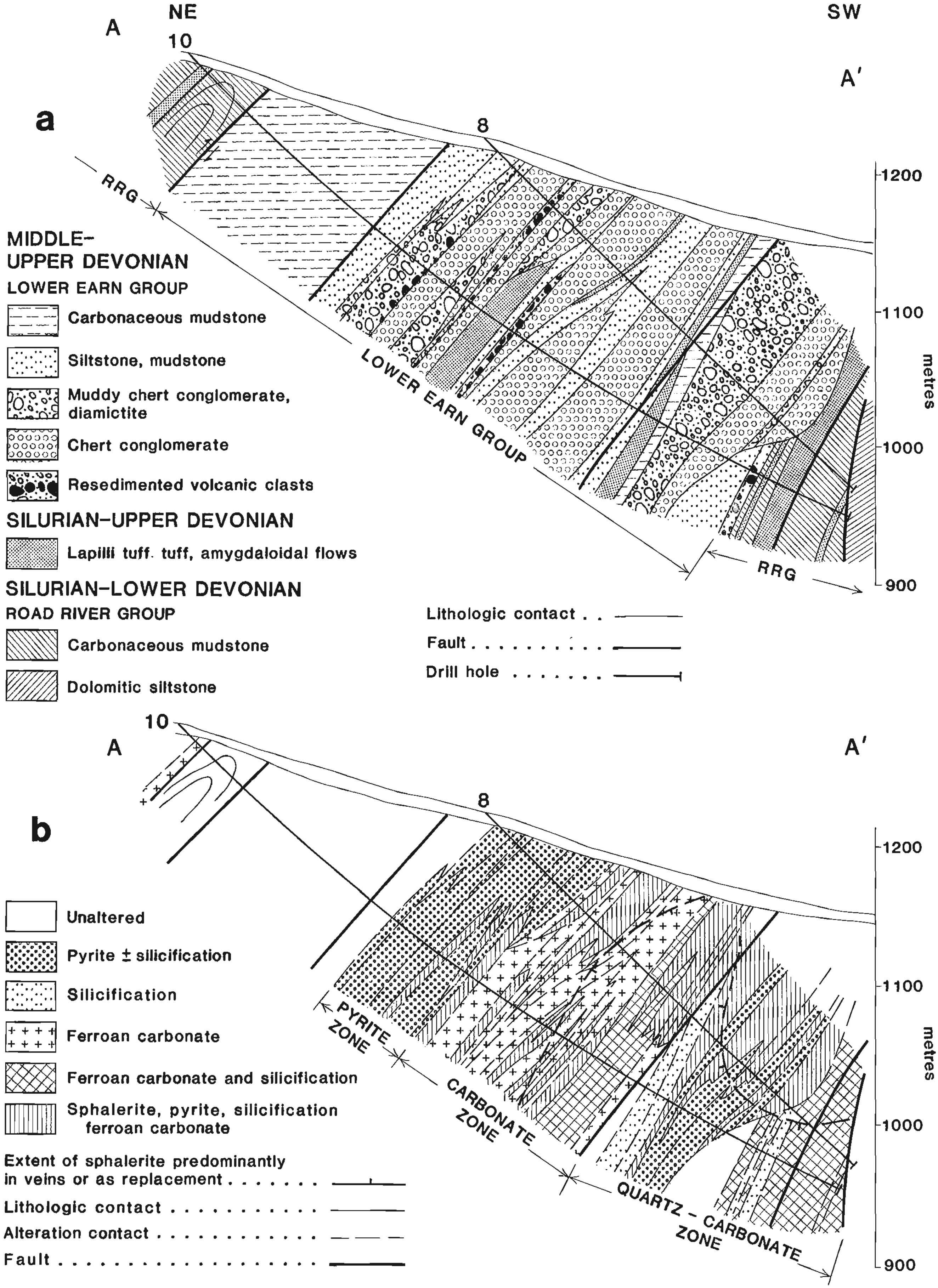

Figure 3. North-south structural cross-section A-A' through Boundary Creek deposit illustrating distribution of (a) lithologies and (b) alteration/mineralization types.

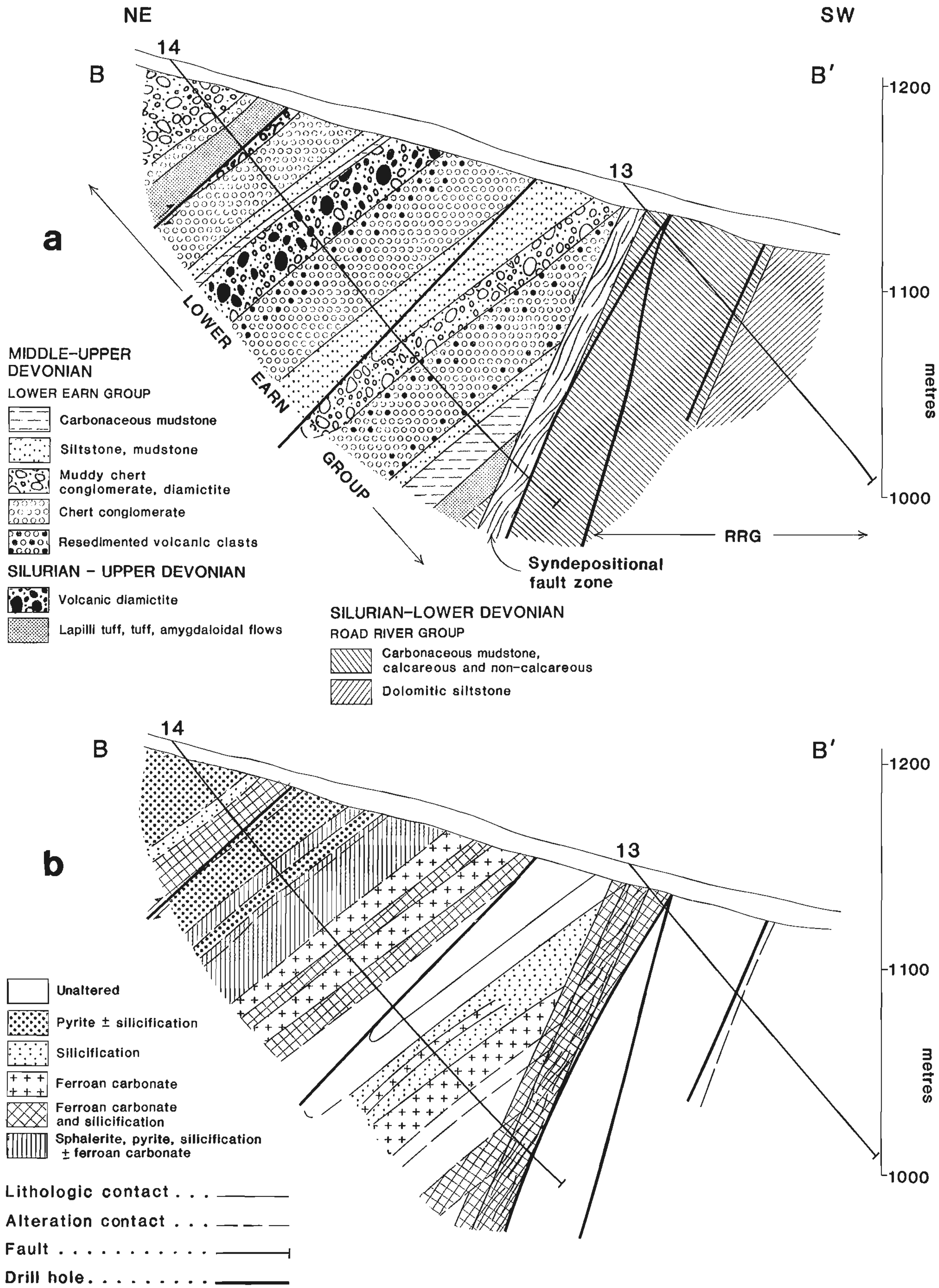

Figure 4. North-south structural cross-section B-B′ through Boundary Creek deposit illustrating distribution of (a) lithologies and (b) alteration/mineralization types.

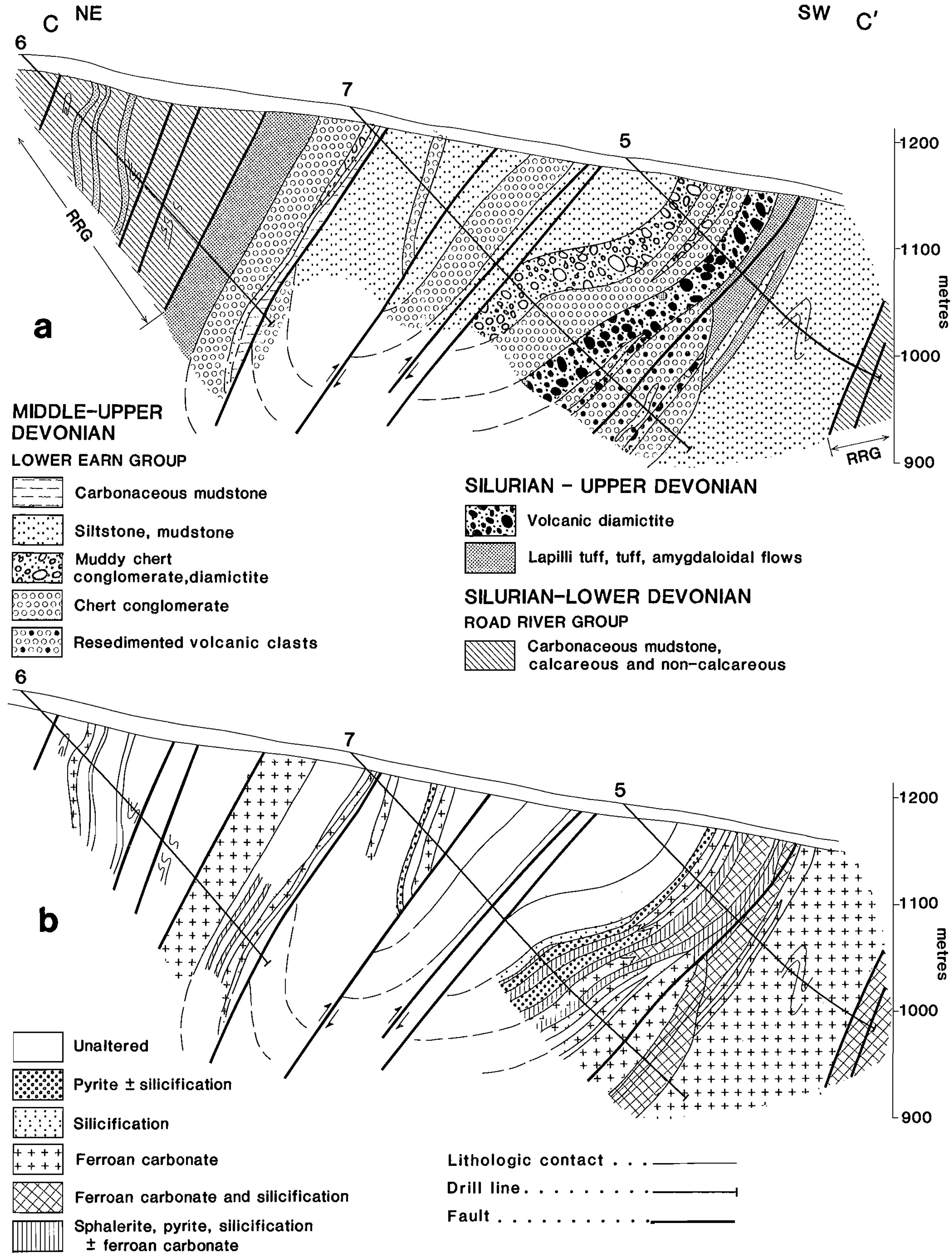

Figure 5. North-south structural cross-section C-C' through Boundary Creek deposit illustrating distribution of (a) lithologies and (b) alteration/mineralization types.

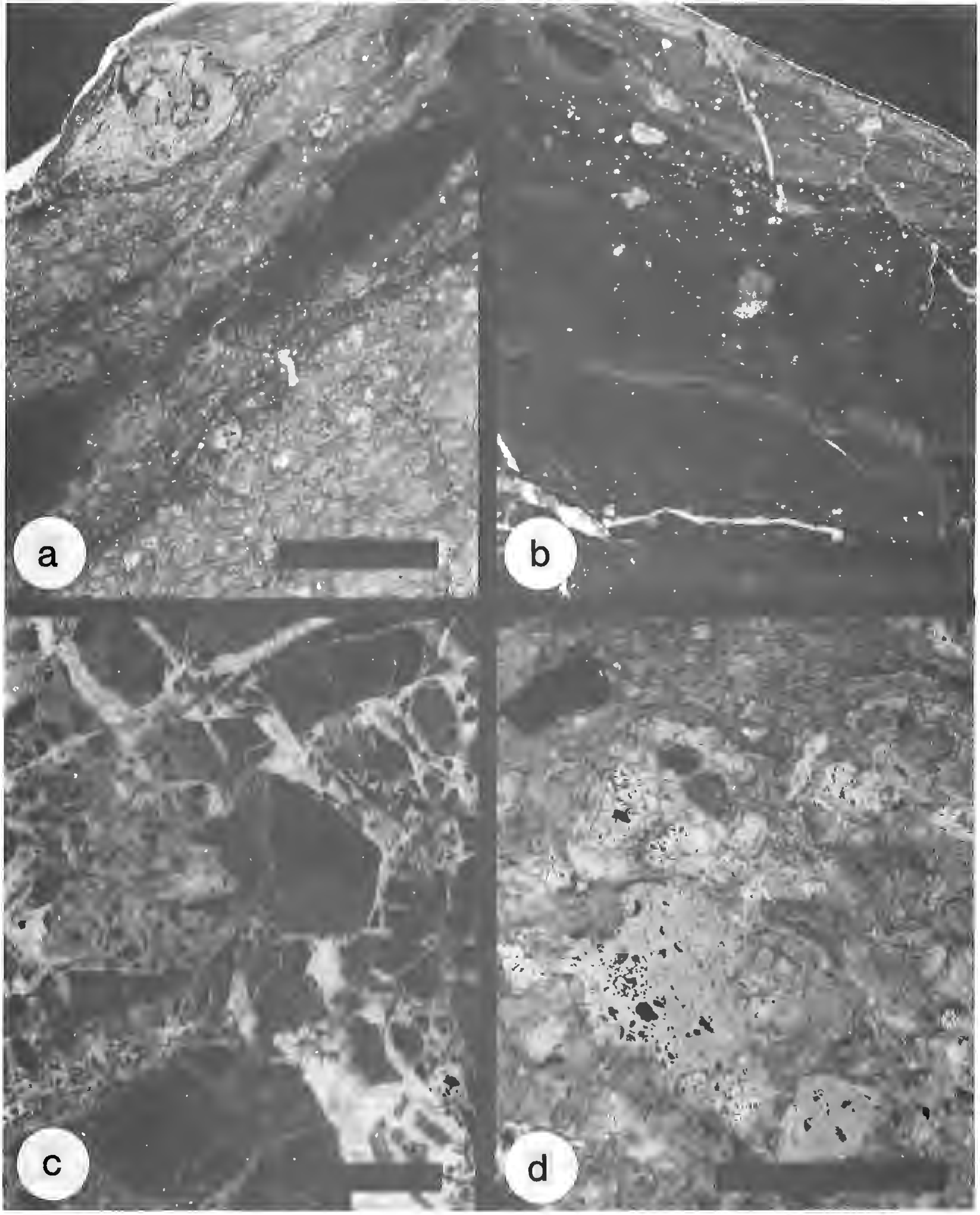

Figure 6. Photographs of representative rock types from the Boundary Creek deposit. Scale bars are one centimetre. **(a)** Cataclastic rock within syndepositional fault zone composed of fragments of ferroan carbonate (grey) ferroan carbonate breccia (b), and mudstone (black) (DH14-260.75m). **(b)** Cataclastic rock within syndepositional fault zone composed of polymictic fragments including ferroan carbonate, within fine-grained foliated matrix (DH14-260m). **(c)** Ferroan carbonate breccia, syndepositional fault zone (DH13-13.7m). Silicified mudstone fragments (black) in siderite matrix (grey) and partially replaced by buckshot siderite, cut by late ankerite breccia (light grey). **(d)** Lapilli tuff, Lower Earn Group (DH10-422.9). Amygdular volcanic and mudstone fragments in a matrix of ankerite cement.

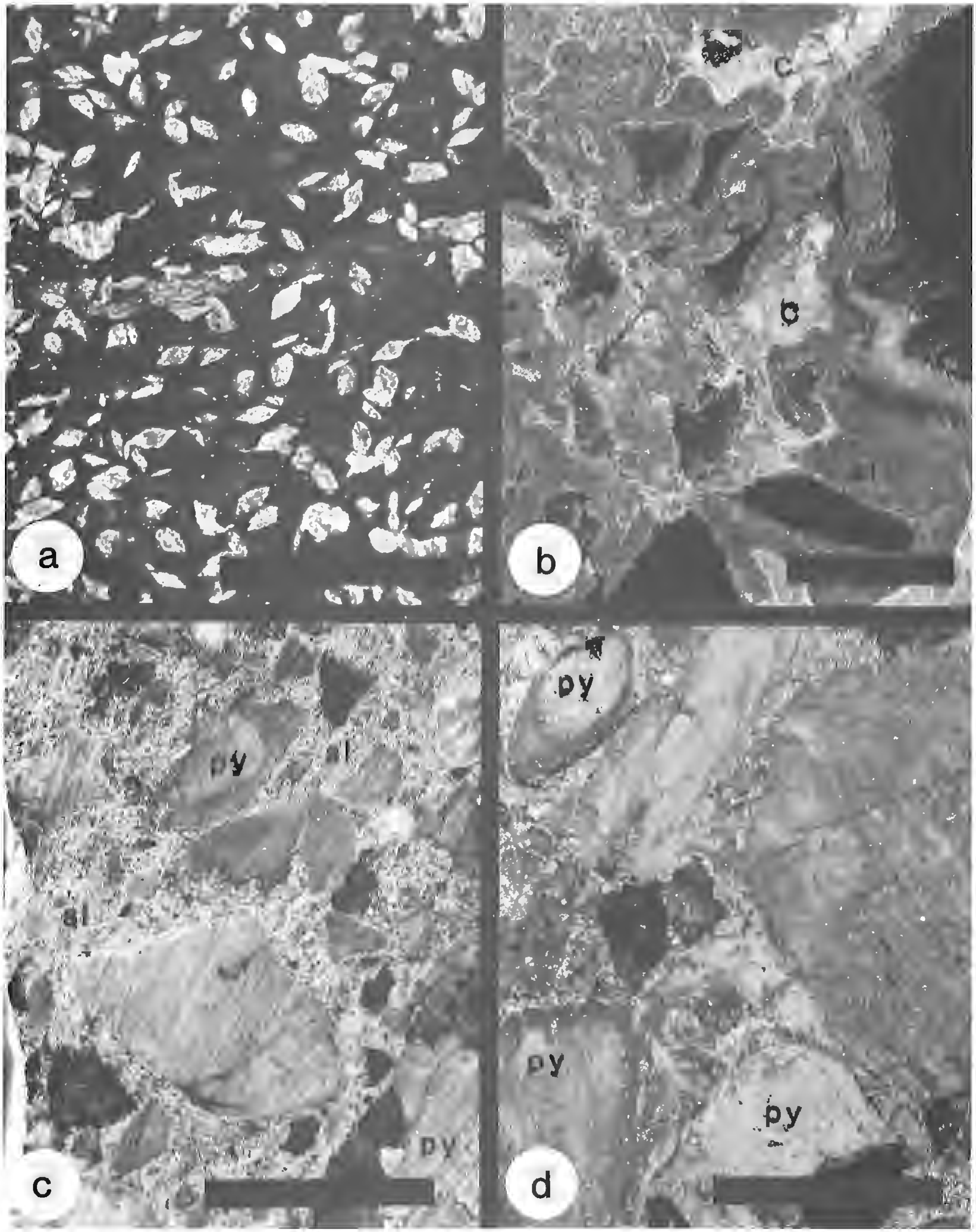

Figure 7. Photographs of representative rock types from the Boundary Creek deposit. Scale bars are one centimetre. **(a)** Disseminated euhedral crystals of siderite (light grey) or 'buckshot' siderite in silicified calcareous mudstone (black) (DH10-547.3m). **(b)** Vein breccia within silicified mudstone. Mudstone fragments (black) rimmed by early colliform sphalerite (sl) and younger ferroan carbonate (c) (DH10-369.1m). **(c)** Conglomerate with sphalerite cement (sl) and local replacement of clasts by pyrite (py) (DH10-278.7m). **(d)** Conglomerate with pyrite cement and abundant replacement of chert clasts by pyrite (py) (DH10-245m).

STRATIGRAPHY OF THE BOUNDARY CREEK AREA

To the east and west of the Boundary Creek area, Silurian-Devonian silty limestone is overlain by the middle turbidite member of the Lower Earn Group. Locally, lower member cherts of the Lower Earn Group (Abbott, 1986) are preserved below this unconformity. Ferroan carbonate altered lapilli tuffs, tuffs and flows occur interbedded with or at the top of the Silurian-Devonian silty limestone; 10 miles to the northeast of Boundary Creek such volcanic rocks disconformably overlie Ordovician to Silurian strata. On the north limb of the syncline, calcareous mudstone and sandstone with interbedded tuffs correlate with the Silurian-Devonian silty limestone. Silurian siltstone of Abbott (1983), and altered calcareous shales of the Silurian-Devonian silty limestone occur on the south limb of the syncline.

The Lower Earn Group in the Boundary Creek area can be divided into two major lithological assemblages: (1) a lower coarse-grained lithofacies (LCL) of conglomerate, muddy conglomerate, volcanic diamictite, volcanic tuffs, siltstone and mudstone; and (2) overlying carbonaceous siliceous mudstone equivalent to the upper siliceous shale member of Abbott (1986) and the Unit 3B of Carne (1979). Though the chert conglomerates of the LCL must correlate with those of the regionally widespread middle turbidite member, the latter lacks volcanic rocks, resedimented units such as diamictite and muddy conglomerate, and multiple rather than a single conglomerate horizon. Volcanic rocks are rare elsewhere within the Lower Earn Group. The LCL is limited in areal extent as typical middle turbidite member conglomerate is exposed immediately to the west and east of the Boundary Creek area. The anomalous nature of the LCL is interpreted to reflect the local environment adjacent to a Late Devonian syndepositional fault.

LITHOLOGICAL UNITS

Rapid facies changes, repetition of similar lithologies, structural complexity and poor surface exposure complicate our understanding of the stratigraphy of the Boundary Creek area. Emphasis here is given to description of lithological units.

Dolomitic siltstone

The oldest rocks drilled at the Boundary Creek prospect are dolomitic siltstones that we correlate with the Silurian orange weathering bioturbated mudstone unit of Abbott (1983). Massive cream to orange weathering, pale grey dolomitic siltstone to fine-grained sandstone contains 2-10 % distinctive dark argillaceous wispy laminae or bands and is interbedded with minor black mudstone beds. In DH-13, the contact with overlying black calcareous mudstone is a transitional zone of interbedded black calcareous mudstone and dolomitic siltstone (e.g. DH-13, 90-120m).

Black calcareous mudstone

Black carbonaceous calcareous mudstone forms the uppermost Road River Group in the Bounday Creek area. The mudstone is massive, faintly laminated or interbedded with varying amounts of calcareous siltstone laminae. A distinctive sub-type is mudstone interbedded with graded, bioclastic calcareous sandstone/siltstone beds up to 15 cm thick that give the rock a distinctive colour-banded appearance. Tuff and lapilli tuff beds to 5 m occur interbedded with parts of the mudstone unit. A carbonaceous, siliceous and non-calcareous mudstone associated with iron carbonate alteration and veining on the south side of the syncline is interpreted as altered calcareous mudstone.

Volcanic rocks

Ferroan carbonate altered lapilli tuffs, tuffs, and minor flows are interbedded with calcareous mudstone of the Road River Group, and mudstone, conglomerate and resedimented conglomerate of the lower Lower Earn Group over a stratigraphic interval of about 400 m. Intervals of interbedded lapilli tuff, tuff-breccia and tuff up to 40 m in thickness compose up to 15 % of the section. The highest tuff horizon occurs within pyritic mudstone at the top of the LCL and marks the end of local volcanism.

Thin-bedded grey massive to finely laminated tuffs interbedded with mudstone and tuffaceous mudstone beds are commonly pyritic. Lapilli tuffs include angular to subrounded clasts of pophyritic volcanic rock (to 80 %), mudstone (to 25 %) and chert (to 2 %) (Fig. 6d). Volcanic fragments are composed of up to 15 % feldspar and mafic phenocrysts to 2mm pseudomorphed by quartz, chlorite and clay, and ferroan carbonate amygdules in a fine-grained matrix of clay, ferroan carbonate and quartz or coarse ankerite cement. Tuff breccias are thicker bedded and contain similar fragments to 30 cm. Ferroan carbonate altered volcanic flows to 25m thick within the basal Lower Earn Group (DH-1-233-258m) are composed of a fine-grained groundmass with up to 20 % altered feldspar and mafic phenocrysts, and local amygdule-rich zones.

Siltstone and mudstone

Dark grey to black, siltstone and mudstone consisting of varying proportions of thin beds or laminae of siltstone, black mudstone and sandstone is interbedded with the conglomerates, diamictites and volcanic rocks. Thick silty mudstone occurs locally as the basal portion of the LCL (e.g. DH5-190-290m). Mudstone-rich rock with lesser siltstone laminae has a 'pinstriped' appearance. 'Banded' siltstones are predominantly centimeter-scale siltstone beds separated by thin mudstone laminae; siltstone beds may include a basal sandstone laminae. Massive, cross-laminated or graded sandstone beds up to 3 cm thick are up to 5 % of siltstone-rich sequences. Conformable light yellowish-brown bands of fine-grained ferroan carbonate up to 15 cm thick occur locally within siltstone-rich sections.

Chert conglomerate

Massive chert conglomerate up to several hundred metres in thickness in the Lower Earn Group occurs throughout most of the MacMillan Fold Belt. The Boundary Creek area is unusual as conglomerate is interbedded with resedimented muddy conglomerate, diamictite, volcanic rocks and siltstone. Conglomerate occurs as beds and amalgamated beds 5 to 100 m thick and comprise 40 to 60 % of the LCL. Massive, bedded or normally graded chert sandstone and conglomeratic sandstone beds are interbedded near the top of some conglomerate units.

The conglomerate is well sorted with rounded to subangular, white, grey and lessor black chert clasts (70 to 80 %) up to 3 cm across, and grey to black mudstone/siltstone clasts (5 to 20 %). Volcanic pebbles and boulders up to a metre in diameter are important in some conglomerate units (e.g. DH7-330-385m). While quartz commonly cements conglomerate throughout the MacMillan Fold Belt, at Boundary Creek quartz, ferroan carbonate, pyrite and sphalerite occur as monomineralic or polymineralic interstitial cements in the conglomerate.

Mud-rich conglomeratic rocks

Muddy conglomerate, chert pebble diamictite and mudstone diamictite represent a wide range of mud-rich chert conglomeratic rocks composed of chert pebbles, mudclasts, and mud matrix that occur as units up to 30 m thick, and compose up to 30 % of the LCL. Large mudstone/ siltstone clasts may exceed 2 m in diameter. Muddy conglomerates are differentiated from conglomerates by the presence of interstitial mudstone rather than a cement (e.g. quartz, carbonate) though both are grain-supported. Some conglomerate beds are transitional upwards into muddy conglomerate. Muddy conglomerate is transitional to matrix-supported chert pebble diamictite. The poorly sorted nature of mud-rich conglomeratic units, and the predominance of mud as a matrix for chert pebbles rather than clasts of mudstone and conglomerate suggests a local source of unlithified chert pebble gravels and interbedded muds. Similar chert pebble diamictites interbedded with the Jason stratiform zinc-lead deposit occur adjacent to a syndepositional fault (Turner, 1986).

Volcanic diamictite

Volcanic diamictite is a brown to grey coarse fragmental rock composed of subrounded to angular volcanic clasts up to boulder size, chert pebbles, and mudstone fragments within a grey to black fine-grained matrix . The dominant types of clasts within the volcanic diamictite are lapilli tuff pebbles and boulders altered to ferroan carbonate, quartz and clay, grey chert clasts up to 10 cm (10-30 %), black mudstone clasts up to 15 cm (10-40 %), and coarse-grained ferroan carbonate up to 2 cm. The fine-grained matrix varies in composition from ferroan carbonate-, chlorite- and pyrite-altered tuff to black mudstone. The bulk of the volcanic diamictite occurs on the south limb of the syncline at several stratigraphic levels, centred on the zone of altered and sulphide-rich strata. A major diamictite unit up to 45 m thick can be correlated for 1500 m along strike and is an important marker unit (e.g. DH5-123-153m) (Fig. 3,6). Diamictite consisting of volcanic boulders locally overlies lapilli tuff units from which it is partially derived (e.g. DH-5, 123-190m).

Carbonaceous mudstone

Black, carbonaceous, locally siliceous mudstone overlies the LCL and is correlated with the upper siliceous shale member of the Lower Earn Group (Abbott, 1986) and Unit 3B of Carne (1979). The mudstone varies from massive to faintly laminated and can have a gritty appearance. Pyrite, up to 5 %, occurs as disseminated grains up to 0.5 mm and less commonly as laminae, blebs and veinlets. In DH-10, the uppermost LCL is 40m of siltstone-mudstone with thin interbeds of diamictite, conglomerate and tuff in fault contact with overlying carbonaceous mudstone greater than 110m thick. On the west side of Boundary Creek area stratiform barite lenses of the Kobuk prospect occur within the upper carbonaceous mudstone near the contact with overlying sandstone of the Upper Earn Group.

ALTERATION AND MINERALIZATION

General character

The extent of altered and sulphide-rich rock at Boundary Creek is limited to the Lower Earn Group lower coarse lithofacies of conglomerate, diamictite, mudstone and volcanic rock and calcareous mudstone of the upper Road River Group that underlie these strata. In surface plan, the cross-strike trend of altered rock is to the northeast. This altered and mineralized zone span 400 m of stratigraphic thickness, and extends at least 1500 m along strike. East and west of the Boundary Creek area Lower Earn strata lack diamictites and volcanics, and multiple conglomerate horizons, and lack significant alteration or sulphide content. Highly pyritic mudstones at the top of the LCL (e.g. DH10-208.5-210.5m) are the youngest sulphide-rich strata.

Altered and mineralized rocks are composed of quartz, ferroan carbonate, clay, pyrite, sphalerite, and minor galena, chalcopyrite and sericite. These minerals occur as interstitial cements within conglomerates, as replacement bodies, or in veins. The extent of this altered and sulphide-bearing rock is referred to here as the Boundary Creek deposit.

Zonation

A textural zonation with increasing depth occurs in the central portion of the Boundary Creek deposit. Depth in the following discussion refers to stratigraphic position and therefore paleo-depth below the seafloor. Sulphides and carbonates commonly occur as interstitial cements at shallow levels; and as replacement of sedimentary and volcanic rock, and as veins at depth. The widest and most abundant veins are associated with the syndepositional fault. This vertical zonation suggests formation in a hydrothermal upflow zone along the syndepositional fault.

A zoning of sulphide minerals with depth is also present. Pyrite is the dominant sulphide in the shallowest part of the deposit. Sphalerite occurs with quartz as an intersitial cement in conglomerates or as replacement of volcanic rocks, or in veins interbanded with ferroan carbonate at intermediate and deeper levels. Highest zinc grades are associated with conglomerate and volcanic diamictite, and with sphalerite veins cutting silicified mudstone between conglomerate units or within the syndepositional fault zone. Galena is only significant in the deeper portions of the deposit, and chalcopyrite is limited to the syndepositional fault. Silicification is common throughout the intermediate and deep portion of the deposit; intense silicification is associated with bleaching of organic content and with darker sphalerite. Ferroan carbonate is an important cement within some conglomerates at intermediate levels and as disseminated euhedra (''buckshsot siderite''), fine-grained bands, and veins in the intermediate and deep portion of the deposit. Carbonate and sphalerite rarely occur together except within veins.

Although vertical position within the deposit is important, lithological type exerts the strongest control on the textural character of mineralization. Therefore, the alteration and sulphide occurrence is discussed under host lithologies.

Calcareous mudstone

Calcareous mudstones on the south limb of the syncline are commonly silicified; pervasive ferroan carbonate alteration is less widespread and occurs as 'buckshot' siderite (Fig. 7a). Vein stockworks are more locally developed and include sphalerite, pyrite, ferroan carbonate muscovite and galena. Vein minerals are often zoned.

Alteration zonation is recognized locally (DH10-540-570m) where unaltered calcareous mudstone have been progressively altered to silicified mudstone with buckshot siderite, and to intense silicification and carbonatization characterized by medium-grained dark grey ferroan carbonate (siderite?) replacement bands and disseminated ankerite. Sulphide veins do not appear directly related to pervasive alteration types; pyrite-sphalerite veinlets and breccias lack alteration selvages and occur within both quartz-buckshot siderite alteration, and intensely silicified and ferroan carbonate banded rock. Where timing relationships can be established (e.g. DH10-542.6m), early silicification is followed successively by dark grey ferroan carbonate, buff coarse-grained ankerite, and latest pyrite replacement.

Mudstone-siltstone

Pervasive alteration within Lower Earn Group mudstone and siltstone units is uncommon except adjacent to altered conglomerate, diamictite or volcanic units. However, pale yellowish grey bands of fine-grained ferroan carbonate 1 to 5 cm thick may compose up to 5 % of the rock. Though typically concordant, these bands locally cut bedding thus indicating a diagenetic rather than sedimentary origin. Thin mudstone beds between conglomerate units are commonly silicified with abundant veins (Fig. 7b). Crosscutting relationships suggest ferroan carbonate veins generally predate sphalerite veins (e.g. DH8-195-210m). Zoned veins commonly display central sphalerite and outer ferroan carbonate or pyrite (e.g. DH8-300m), but exceptions occur (Fig. 7b).

Conglomerate

Hydrothermal minerals occur both as cement and as replacements of clasts in conglomerates (Fig. 7c, 7d). Mineral zonation from distal to proximal sphalerite-bearing conglomerate are: (1) quartz cement; (2) pyrite cement and clast replacement with minor grey clay intersitial clay; (3) silicification of clasts; (4) pale fine-grained sphalerite; (5) darker medium-grained sphalerite with sericite; and (6) partial destruction of carbonaceous matter (bleaching) with intense silicification. This zonation can be mapped as mineral zones: (1) quartz; (2) pyrite; (3) pyrite + silicification; (4) pyrite + sphalerite + quartz; (5) pyrite + sphalerite + intense silicification + bleaching. It is unclear if ferroan carbonate cement fits into this zonation. Ferroan carbonate and sphalerite rarely occur together as cement; instead sphalerite-bearing veins cut ferroan carbonate cemented conglomerate.

Vertical mineralogical zonation occurs in some conglomerate beds (e.g. DH10-230-270m). The upper portions of conglomerate beds are characterized by pyrite cement while in basal portions sphalerite is the dominant cement, and clasts are silicified, bleached and partially replaced by pyrite. These trends suggest the basal gravels, likely because of better sorting, were more permeable to hydrothermal fluids and hence more highly altered and mineralized. At deeper levels (e.g. DH10-300-306m), some conglomerate beds are zoned from ferroan carbonate cement at the top to sphalerite-quartz-pyrite at the base.

Mud-rich conglomerate

Altered muddy conglomerate and chert diamictite are typically pyrite-rich and sphalerite- and ferroan carbonate-poor relative to conglomerates. In the shallower levels of the deposit, extensive pyrite replacement of muddy conglomerate units occurs as disseminations, discordant zones, or massive pyrite rock metres in extent. In deeper parts of the deposit, massive pyrite replacement is associated with sphalerite and centred on discordant veins of colliform pyrite, marcasite, and sphalerite (Fig. 8a).

Volcanic rock

Volcanic rocks throughout the MacMillan Fold Belt are highly altered to ferroan dolomite or calcite. At Boundary Creek, volcanic rocks are characterized by extreme iron enrichment (to 40 %) and are altered to siderite, ankerite, quartz, clay, pyrite and chlorite. Distinctive blue green clay (montmorillinite?) amygdules occur locally in sideritic and silicified volcanics (e.g. DH10-10-17m). Sphalerite is not common within volcanic rocks except near contacts with sphalerite-bearing conglomerate (e.g. DH10-295m), in veins (e.g. DH10-420-432m), or within the reworked (higher permeability?) top of tuffaceous units (e.g. DH10-513-517m).

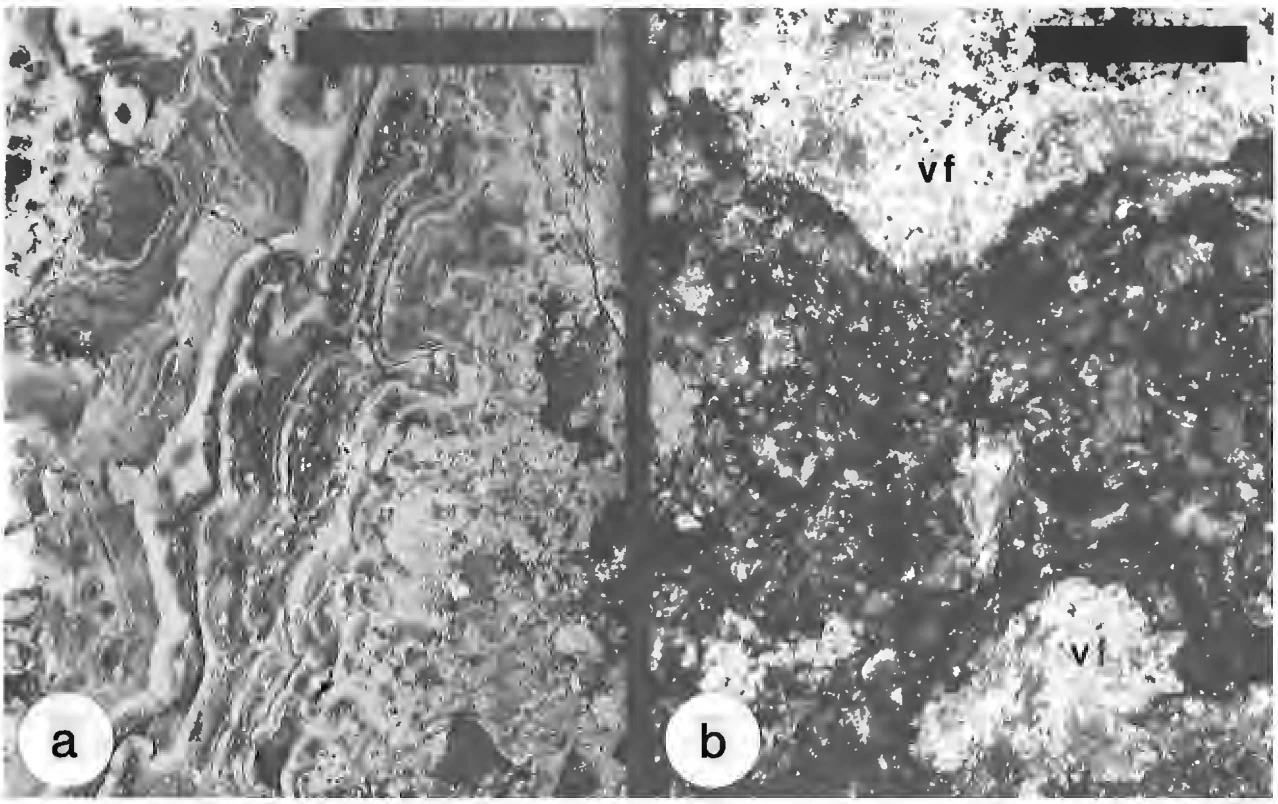

Figure 8. Photographs of representative rock types from the Boundary Creek deposit. Scale bars are one centimetre. **(a)** Contact between colliform pyrite vein (left) and muddy conglomerate wallrock largely replaced by pyrite (lower right)(DH10-467.35m). **(b)** Volcanic diamictite with volcanic fragments (vf) altered to quartz (grey) and pyrite-sphalerite (white), and silicified mudstone fragments (black) in silicified muddy matrix (DH10-507.3).

Volcanic diamictite

Both the volcanic clasts and tuffaceous matrix of volcanic diamictites are typically altered to ferroan carbonate with variable amounts of quartz, clay, chlorite, pyrite, sphalerite, and galena (Fig. 8b). Where the matrix is mudstone it is commonly silicified. Sphalerite and galena are commonly associated with grey silicification (e.g. DH8-310-312m); diamictites lacking sulphides are altered only to ferroan carbonate. In general, volcanic diamictites are significantly more altered and sulphide-rich than volcanic tuffs. In drillhole 10 (506-540m), ferroan carbonate altered tuffs are overlain by sphalerite-bearing volcanic diamictites altered to quartz, clay and pyrite.

Syndepositional fault

The syndepositional fault zone in DH14 consists of two cataclastic zones (261.5-262.5m; 265-268m) bounding and overlain structurally by veined and altered rock. The cataclastic matrix of fine-grained quartz, ferroan carbonate, and sulphides contains angular to rounded fragments of fine-grained ferroan carbonate, medium-grained ankerite veins, and siliceous mudstone (Fig. 6a,6b) and is cut by sugary sphalerite veins, and breccia veins of ankerite, chalcopyrite and pyrite. Veins lack alteration selvages. Local zones of ferroan carbonate breccia are largely replaced by pyrite. Between the cataclastic zones, silicifed mudstone is cut by a stockwork of medium to coarse-grained ankerite veins, large veins of coarse-grained ankerite, pyrite and chalcopyrite, and veins of pyrite breccia. Above the upper cataclastic zone, a massive ferroan carbonate-chlorite-quartz altered volcanic rock that is interbanded with carbonated lapilli tuff is cut by irregular patches of silicified rock, and up to 50 cm thick veins of banded sugary sphalerite and pink, buff and reddish ferroan carbonate. Minor late quartz or sericite pods occur central to banded veins. Offset wallrocks across some veins suggest formation along faults whereas banded veins reflect repeated openings of fractures and fluctuating physiochemical conditions of the fluid.

Drillhole 13 intersects only the lower portion of the fault zone (DH13-13.6-15.8), where textures in the ferroan carbonate breccia indicate a sequence of: (1) early silicification; (2) brecciation and deposition of siderite as cement, clast replacement and coarse-grained 'buckshot'; (3) brecciation and deposition of medium grey ferroan carbonate; (4) brecciation and deposition of orange weathering buff ankerite; and (5) planar quartz, and sphalerite-ankerite veins in subparallel sets.

SYNDEPOSITIONAL TIMING OF THE FAULT

The presence within the Lower Earn Group at Boundary Creek of (1) abundant resedimented sedimentary and volcanic material locally derived from uplifted Lower Earn sediments, (2) a fault zone associated with major base metal veins representing a hydrothermal fluid upflow zone, (3) structural elevation of the Road River Group along this fault, and (4) the presence of volcanic tuffs rare elsewhere in the Lower Earn Group collectively argue for the existence of a submarine syndepositional fault that was the locus of basaltic eruptions until Late Devonian time and a conduit for hydrothermal fluids during Late Devonian time. The resedimented material is comparable to the debris apron of resedimented mudstone, siltstone, mud and chert pebbles adjacent to the syndepositional fault at the nearby Jason stratiform zinc-lead deposit (Turner, 1986).

SHALLOW SUB-SEAFLOOR SULPHIDE DEPOSITION

It is possible that the Boundary Creek deposit represents (1) exhalative sulphide deposition coincident with deposition of the lower coarse lithofacies, (2) sulphide deposition at a shallow depth below the seabottom during the Late Devonian, or (3) sulphide deposition following Late Devonian time (e.g Cretaceous, Tertiary). The abundance of hydrothermal minerals as replacements and veins indicates a post-sedimentary formation. In addition, an exhalative model would predict the occurrence of laminated base metal sulphides within fine-grained mudrock interbeds within the lower coarse lithofacies; such laminated suphides are absent. The possibility that mineralization postdates the syndepositional fault is considered unlikely because: (1) ferroan carbonates and sulphides occur as fragments and disseminations within the cataclastite zones, interpreted to be the syndepositional fault; (2) banded sphalerite-carbonate veins are associated with the cataclastite zone; gouge zones reflecting younger faulting are not mineralized; and (3) the strong association of mineralization/alteration and fault scarp related debris. A shallow sub-seafloor origin is consistent with the strong vertical textural and mineralogical zonation within the deposit, and evidence for permeability control during mineralization. The abundance of hydrothermal cements in conglomerate suggest formation prior to loss of primary permeability, likely during early diagenesis.

The occurrence of sphalerite-pyrite in the highest conglomerate at the top of the LCL sets a minimum age on mineralization. It is possible that pyrite laminae in mudstones immediately overlying the LCL are syngenetic in origin and reflect venting of the hydrothermal fluids on the seafloor although a sub-seafloor origin for pyrite cannot be ruled out.

COMPARISON WITH NEARBY SEDEX DEPOSITS

The Boundary Creek deposit appears to have formed from a Late Devonian geothermal fluid similar to fluids that formed nearby Zn-Pb-barite stratiform deposits. It differs, however, from nearby stratiform deposits because sulphides were deposited below, rather than at the seafloor, though it is possible that a related exhalite base metal horizon may yet be discovered. The fact that hot, base metal-rich fluids do not appear to have vented significantly on the seafloor is probably due to rapid cooling resulting from subseafloor boiling or mixing with seawater in permeable gravels. The abundance of carbonate minerals may reflect boiling processes; such boiling may suggest a shallower water setting or a more CO_2-rich hydrothermal fluid at Boundary Creek relative to nearby stratiform deposits. The absence of barite in the deposit may argue against subseafloor mixing with seawater.

At both the Tom (Carne, 1979) and the Jason (Turner, 1986) properties, stratiform deposits occur near the base of the upper siliceous mudstone member (Unit 3B of Carne, 1979); if the pyritic mudstones overlying the lower coarse lithofacies represent the time of hydrothermal activity at Boundary Creek, then hydrothermal activity at Boundary Creek was approximately coeval with formation of the nearby stratiform deposits.

Fault scarp debris within the Lower Earn Group occurs at both Boundary Creek as well as the Jason and Tom stratiform deposits and indicate Late Devonian hydrothermal activity in the MacMillan Fold Belt was focused along syndepositional faults. The Boundary Creek deposit formed near the intersection of a northeasterly-trending syndepositional fault and the westerly trending Hess fault; the latter is also interpreted to have been active during the Late Devonian (Abbott, 1986). This structural setting is very analogous to that of the Jason deposit (Turner, 1986).

Widespread silicification, abundant ferroan carbonate and minor sericite, and the presence of pyrite, sphalerite and galena at Boundary Creek as well as the Jason and Tom deposits (Turner, 1986; Ansdell et al., 1989) suggest formation from similar hydrothermal fluids. A common paragenesis of early fine-grained siderite and later coarser-grained ankerite, as well as chalcopyrite, occurs in the syndepositional fault zones of both Boundary Creek and Jason. Boundary Creek, however, differs by a much higher ratio of sphalerite to galena, and the absence of pyrrhotite and minor pyrobitumen; the latter suggest a less reduced or more sulphur-rich fluid at Boundary Creek.

Unlike nearby stratiform deposits, Boundary Creek is associated with a volcanic centre, albeit a minor one. Although it is not clear if this association is coincidental or causal in nature, Boundary Creek provides a potential link between sediment-hosted stratiform deposits and volcanic processes. If basaltic magmatism was coeval with the formation of stratiform deposits in the MacMillan Pass area, it presents a mechanism for generating heated fluids to 250°C as noted in fluid inclusions at the Jason and Tom deposits (Gardner and Hutcheon, 1986; Ansdell et al., 1989). Though alteration makes the protolith difficult to interpret, volcanic rocks at Boundary Creek appear to represent a submarine basaltic centre dominated by pyroclastic eruptions. It is likely that these volcanic rocks are of alkaline affinity similar to other volcanics of Paleozoic age within Selwyn Basin (e.g. Mortensen and Godwin, 1982).

Drilling at Boundary Creek indicates large volumes of distinctly sub-economic zinc, and traces of lead and silver. The mineralized system at Boundary Creek contains a similar tonnage of Zn but subordinate Pb when compared to the exhalative Tom and Jason deposits.

Our understanding of hydrothermal fluids that form stratiform sediment-hosted zinc-lead deposits is limited by the lack of associated subseafloor vent complexes with vein minerals deposited from hydrothermal fluid prior to significant mixing with seawater. Temperature, salinity, major cation ratios, and sulphur and oxygen fugacity of the hydrothermal fluids have been estimated for only the Jason, Tom and Silvermines SEDEX deposits (e.g. Gardner and Hutcheon, 1985; Ansdell et al., 1989). Boundary Creek has the potential to make an important contribution in this regard because it formed at the same time as the nearby stratiform deposits and contains well preserved large sphalerite-carbonate feeder veins.

ACKNOWLEDGMENTS

The discovery of the Boundary Creek prospect and subsequent development of geological understanding is due to many Cominco geologists,whose contributions are acknowledged. We thank Cominco Ltd. for support and permission to publish this interim report on our ongoing exploration and study of the Boundary Creek project. We also thank Wayne Goodfellow for a helpful review of our manuscript.

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Structural controls on veins of the Mount Skukum gold deposit, southwestern Yukon[1]

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Love, D.A., Structural controls on veins of the Mount Skukum gold deposit, southwestern Yukon; in Current Research, Part E, Geological Survey of Canada, Paper 90-1E, p. 337-346, 1990.

Abstract

Eocene volcanic rocks of the Mount Skukum volcanic complex host the Mount Skukum epithermal gold deposit, which is approximately 65 km southwest of Whitehorse. The deposit consists of one major vein, the Cirque Zone, and two smaller vein systems, the Brandy and Lake zones. The strain ellipse associated with Brandy Vein formation was oriented with its axis of maximum shortening north-northeast, its axis of maximum extension east-southeast, and its intermediate axis near-vertical. The Cirque Vein formed in a strike-slip fault, and higher grades and thicknesses occurred where the fault refracted as it crossed a rhyolite dyke. The fractures and faults that host the veins are Riedel shears that developed in a zone of high strain localized between the Tintina and Shakwak faults in Eocene time. Gold mineralization is hosted in brittle tectonic structures, and may not be related to volcanic activity.

Résumé

Des roches volcaniques éocènes du complexe volcanique du mont Skukum renferment le gisement aurifère épithermique du mont Skukum qui se trouve à environ 65 km au sud-ouest de Whitehorse. Le gisement est constitué du filon important de la zone de Cirque et de deux petits réseaux filoniens : la zone de Brandy et la zone de Lake. L'ellipsoïde des déformations associé à la formation de Brandy Vein a été orienté comme suit : axe de raccourcissement maximum nord-nord-est, axe d'allongement maximum est-sud-est et axe intermédiaire presque vertical. La veine de Cirque s'est formée dans une faille à rejet horizontale, et des teneurs ainsi que des épaisseurs plus grandes se manifestent là où la faille a subi une déviation au moment où elle traversait un dyke de rhyolite. Les fractures et les failles où se trouvent les filons sont des zones de cisaillement de Riedel qui se sont formées dans une zone de fortes contraintes localisée entre la faille de Tintina et de la faille de Shakwak, au cours de l'Éocène. La minéralisation en or est logée dans des structures tectoniques fragiles et peut ne pas être associée à une activité volcanique.

[1] Contribution to Canada-Yukon Mineral Development Agreement 1985-1989. Project carried by Geological Survey of Canada, Mineral Resources Division.

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PURPOSE

This is a preliminary report of structural relationships of gold-bearing veins at the Mount Skukum deposit, together with a reinterpretation of available geological data. This study is based on approximately three and a half months of mapping on surface, one month mapping underground, and two months logging drill core.

LOCATION AND MINING HISTORY

The Mount Skukum mine lies within the Mount Skukum volcanic complex, in the Wheaton River district, approximately 65 km southwest of Whitehorse (Fig. 1). Exploration in the Wheaton River district began with the discovery of silver and antimony veins in 1893. AGIP Canada discovered the Mount Skukum vein deposit in 1980 by reconnaissance stream sediment sampling. The Brandy veins were subsequently discovered by soil sampling, and the Lake veins by prospecting and rock chip sampling (McDonald et al., 1986). Erickson Gold Mines Ltd. began underground development of the Cirque Zone in 1984. Proven reserves at that time were 148 980 tonnes of ore with average gold and silver grades of 24.98 g/t and 20.5 g/t, respectively (McDonald et al., 1986). Production from the Cirque Zone began in February 1986 at an average rate of 300 tons (272 tonnes) per day. Development towards the more westerly Brandy zone began in 1986 and ore was produced from it for several months early in 1988. Development towards the Lake Zone began in 1987, but the ore shoots proved to be smaller than expected and were not mined. Milling, mining, underground exploration, and development ended in the summer of 1988 owing to exhaustion of the ore reserves. The mine produced approximately 80,000 oz of gold in its brief lifetime.

Previous work

M.J. Pride (née Smith) is engaged in a study of the overall geology of the Mount Skukum volcanic complex as part of a Ph.D. program (Smith, 1982, 1983; Pride 1984, 1986). B.W.R. McDonald, for his M.Sc., described the geology of the Mount Skukum deposit and studied the fluid inclusions and stable isotope geochemistry of the Cirque Zone (McDonald, 1987; McDonald et al., 1986, McDonald and Godwin, 1986). Doherty and Hart (1988) mapped all but the westernmost part of the Mount Skukum volcanic complex as part of 1:50 000 scale mapping contract under the Canada — Yukon Economic Development Agreement.

Tectonic setting

In the Paleocene epoch the absolute motion of the Kula-Pacific plate, which was actively subducting under North America, was north-northeastwards, and that of the North American plate was west-southwestwards. By Middle to Late Eocene time the absolute motion of the Kula-Pacific plate was northwestwards, whereas the absolute motion of the North American plate remained unchanged; thus their relative motion changed from northeastward convergence, and subduction, to northwestward transcurrent fault movement (Fig. 1, and Engebretson et al., 1985). The earlier

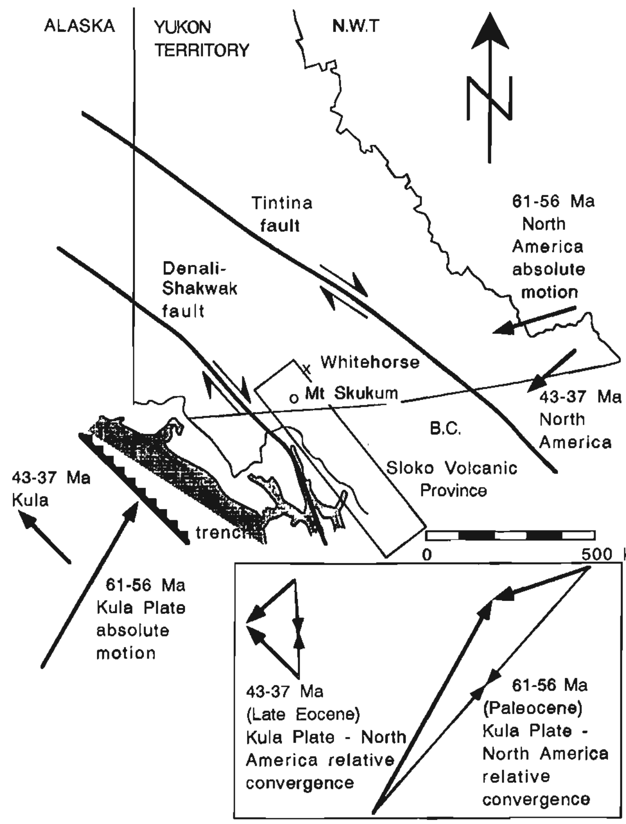

Figure 1. Location map of the Mount Skukum volcanic complex, showing Paleogene plate motions that affected the Yukon Territory (from Engebretson et al., 1985).

stress conditions produced the compressional structural trends characteristic of the Cordillera, whereas the later stress conditions resulted in right-lateral, strike-slip movement on the major northwest-trending faults such as the Tintina and Shakwak faults, and associated extensional strain.

General Geology

The Mount Skukum volcanic complex comprises intermediate and felsic volcanic rocks and derived sediments lying unconformably on Cretaceous granitic rocks of the Coast Plutonic Complex, folded Jurassic sedimentary rocks, and Precambrian (?) Yukon Group metasedimentary rocks. The complex is preserved as one or more down-faulted blocks, which overlap the boundary between the Coast Plutonic Complex and the Stikine Terrane. The volcanic complex is elliptical in plan, and covers an area of about 140 km^2. Part of an andesitic stratovolcano forms the western and southern parts of the complex. The eastern part is dominated by felsic volcaniclastic rocks, in which a cauldron-subsidence structure has been postulated (Pride, 1985, 1986). A quartz-feldspar — phyric rhyolite stock intrudes the centre of the complex.

The gold mineralization at the Mount Skukum deposit is hosted by quartz-carbonate-sericite veins that underlie Main Cirque, at the head of Butte Creek, in the central-western part of the complex (Fig. 2). Mineralized veins cut most of the volcanic and hypabyssal rocks exposed in the deposit area, namely, andesite flows, and massive to brecciated rhyolite dykes, but some narrow quartz-porphyritic intermediate dykes cut the veins. Multiple fracturing and filling is evident in all veins. Cockade and drusy textures, crustiform layering, colloform banding, and bladed calcite are common, all indicating open-space filling. Both adularia-sericite and quartz-alunite alteration are present in the Mount Skukum area, but the gold mineralization is associated only with the adularia-sericite — type alteration. The quartz-alunite — bearing assemblage is described in more detail in Love (1990). The regional geology of the area within a few kilometres radius of the mine as been mapped by Love (1989). The volcanic stratigraphy and geological history of the Main Cirque area interpreted from this mapping is described in Love (1990).

The vein systems in the Main Cirque area comprise three major, and many minor, north- to northeast-trending fault-hosted veins, with steep to vertical dips. The three major vein systems are the Cirque, Brandy and Lake zones. The Cirque Zone, the largest of these, strikes about 035° and dips 80° east. The ore zone within the Cirque Zone was about 200 m long, extended 80 m down dip, and averaged 5 m wide. The ore zone was contained in a fault that is at least 1.5 km long, in an area where the fault changes strike as it transects a north-trending rhyolite dyke. The Brandy and Lake zones are narrower, west-dipping vein systems, lying, respectively, 300 and 600 m west of the Cirque vein, and oregrade zones in them are approximately 100 m higher in elevation than the Cirque Zone.

Structural geology of the Cirque Zone

Structures in the Cirque Zone could not be mapped because mining there had been completed by the time this project was started. The following reconstruction of the geology of the Cirque Zone is based on core logging and company records.

The Cirque Zone occurs in a fault that can be traced for more than 1.5 km across most of the eastern side of Main Cirque. The fault strikes approximately 055° and dips very steeply to the east across most of the cirque. However, where this fault intersects a large north-trending rhyolite and rhyolite breccia dyke, its strike changes to about 035°, but its dip does not change (Fig. 2). the ore zone is restricted to that part of the fault that transects the dyke.

The abrupt change of strike of the fault is interpreted as the result of refraction because of the difference in mechani cal properties between the andesite and rhyolite. Continued movement on this faceted fault would have produced parallelogram-shaped or lens-shaped, prism-like openings (Fig. 3). The apparent sinistral horizontal offset of the rhyolite dyke, by this fault, and the refraction of the strike of the fault in the dyke, indicate that movement on the fault was horizontal sinistral and the maximum stress axes must have been oriented approximately north-south. If the movement on the fault was normal, refraction of the fault in the rhyolite dyke would have resulted in a change in dip but no significant change in strike. The result of continued movement on the fault was that the refracted part opened up into a tabular body (Fig. 3).

The "Wheaton Lineament", which is a prominent, northeast-trending, 40 km-long lineation detected by remote sensing (Doherty and Hart, 1988), passes approximately 6 km east of the Cirque Zone. Doherty and Hart (1988, p. 51) commented: "The 'Wheaton Lineament' probably represents a deep-seated structure which was

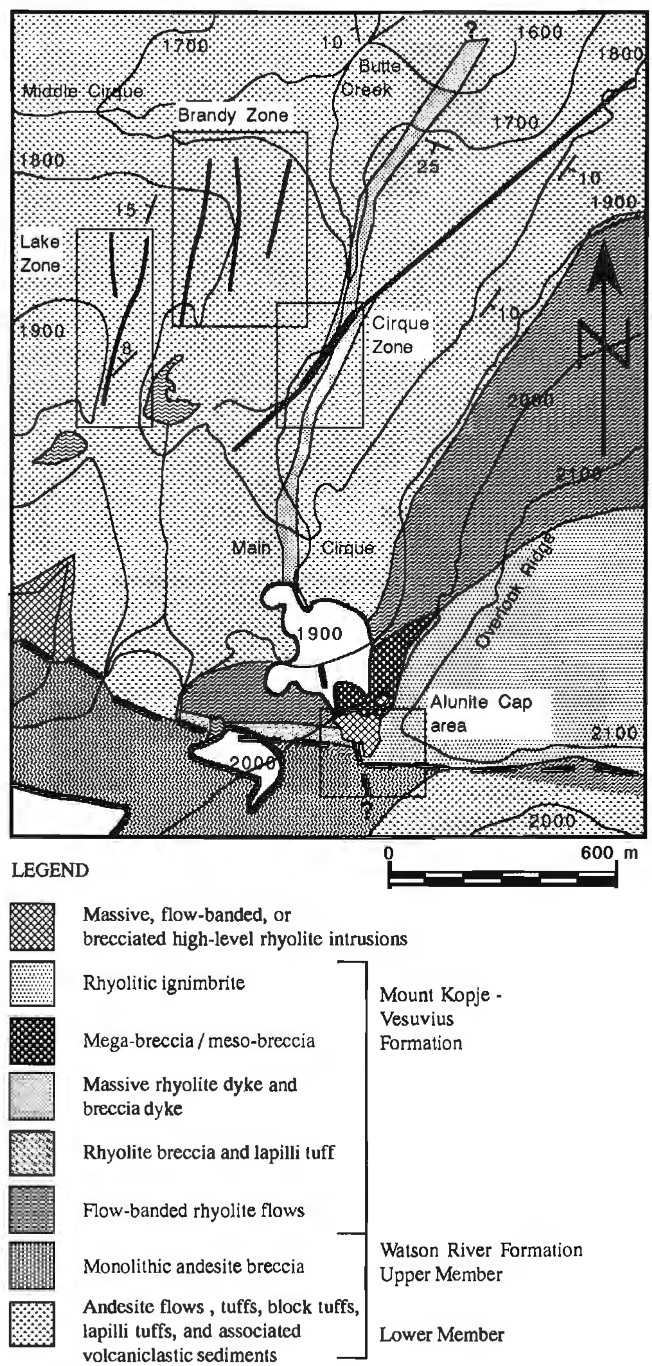

Figure 2. Simplified geological map of the Main Cirque area, showing the locations of the major vein systems.

active during pre-Tertiary time and reactivated during post-Eocene time." The fault that hosted the Cirque Zone has approximately the same orientation as this lineament, and may also have been a reactivated deep-seated structure.

The Cirque Zone occurs in the refracted part of the fault where the rhyolite dyke is in the footwall of the fault. A schematic diagram of the structural setting of the Cirque Zone (simplified from a reconstruction based on drill hole information; Love (1989) showed that dyke — fault intersection plunges to the south, but ore grades do not persist below about 1635 m elevation (Fig. 3C). This apparent horizontal cut-off of ore grades could represent either a stratigraphic control on vein formation, or the deepest level at which boiling took place in the vein. The other side of the fault, where the dyke is in the hanging wall of the fault, and also the down-plunge extension of the dyke — fault intersection, could both represent significant local exploration targets.

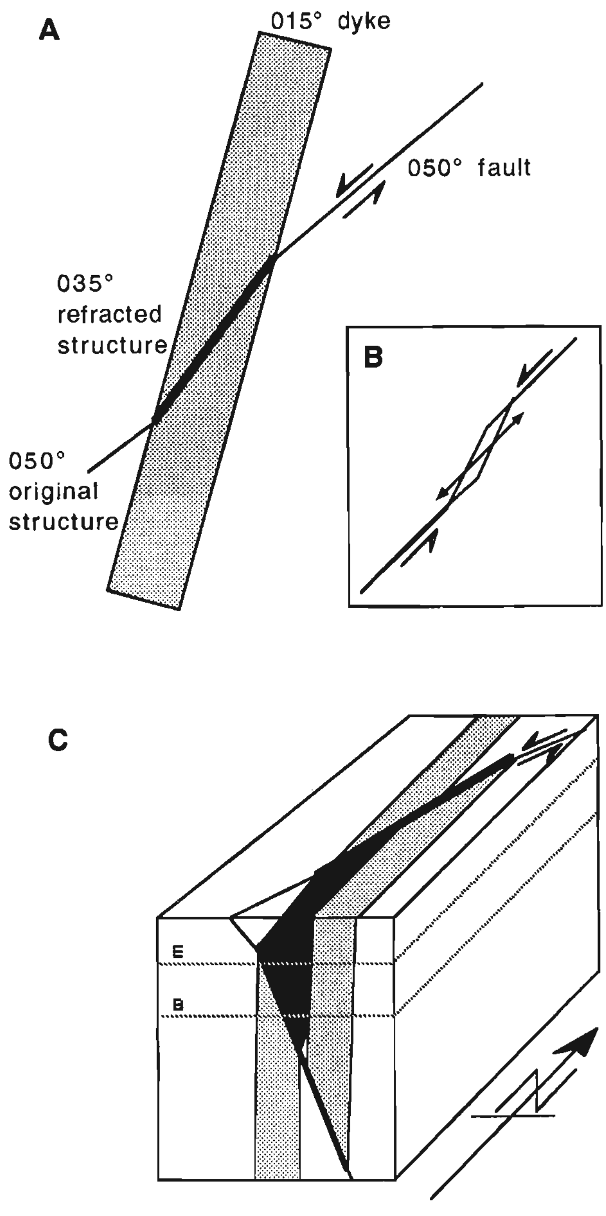

Figure 3. Structural setting of the Cirque Zone. A: refraction of the Main Cirque fault in a rhyolite dyke. B: parallelogram-shaped tabular opening developed as a result of movement on a non-planar fault. C: schematic block diagram of the Cirque Zone, B = base of the ore zone, E = present erosion level.

The Cirque Zone contains a wider vein than the Brandy and Lake zones, probably because of a combination of two factors: first, in the Cirque Zone a wide rhyolite dyke was present, and the consequent refraction of a fault resulted in the fault having an orientation suitable for continued movement to cause an opening to form; and second, fault movement was sufficient to open a large space.

Structural geology of the Brandy Zone

Presented in this section are the results of a detailed study of the structure of the Brandy veins. The structural analysis is based on structural measurements of vein-filled fractures and the orientation of slip lineations on the fracture walls, defined by underground mapping and drilling.

Five closely-spaced, parallel zones of veining were defined during exploration, development and mining in the Brandy Zone area, and were numbered Brandy veins 1 to 5, in order of discovery. BV2 was the only zone developed and mined, and thus the only one exposed for detailed mapping. Brandy Vein 2 was exposed underground for approximately 200 m of strike length, and over a vertical interval of about 50 m.

The objectives of this vein analysis are to: (1) define the geometry of the vein system; (2) determine the orientation and shape of the strain ellipsoid that describes the strain effected by movement on the vein-hosting fractures; and (3) determine the orientations of the stress axes for the stress field that caused the formation of the fractures and the pattern of movement in them.

Overall, BV2 strikes about 018°, and its dip varies between 70° and 55° to the west. Brandy Vein 2 is made up of a series of short, faceted veins, in some places joining in sharp, obtuse angles, or elsewhere disposed in an en echelon array. Some of the veins are curved. The wall rocks to the veins are not foliated, and a "jigsaw" texture of angular wall rock fragments that have not been transported or rotated far, is common in the veins, suggesting they are hosted in brittle fractures. Slickensides are common on vein walls, indicating these were shear, not tension fractures.

Analysis of fractures

The poles to BV2 vein-filled fractures (Fig. 4A) cluster around four point maxima at 100/16, 082/22, 120/13 and 300/04 (Fig. 4B). It is inferred that the veins in the Brandy Zone system are hosted in four main sets of fractures, at 190/74, 172/68, 210/77 and 030/86, in order of decreasing abundance (Fig. 4C).

From vein attitudes, the orientation and shape of the strain ellipsoid and the orientations of the stress axes can be estimated, assuming the veins formed by plane strain. Poles

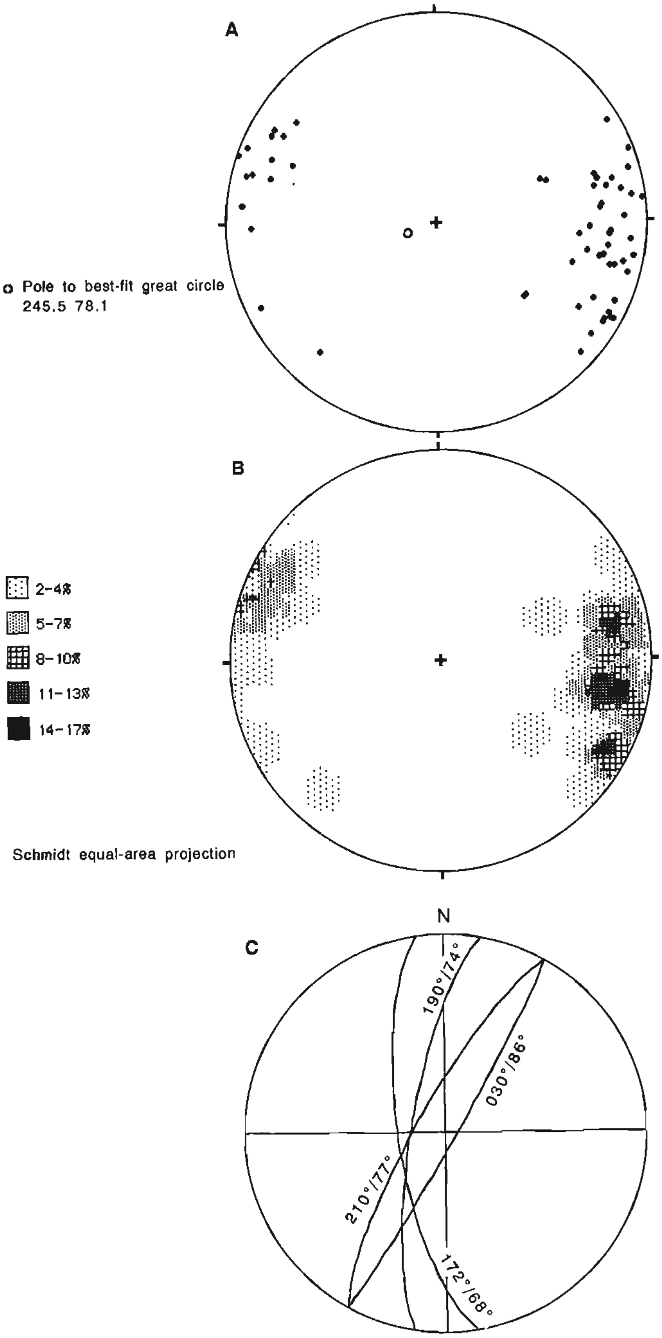

Figure 4. Stereonets of the Brandy Zone shear veins. A: poles to veins. B: contoured per 1 % area. C: an approximation of the four main vein sets in A and B. N = 60.

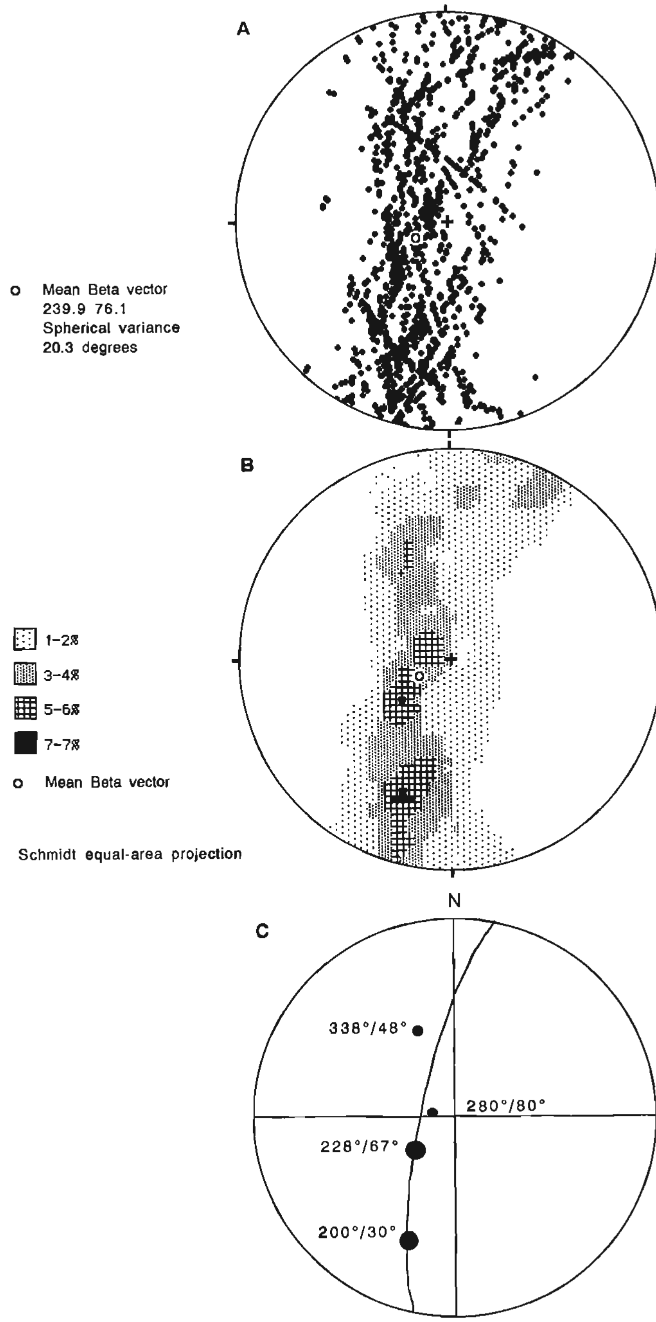

Figure 5. Stereonets of the lines of intersection of Brandy Zone shear veins, a Beta diagram. A: intersection lines. B: contoured per 1 % area. C: an approximation of the four major lines of intersection of veins in A and B. N = 1768.

to fault planes should form a girdle defining a great circle (π circle) that contains s_1 and s_3, the pole to which (π pole) is s_2. The poles to the Brandy Zone fractures do not clearly define a girdle, but the best-fit great circle through them is 336/12 and the pole to this, the orientation of s_2, estimated by this method, is 246/78 (Fig. 4A).

To check the interpretation above, the intersections of all the fracture planes were plotted in a "Beta diagram" (Fig. 5A) and the mean of the intersections calculated at 240/76. If the assumption of plane strain is correct, the intersections of all the fractures should form a point maximum (the β maximum), which corresponds to the orientation of s_2. However, rather than a single maximum, the fracture intersections are distributed in a girdle that has two large, equal maxima at 200/30 and 228/67, and two smaller maxima at 280/80 and 338/48 (Fig. 5B and C). These maxima of vein intersections are close to the intersections of the planes corresponding to the maxima on the π plot (Fig. 4B and C). The calculated mean Beta vector is between the two larger maxima. Geometrically, the β maximum should correspond to the π pole, but as noted by Phillips (1971, p. 68) in practice the "β axes are most satisfactorily located graphically from the π circle rather than directly from great circle intersections."

Analysis of lineations

All the lineations measured are slickensides on wall-rocks at the vein margins. Lineations (Fig. 6A) cluster around three point maxima (Fig. 6B), the largest one at 027/07, and the others at 002/18 and 341/23. The largest cluster of lineations coincides with the intersection of the planes corresponding to the two main maxima in the p diagram (Fig. 5C), the 210/77 and 030/86 ones. The other two clusters of lineations are on the planes corresponding to the other two maxima in Figure 5C. The best-fit great circle through the lineations is 222/26, and its pole is 132/64 (Fig. 6C).

For faults formed under conditions of plane strain, slip lineations on conjugate faults should lie in the s_1 — s_3 plane, where it intersects the faults. The s_1 axis should bisect the two maxima of lineations corresponding to the intersections of the s_1 — s_3 plane with the conjugate faults. For the Brandy Zone, the s_1 — s_3 plane should be the best-fit great circle through the lineations, 222/26, but there is a broad cluster of slip lineations covering about 60° of arc rather than two maxima. If the orientation of s_1 were estimated to fall in the middle of this cluster, it would be approximately 006/18. The s_2 axis should be the pole to the s_1 — s_3 plane, which for the Brandy Zone would be 132/64. However, these estimates of strain axis orientations from lineation analysis are not the same as those from vein orientation analysis.

DISCUSSION

The orientation of s_2 was determined by fracture analysis to be about 243/77, and by lineation analysis to be about 132/64. The s_1 — s_3 plane was determined by fracture analysis to be about 333/13, and by lineation analysis to be about 222/26. The inconsistency of estimates of the strain axes from fracture and lineation analyses indicates that the assumption of plane strain, upon which these analyses are based, is incorrect. Because there are more than two clusters of lineations and more than one cluster of fracture intersections, it is likely that the Brandy Zone fracture sets developed in a three-dimensional strain field. The orientation of the strain axes in the case of three-dimensional strain can be estimated using Krantz's (1988) odd axis model.

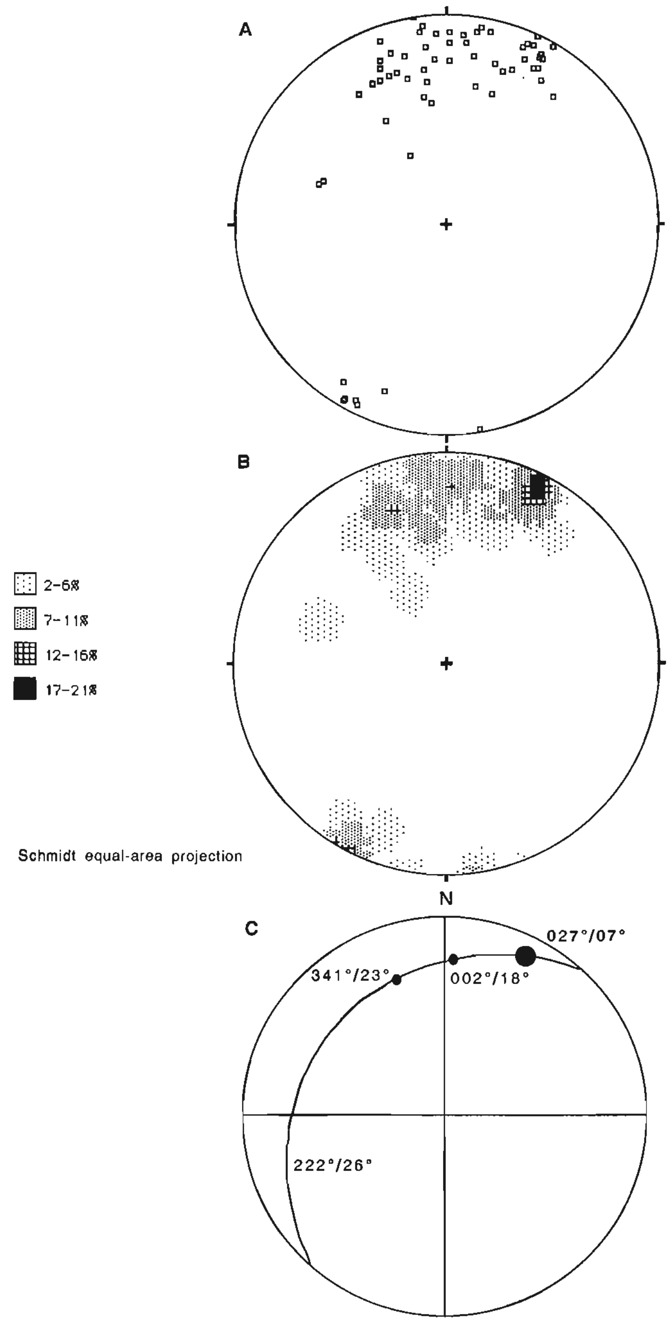

Figure 6. Stereonet of slip lineations on Brandy Zone vein walls. A: lineations. B: contoured per 1% area. C: an approximation of the three maxima of slip lineations, and a great circle through them. N = 61.

Three-dimensional Strain

In general three-dimensional strain there are three principal axes of extension: $\varepsilon_x > \varepsilon_y > \varepsilon_z$. One of these principal strain axes will have a sign opposite to the other two, and this one is termed the odd axis. The other two principal strain axes are termed the intermediate axis and the similar axis (which is either ε_x or ε_z and has the same sign as the intermediate axis). For irrotational plane strain ε_y is zero, and the slip vectors are defined by the intersections of the conjugate faults and the $\varepsilon_x - \varepsilon_z$ plane. However, for general three-dimensional strain ε_y has an absolute magnitude greater than zero.

Provided none of the principal axes of extension is zero, a minimum of four sets of faults in orthorhombic symmetry are necessary to effect three-dimensional strain. Also, in the case of three-dimensional strain, none of the slip vectors will lie in any of the principal strain planes. Rather, the slip vector for each of the fault sets is the intersection of the fault plane and a second plane containing the pole to the fault plane and the odd axis (Fig. 7). In other words, the slip vector, fault pole and odd axis are coplanar. Because the odd axis lies in the plane of the fault pole and the slip vector for all fault plane orientations, it is possible to locate the odd axis by a stereonet construction. The plane containing each fault pole and associated slip vector can be represented as a great circle (GC in Fig. 7), and the common intersection of these great circles or their average intersection is the odd axis. Also, the poles to the great circles should cluster around the intermediate strain axis.

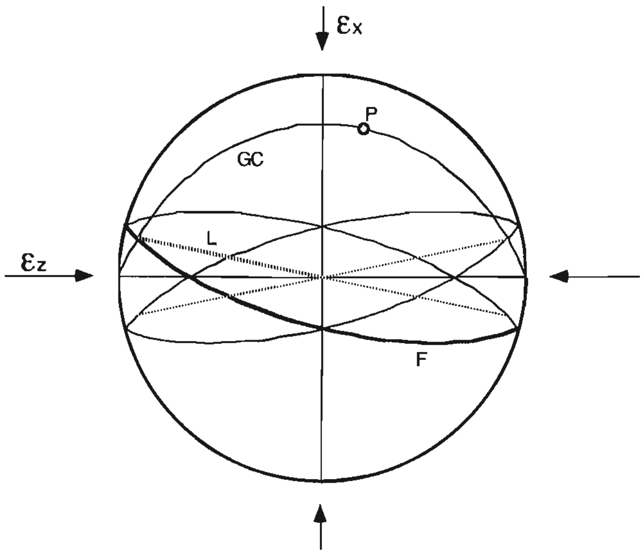

Figure 7. Stereonet of orthorhombic faults arranged in four sets (one heavy and three light solid arcs) symmetrical about the principal strain axes, and also showing slip lineations (one heavy and 3 light dotted lines). F: one of the faults/fractures. P: pole to F. ε_x: axis of maximum extension. ε_z: axis of minimum extension (maximum shortening). L: slip lineation of F. GC: great circle through ε_z, L, and P.

The intersections of all the great circles through the lineation and the pole to the fracture containing the lineation should ideally be a point maximum, but the Brandy Zone data results in a girdle (Fig. 8A) that has a maximum at approximately 015/13 (Fig. 8B): this maximum is interpreted as the odd axis. The other method to determine the odd axis is to plot the poles to the great circles, and to fit a plane through them; its pole should be the odd axis. Application of this second method to the Brandy Zone data yields a cluster rather than a girdle of poles (Fig. 9A and B). A best-fit plane through this cluster of poles (Fig. 9A) results in an odd axis orientation of 018/29. Average of these two results could be calculated, but because of the potential error involved in estimating a great circle through a cluster of points, the result of the first method, 015/13, is considered the best estimate of the orientation of the odd principal strain axis.

The other two principal strain axes must lie in a plane perpendicular to the odd axis: that is, in the plane 105/77. There are two methods of estimating the other principal

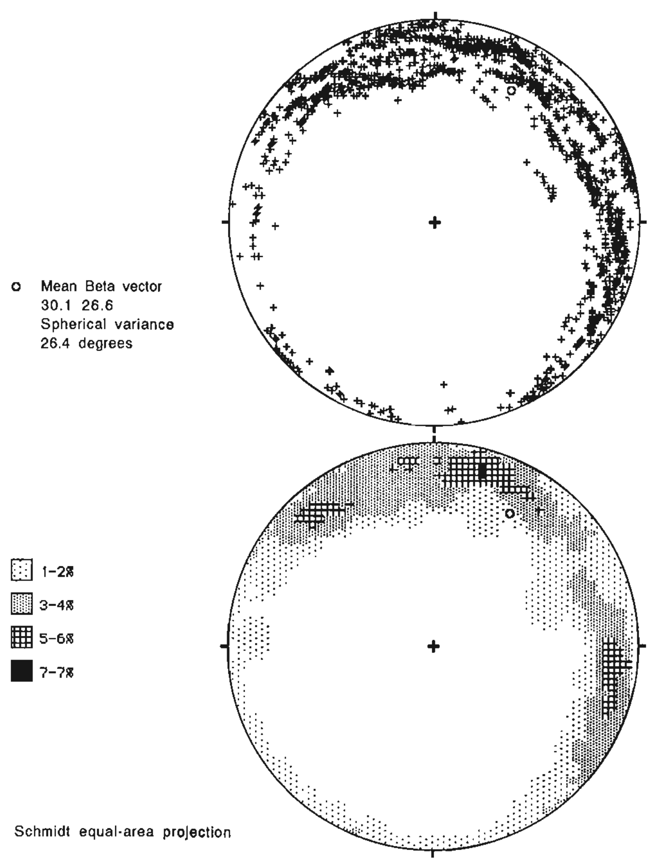

Figure 8. Stereonet construction for odd axis determination, showing the intersections of planes, where each plane contains the pole to a vein and the slip lineation on that vein's wall. The method is explained in the text and illustrated in Figure 7. Top: lines of intersection. Bottom: contoured per 1% area. N = 1538.

strain axes, each of which involves estimating one of them and solving geometrically for the other. The first method is to estimate two sets of great circles containing the lineation and the pole to the plane in which that lineation occurs, that intersect at the odd axis, and the similar axis should bisect the acute angle between them. The second method is to estimate the intermediate axis, which should bisect two equal maxima of poles to the same great circles, or coincide with one large maxima of these poles, and lie on the plane 105/77. Because the great circles do not form two obvious sets that intersect at the odd axis, whereas there is an obvious maximum of poles (Fig. 8, 9), the second method is used here. The largest cluster of poles to the great circles is oriented at 245/72, and lies on the 105/77 plane (Fig. 9). This is interpreted as the intermediate axis. The similar axis is perpendicular to both the odd axis and the intermediate axis, and therefore is oriented at 119/13.

Whether the odd axis is the ε_x or ε_z axis is determined from other structural geological information. Shortening in the odd axis direction would mean that ε_z is the odd axis. Because slip lineations cluster around the shortening direction, and in the Brandy Zone the lineations cluster near the odd axis, it is concluded that the odd axis for the Brandy Zone was ε_z. The orientations of the three principal strain axes are shown on the stereonet in Figure 10 and a schematic diagram of the fractures that host the veins, in the block diagram, Figure 11. The ellipsoid that represents the strain associated with vein formation was oriented with its axis of maximum shortening north-northeast, its axis of maximum extension east-southeast, and its intermediate axis, which was also an axis of extension, near-vertical.

The ε_y — ε_z plane approximates the overall trend of BV2, 018/55-70. Because the overall attitude of this vein system is approximately perpendicular to ε_x it appears to have formed with the orientation of a tension vein. Thus, Brandy Vein 2 may have formed as a series of four connected sets of shear veins that had the general orientation (but not the form) of a tension vein. Curved veins could have formed where strain was transferred from one fracture set to another during fracture development.

The fractures that host the Brandy veins are shear fractures formed by horizontal compression, and strike-slip movement as well as dilation accompanied vein emplacement. McDonald et al. (1986) suggested that the step-like topography of Main Cirque, which contains the ore deposit, was controlled by high-angle normal faults, and that the veins formed in the bounding faults between the down-dropped blocks, when they were reactivated as a result of resurgent doming. The structural analyses of both the Brandy and Cirque zones preclude this possibility.

Strain and Regional Tectonics

Strain axes oriented such as those described above could have been developed between two dextral strike-slip faults. An approximation of the shape and orientation of the strain ellipsoid determined from analysis of the Brandy veins, and its relationship to dextral strike-slip faults, are shown in Figure 12A.

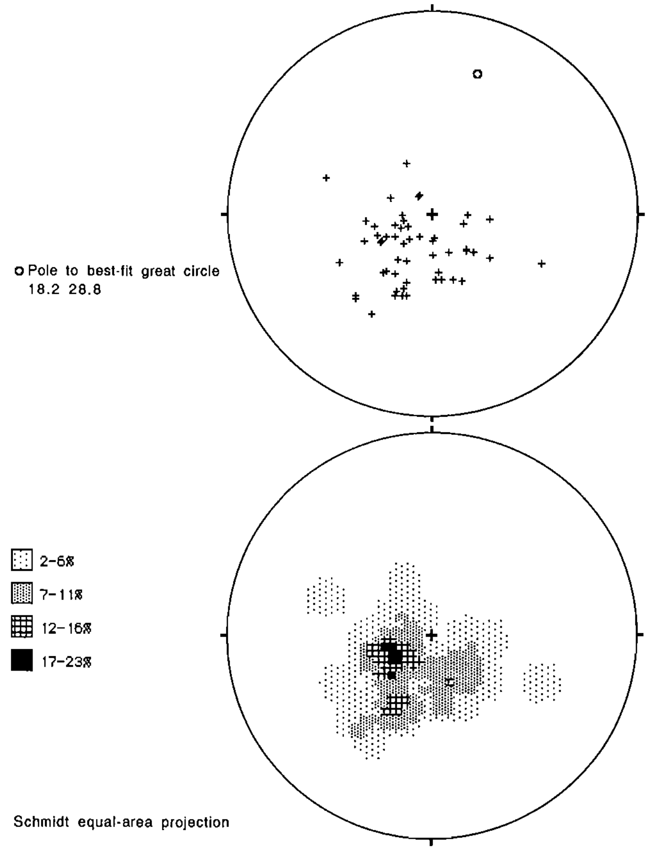

Figure 9. Stereonet construction for odd axis determination, showing the poles to planes, where each plane contains the pole to a vein and the slip lineation on that vein's wall. The method is explained in the text and illustrated in Figure 7. Top: poles to planes. Bottom: contoured per 1% area. N = 56.

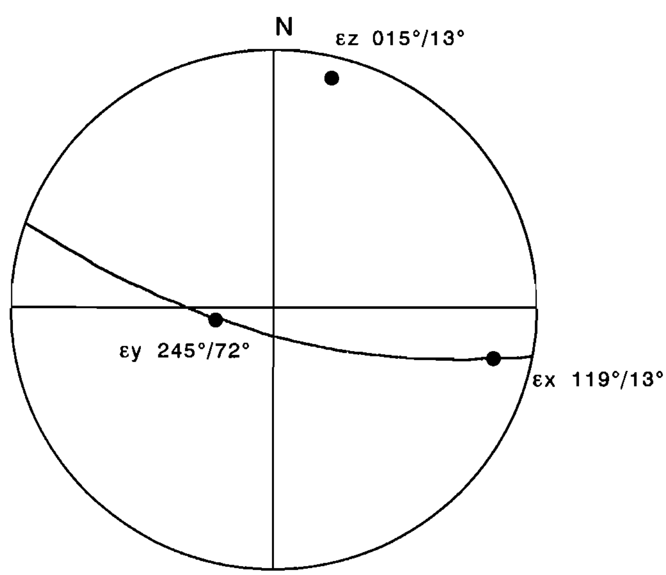

Figure 10. Stereonet showing the estimated principal strain axes during vein formation, based on analysis of veins and lineations in the Brandy Zone. The axes of extension are $\varepsilon_x > \varepsilon_y > \varepsilon_z$, where ε_z is negative, and, ε_x and ε_y are positive.

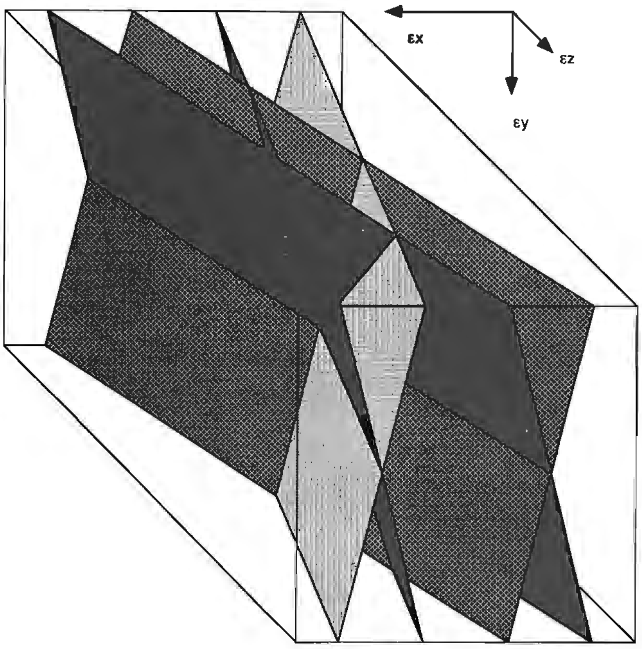

Figure 11. Schematic block diagram of four sets of faults/fractures in orthorhombic symmetry, in which the Brandy veins formed, and the principal strain (extension) axes that would have been responsible for their formation.

The preferred interpretation of the structural and tectonic conditions that led to formation of the Brandy Zone veins is that the veins are hosted in Riedel shear fractures that developed as a result of strain localized between the Tintina and Shakwak faults in Eocene time (Fig. 12B). Unfortunately, it was impossible to obtain any reliable sense-of-shear data on the various vein sets that would confirm this interpretation. The timing of dextral transcurrent faulting on the Tintina Fault has been constrained to Middle Cretaceous to Late Eocene or Oligocene times (Gabrielse, 1985), and on the Denali-Shakwak fault to the mid-Tertiary (Monger et al., 1982). The plate convergence rates and vectors between the Kula and North American plates changed between earliest Eocene and Late Eocene times (Fig. 1). This change in plate movement could have significantly affected the orientation of stresses in the zone lying between the Tintina and Shakwak faults. In earliest Eocene time the relative motion vectors were oriented northeast-southwest, virtually perpendicular to the two major transcurrent faults (Fig. 1), and there would have been little, if any, strike-slip movement on them. By Late Eocene time the relative motion vectors were reoriented to north-south (Fig. 1), and strike-slip movement on those faults would have been possible. The axis of maximum compression, in a zone weakened by fracturing between two faults, should be at 45° to the directon of movement of the faults (Casey, 1980). An orientation of s_1, at 45° to strike-slip movement vectors on the Tintina and Shakwak faults, would correspond to 010° to 015° and horizontal, virtually identical to the orientation of the major principal strain axis determined from the Brandy veins. Vein formation, therefore, may have been related to regional tectonics of the southwestern Yukon, and if so, must have occurred when strike-slip movement on the Tintina and Shakwak faults predominated. Vein formation may not be structurally related to the volcanic evolution of the Mount Skukum Complex.

Sinistral horizontal movement on the Main Cirque fault and the observed refraction are consistent with the maximum principal stress orientation of 015°/13°, as determined from the Brandy Zone. The Main Cirque fault can be classified as an R Riedel shear (Fig. 12B), assuming it was formed by the same stress conditions that controlled the orientation of the Brandy Zone. Intersections of dykes and other faults with the same orientation as the Main Cirque fault, and also conjugate, R Riedel shear faults, could be good mineral exploration targets.

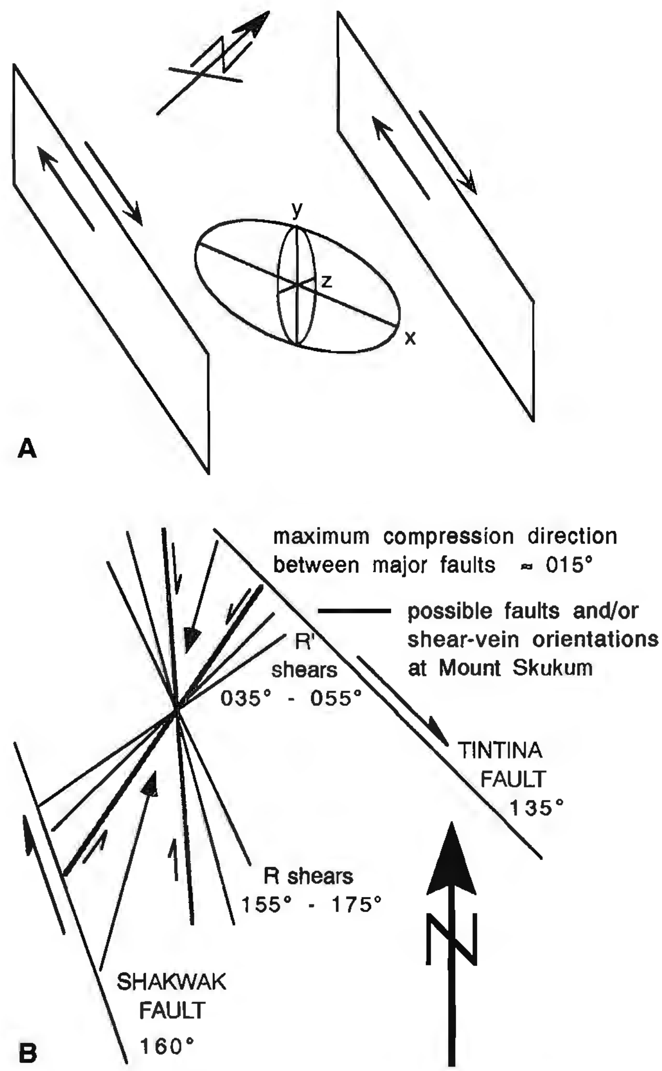

Figure 12. A possible model for the tectonic conditions responsible for the strain indicated by vein-filled faults and shear-fractures at Mount Skukum. A: strain ellipsoid orientation between two subparallel strike-slip faults. B: stresses and secondary shears that may have formed between the Shakwak and Tintina faults.

ACKNOWLEDGMENTS

This report presents some of the initial results of a Ph.D. study of the Mount Skukum gold deposit. I thank Mount Skukum Gold Mining Corp. for their support of the project. The study has benefited from discussions with their geologists, especially B. McDonald and R. Basnett. I also thank my supervisors at Queen's University, A.H. Clark and C.J. Hodgson, for their constructive and stimulating discussions and critical comments on the manuscript. This project is funded by the Canada — Yukon Mineral Development Agreement. And last but not least, I thank W.D. Goodfellow, of the G.S.C., for establishing and funding this project, and for his valuable discussions. This work was partly supported by a scholarship from the Natural Sciences and Engineering Research Council of Canada.

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Results of a biogeochemical orientation study on seaweed in the Strait of Georgia, British Columbia

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Dunn, C.E., Results of a biogeochemical orientation study on seaweed in the Strait of Georgia, British Columbia in Current Research, Part E, Geological Survey of Canada, Paper 90-1E, p. 347-350, 1990.

Abstract

*Samples representing the three main groups of seaweed - brown, green, and red - were collected from two sites along the shores of Texada Island, British Columbia. Instrumental neutron activation analysis of 43 elements in ashed tissue showed that As, U, and Sr are appreciably more enriched in the brown seaweeds than in the green or red types. Consideration of availability and practicality of collection of several species, and the value of the chemical information derived from them, indicate that the brown seaweeds (especially **Sargassum**) may be of use in identifying certain types of near shore mineralization, and in environmental monitoring.*

Résumé

Des échantillons représentant les trois groupes d'algues, soit les brunes, les vertes et les rouges, ont été prélevés à deux endroits le long de l'île Texada en Colombie-Britanique. Le dosage par activation neutronique de 43 éléments dans les tissus carbonisés a montré que les concentrations d'As, d'U et de Sr sont beaucoup plus élevées dans les algues brunes que dans les algues vertes ou rouges. L'examen de la disponibilité et de la praticabilité de la cueillette de plusieurs espèces ainsi que de la valeur de l'information chimique qui en découle, indique que les algues brunes (particulièrement Sargassum) peuvent servir à déterminer certains types de minéralisation en zone littorale et à contrôler le milieu.

INTRODUCTION

A few studies have investigated the chemistry of seaweed (macro-algae) to ascertain its use in mineral prospecting and environmental monitoring (Black and Mitchell, 1952; Bryan, 1969; Asmund et al., 1975; Bollinberg, 1975; Cooke, 1978; Sharp and Bolviken, 1979; Bollinberg and Cooke, 1985). Results have established, for example, that certain seaweed species are enriched in Pb and Zn near deposits of galena and sphalerite on the coast of western Greenland (Bollinberg and Cooke, 1985).

Since there is no record of this approach to biogeochemical prospecting along Canada's coasts, an orientation study was undertaken in the intertidal zone of Texada Island, British Columbia, in June, 1989. Common seaweeds were collected at two sites: one near skarn-hosted copper-gold mineralization; the second in a 'background' area of basalt.

SAMPLING AND SAMPLE PREPARATION PROCEDURES

Nine species of seaweed (Table 1) were collected at low tide by plucking the entire plant from the rocks. Holdfasts, the root-like attachment organs, were not collected. Species characteristic of each tidal zone were collected: two species from the upper tidal zone (green seaweed); five from the mid-tidal zone (brown seaweed); and two from the low tidal zone (red seaweed). Care was taken not to include rock fragments, and this proved time consuming for the green seaweeds collected here, since they encrust the rocks.

Seaweed contains a high water content, therefore where possible, samples weighing up to 1 kg were collected to ensure an adequacy of material for analysis. Samples were left to air dry for several weeks. The strong and unpleasant odour of the drying samples warrants the development of a procedure to squeeze the water out of the seaweed on site, to help accelerate the drying process and reduce the large amounts of sodium chloride that precipitate in the seaweed upon drying. For this study no such elimination of sea water was undertaken.

Table 1. Seaweed Species Obtained for this study

Botanical Name	Common Name	Description*	Type
Agarum fimbriatum	Laminaria (Kelp)	Dark brown ovate blades, with central rib; 20-80 cm long, 15-25 cm wide	Brown
Costaria costata	Laminaria (Kelp)	Dark brown ovate blades, with 5-7 longitudinal ribs; 10-30 cm wide, 50-250 cm long	Brown
Fucus gardneri	Wrack; rock weed	Olive-green, flattened branching structure with central rib to each branch; up to 50 cm high	Brown
Gigartina exasperata	Red algae (rough)	Reddish ovate blades with papillae (small protuberances); up to 45 cm high	Red
Nereocystis luetkeana	Bladder kelp; bull kelp	One of the largest brown algae. A long (up to 25 m) cylindrical stipe with a spherical float from which there are up to 20 blade-like fronds.	Brown
Sargassum muticum	Eel Grass; Gulf weed	Yellowish brown with profuse growth of filiform branches; up to 2 m high	· Brown
Schizymenia pacifica	Red algae (smooth)	Reddish, slimy ovate blades; up to 60 cm high	Red
Scytosiphon bullosus	--	Pale yellowish-brown, smooth and slimy, rounded, 1-10 cm high	Green
Ulva lactuca	Sea lettuce	Bright-green, ovate fronds 18-60 cm high. Ruffled like lettuce	Green

* Descriptions summarized from Scagel, R.F. (1967)

The dried samples were ashed at 470°C for 24 hours, after which some samples appeared still to contain charcoal even though all organic matter has usually volatilized under these conditions. Further ashing at this temperature failed to change this appearance indicating that it was not carbon; hence, in view of the chemistry of sea water and the high K content of the seaweed, it is probable that the dark crusty component was a complex chloride of Na, K, and Mg. By heating a portion of the material to 600°C it became greyish-white and hard, and fused to the sides of the porcelain crucible. Such annealing of the material at this temperature also suggests that the matrix is predominantly a salt.

The crusty samples obtained from heating at the lower temperature were powdered with a mortar and pestle, homogenized, and submitted for multi-element instrumental neutron activation analysis (INAA). Two counting procedures were performed; the first to obtain counts on short-lived isotopes of Al, Cl, Mn, Ti, V, Cu, Mg, and I; and the second for the longer-lived isotopes of 35 elements. Data for Al may reflect some contamination from the trays used for ashing, but patterns of variation for the different seaweeds give an indication of the relative concentrations of Al.

RESULTS

Data from the analysis of samples from the two sites are presented in Table 2. Site 1 is near the town of Vananda (Fig. 1), down-slope from the Cu-Au deposit of the abandoned Little Billie mine. Here Triassic marble is intruded by rocks ranging in composition from granodiorite to gabbro (Ettlinger and Ray, 1988). Site 2 is an area with no known mineralization, near Gillies Bay, on the opposite side of the island where outcrop is Triassic basalt.

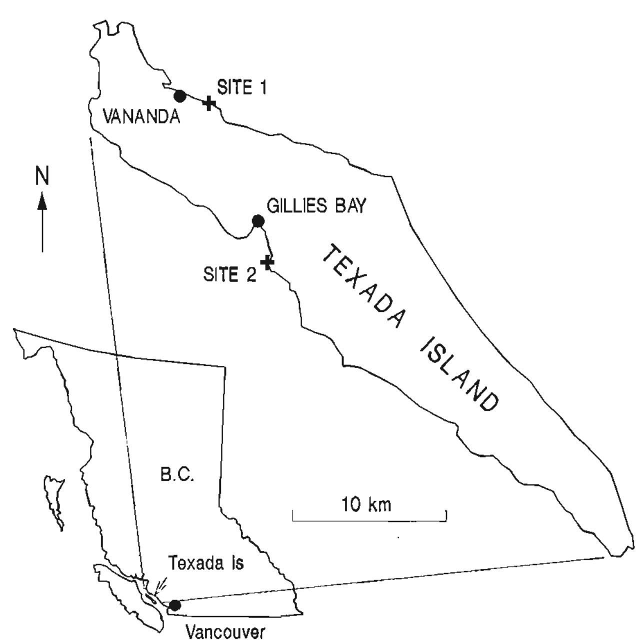

Figure 1. Location map showing sites of seaweed collection.

Table 2. Chemical Composition of Ashed Seaweed — Texada Island

		Sargassum (Eel Grass) Brown		Fucus (Wrack) Brown		Agarum (Laminaria) Brown		Costaria (Lamininaria) Brown	Nereocystis (Kelp) Brown	Ulva (Sea lettuce) Green		Scytosiphon Green	Schizymenia Red Algae (Smooth)	Gigartina Red Algae (rough)
		Marble*	Basalt**	Marble*	Basalt**	Marble*	Basalt**	Marble*	Marble*	Marble*	Basalt**	Marble*	Marble*	Marble*
Al+	ppm	43	8420	97	702	310	3630	223	132	2570	57400	3100	64	90
As	ppm	140	60	24	12	64	33	32	25	13	2.7	<2	12	12
Au	ppb	<5	<5	<5	<5	<5	<5	<5	<5	<5	<5	15	<5	11
Ba	ppm	<30	76	68	<30	72	81	<30	44	51	270	<30	<30	<30
Br	ppm	780	1200	1500	1200	3300	2000	1800	1700	670	340	1600	740	800
Ca	%	4.1	4.7	3.1	3.6	8.6	3.5	5.9	2.3	12.3	4.8	9.4	1.3	1.1
Cl	%	27.3	20.4	21.7	23.0	21.2	25.1	22.4	28.2	12.7	8.3	19.9	4.95	10.8
Co	ppm	2	3	3	2	4	3	3	2	1	6	2	2	2
Cr	ppm	<1	7	2	1	<1	<3	<1	2	<1	34	<2	7	<1
Fe	ppm	<500	3500	<500	<500	<500	1600	<500	750	900	11590	1100	<500	<500
Hf	ppm	<0.5	0.7	<0.5	<0.5	<0.5	<0.5	1.4	<0.5	<0.5	2.4	<0.5	<0.5	<0.5
I	ppm	39	21	<5	18	460	810	95	570	<5	<5	<5	<5	<5
K	%	28.9	15.7	10.6	15.4	24.6	24.3	27.7	28.7	9.0	6.4	13.7	15.1	17.2
Mg	%	0.14	1.2	1.1	0.8	0.67	0.5	0.67	0.45	1.6	N.D.	0.7	1.3	0.89
Mn	ppm	25	76	54	32	35	24	23	10	53	44	50	20	11
Mo	ppm	3	<2	N.D.	<2	<2	<2	<2	<2	15	<2	<2	<2	<2
Na	%	10.1	11.8	17.4	15.1	11.3	10.6	9.9	12.1	11.0	7.8	14.2	20.9	14.7
Ni	ppm	<50	<50	<50	<50	<80	<50	<50	<50	<50	89	<50	110	<50
Rb	ppm	73	44	45	41	53	68	71	80	21	16	23	47	54
Sb	ppm	0.2	0.5	0.2	<0.1	0.3	0.4	0.1	<0.1	0.3	0.6	0.5	<0.1	0.3
Sc	ppm	<0.1	1.1	<0.1	<0.2	<0.2	0.6	0.2	<0.1	0.2	6.4	0.3	0.1	<0.1
Sr	ppm	5700	6000	3200	2500	4300	2100	3600	3500	1500	580	2100	450	330
Th	ppm	<0.1	0.3	<0.1	<0.1	<0.1	0.2	<0.1	<0.1	<0.1	1	0.2	<0.1	<0.1
U	ppm	1.4	4.6	6	3	5.5	<0.3	2.2	1	0.8	1.6	2.1	<0.2	<0.2
V	ppm	<2	17	<2	<3	<3	16	<3	3	2	81	3	<2	<2
Zn	ppm	24	42	100	39	56	57	72	28	31	29	31	130	91
REE														
La	ppm	1.4	2.2	<0.5	<0.5	1.1	2	<0.5	0.5	<0.5	6.9	5.9	<0.5	0.7
Ce	ppm	<3	4	<3	<3	<3	<3	<3	<3	<3	12	6	<3	<3
Nd	ppm	<5	<5	<5	<5	<5	<5	<5	<5	<5	9	<5	<5	<5
Sm	ppm	0.2	0.3	0.1	0.2	<0.1	0.1	<0.1	<0.1	0.1	1.4	0.6	<0.1	<0.1
Eu	ppm	<0.02	0.15	<0.03	<0.02	<0.05	<0.03	<0.03	<0.02	<0.02	0.5	0.17	<0.03	<0.02
Yb	ppm	<0.07	0.32	<0.09	<0.07	<0.15	0.29	<0.06	<0.08	<0.06	1.17	0.28	<0.07	<0.06
Lu	ppm	<0.05	0.06	<0.05	<0.05	<0.05	<0.05	<0.05	<0.05	<0.05	0.17	<0.05	<0.05	<0.05
Ash	yield (%)	29.7	42.7	30.7	30.8	47.3	43.0	37.6	51.4	36.3	61.9	58.2	26.1	34.1

* Marble — near mineralized (Au/Cu) marble intruded by felsic pluton — Little Billie Mine, near Vanada
** Basalt — on amygdaloidal basalt with no known mineralization — Shelter Point, Gillies Bay
ND — not detected due to interference from other elements
Elements determined, but with concentrations all below detection limits (in ppm in parentheses): Ag(2), Cs(0.5), Cu(20), Hg(1), Ir(0.002), Ti(50), Se(2), Ta(0.5), W(1), Tb(0.5)
\+ Possible contamination of Al from trays used for ashing

This limited data set provides preliminary baseline information which can help in the design of future studies. Table 2 shows certain patterns of element enrichment and depletion, both with respect to species, and to substrate:

1. No species shows a clear tendency to accumulate Au or Cu. Only two species, the red algae (rough) and the green *Scytosiphon bullosus,* yielded Au concentrations above the detection limit of 5 ppb.

2. Arsenic is enriched in the brown seaweeds, notably *Sargassum* and *Agarum*. In all species for which comparative data are available, As concentrations are markedly higher near the mine site than at Gillies Bay.

3. Uranium levels are relatively high in the brown seaweeds, especially *Fucus* and *Agarum*.

4. Strontium levels are high in all the brown seaweeds, especially *Sargassum*. In three of the four species obtained at both sample sites, Sr is higher at the mine site probably relecting the relative enrichment of this element in the carbonate.

5. Iron, Al and REE are relatively enriched in the green seaweeds, with higher levels in the basaltic area than near the mine workings. Conversely, these seaweeds have lower concentrations of K, Rb, Br, and Sr.

6. The red algae (*Gigartina* and *Schizymenia*) have relatively low concentrations of Ca, Sr, and I.

7. Iodine exhibits a wide range in concentrations. This is either due to different abilities of the various seaweeds to accumulate I, or to different chemical speciation of I among the seaweeds resulting in the volatilization of I from some plant species during the ashing process, but not from others. The relative enrichments of I in *Agarum* and the kelp (*Nereocystis*) are striking. The green and red seaweeds all have I concentrations below the detection limit of 5 ppm.

8. In comparison with terrestrial vegetation the most obvious enrichments in the seaweed are Br, I, and Sr (the high levels of Cl, Na, and in part Br are attributable to crystallization of salts from the sea water during drying). Conversely, the seaweeds are relatively depleted in Ba, Mn, and Zn.

CONCLUSIONS

This orientation study shows no obvious concentrations of elements related to mineralization at the abandoned mine workings, except for As. The high levels in the brown seaweeds, especially *Sargassum* and *Agarum*, suggest that these species may be useful indicators of mineralization associated with As, and they may be useful media for environmental monitoring of As contamination. Similarly, relatively high U concentrations in *Fucus* and *Agarum*, and to a lesser degree in the other brown seaweeds, may give indications of uraniferous areas and assist in environmental monitoring.

Seaweeds obtain their nutrients directly from sea water, not through their root-like attachments; however, the influence of the different shoreline lithologies on the seawater chemistry appears to be reflected in the seaweed. The differences in element concentrations of the seaweeds obtained from the two substrates demonstrates that broader studies are required to characterize the affect of underlying rock type on the seaweed chemistry.

From the points of view of both practicality of sample collection, and chemical information derived, brown seaweed (especially *Sargassum*, in part because it is easy to identify) is the favoured medium. The red algae are the least informative, and the green seaweeds are the least practical to collect.

ACKNOWLEDGMENTS

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Reconnaissance observations on the Tim Williams Glacier rock avalanche, near Stewart, British Columbia

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Evans, S.G. and Clague, J.J., Reconnaissance observations on the Tim Williams Glacier rock avalanche, near Stewart, British Columbia; in Current Research, Part E, Geological Survey of Canada, Paper 90-1E, p. 351-354, 1990.

Abstract

In late 1955 or early 1956, a rock avalanche occurred near East Tim Williams Glacier in the northern Coast Mountains of British Columbia. The detached rock mass (approximately 3 × 10^6 m^3) consisted of highly fractured Jurassic tuff and argillite of the Hazelton Group. This rock mass disintegrated as it moved down a steep colluvial slope onto East Tim Williams Glacier. The debris then streamed at high velocity across and down the glacier; its leading edge came to a halt about 3.7 km from the source. The geological and geomorphic conditions associated with the Tim Williams Glacier rock avalanche are common throughout this part of the northern Coast Mountains, indicating the need for caution in the development of mineral resources in the region.

Résumé

À la fin de 1955 ou au début de 1956, il s'est produit une avalanche de pierre près du glacier Tim Williams est dans la partie nord de la chaîne côtière de la Colombie-Britanique. La masse rocheuse détachée (environ 3 × 10^6 m^3) est constituée de tufs et d'argilites jurassiques fortement fracturés du groupe de Hazelton. Elle se désintégrait à mesure qu'elle glissait le long d'une pente raide de nature colluviale, jusqu'à la surface du glacier Tim Williams est. Les débris ont alors traversé à grande vitesse le glacier avant de se diriger vers la bas de ce dernier; le front de cette avalanche s'est arrêté à environ 3,7 km de son point de départ. Les conditions géologiques et géomorphologiques associées à l'avalanche de pierre du glacier de Tim Williams se manifestent couramment dans tout ce secteur de la partie nord de la chaîne côtière; lors de la mise en valeur des ressources minérales de la région exigera donc la prise de certaines précautions.

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INTRODUCTION

A recent large major rock avalanche has been discovered in the Coast Mountains,70 km north of Stewart, British Columbia (Fig. 1). This landslide travelled across and down the East Tim Williams Glacier at the head of Tim Williams Creek, and thus is termed the Tim Williams Glacier rock avalanche. This rock avalanche is of particular interest in view of the intense mineral exploration in the region, centred on the nearby Sulphurets area. It is also one of 19 major rock avalanches known to have occurred in the Canadian Cordillera since 1855. Of these, 13 have occurred in the Coast Mountains, of which 8 have occured in glacial environments (cf. Evans and Clague, 1988). This report presents reconnaissance observations on the Tim Williams Glacier rock avalanche made during a visit to the site in August 1989.

TIMING

The rock avalanche is present on aerial photographs flown in July 1956 (Fig. 2), but does not appear on 1950 aerial photographs. The clearly defined debris margins seen on the 1956 photographs show little distortion attributable to glacier movement; because of this we conclude that the rock avalanche occurred in either late 1955 or early 1956.

DETACHMENT ZONE

A large rock mass (approximately $3 \times 10^6 m^3$) became detached between 2175 and 1850 m elevation on a high steep slope which forms the east valley wall above East Tim Williams Glacier. The rock on this slope consists of heavily fractured, massive to stratified tuffs and overlying thin-bedded argillite, assigned by Alldrick and Britton (1988) to the Jurassic Betty Creek Formation (part of Hazelton Group). One or more major, east-west trending faults cut these rocks in the vicinity of the detachment zone. The broken nature of the rock mass, evident on the 1950 aerial photographs, suggests that the slope had already undergone considerable deformation before detachment.

The dominante discontinuities in the detachment zone are persistent fractures or joints that dip steeply toward the valley (Fig. 3); these define much of the detachment surface.We were unable to determine the orientation of bedding in the highly fractured tuffs in the detachment zone, but the overlying argillites dip gently in a variety of directions.

POST-DETACHMENT BEHAVIOUR

The detached mass disintegrated as it moved down a steep colluvial slope below the detachment zone to the surface of East Tim Williams Glacier (elevation 1475-1525 m).A sigificant volume of colluvium was incorporated into the rock avalanche along this part of the path. The blocky debris swept across the glacier on a northwesterly bearing (approximately 330°) collided with the west valley wall. The leading edge of the debris ran up the slope to a point about 100 m above the glacier surface and was deflected on a northeasterly bearing (approximately 35°) back onto glacier. The

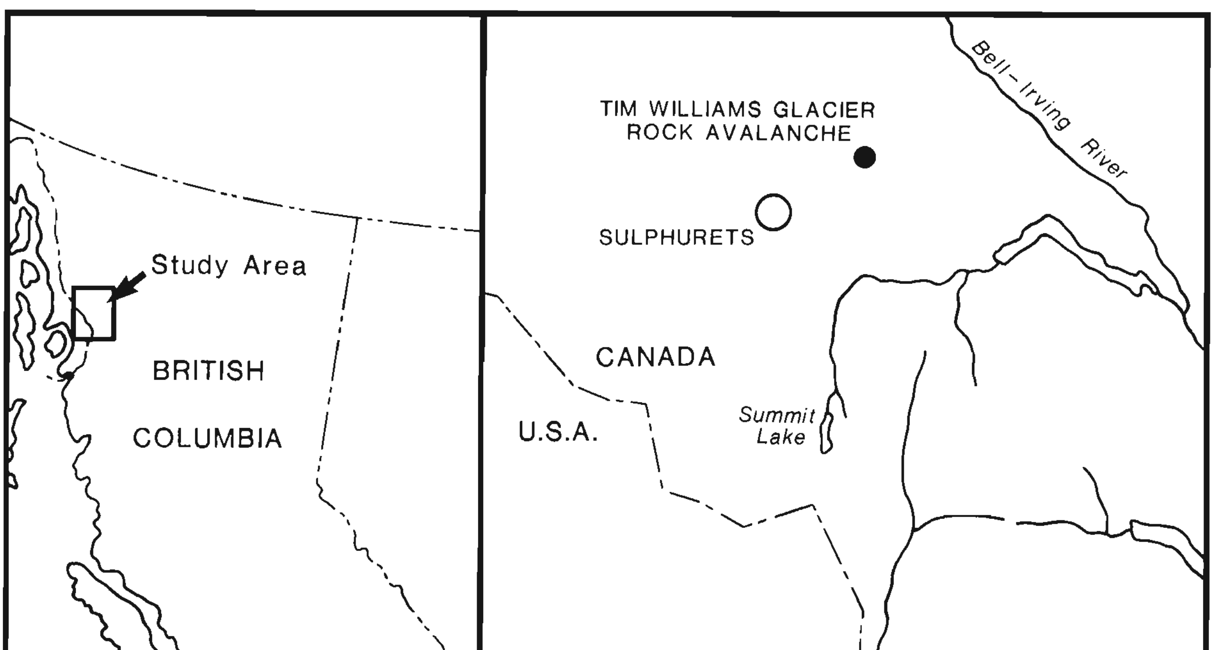

Figure 1. Map Showing location of Tim Williams Glacier rock avalanche.

debris front travelled a further 2.1 km and was partially deflected by a lateral moraine before coming to a halt. Horizontal and vertical distances from the top of the detachment zone to the northern extremity of the debris mass are 3.7 km and 935 m, respectively; this yields a travel angle (Fahrboschung; see Evans et al. 1987) of 14°.

CHARACTERISTICS OF THE DEBRIS

The margins of the rock avalanche debris deposit are sharp and relatively steep. The debris comprises angular fragments which are generaly less than 1 m in longest diameter, reflecting the closely spaced discontinuities of the parent rock mass (Fig. 4A). Some large blocks (Fig. 4B) exceed 10 m in diameter.

Well defined grooves or flowlines, which are parallel to direction of flow, can be seen in the rock avalanche debris in the 1956 aerial photographs (Fig. 2). Similar features

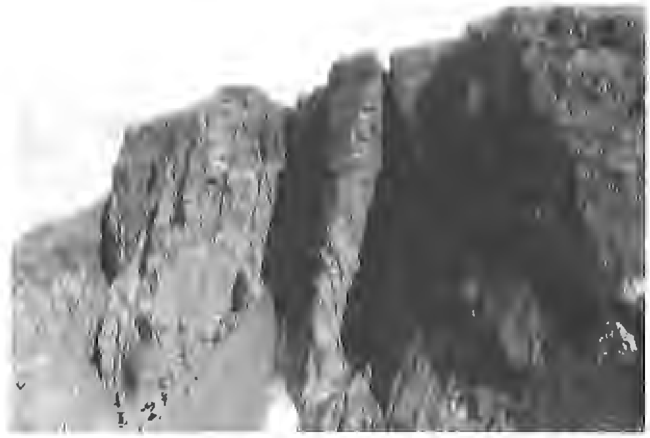

Figure 3. Aerial view of detachment zone showing steeply dipping structural discontinuities, intensely fractured tuff, and overlying thin-bedded argillite.

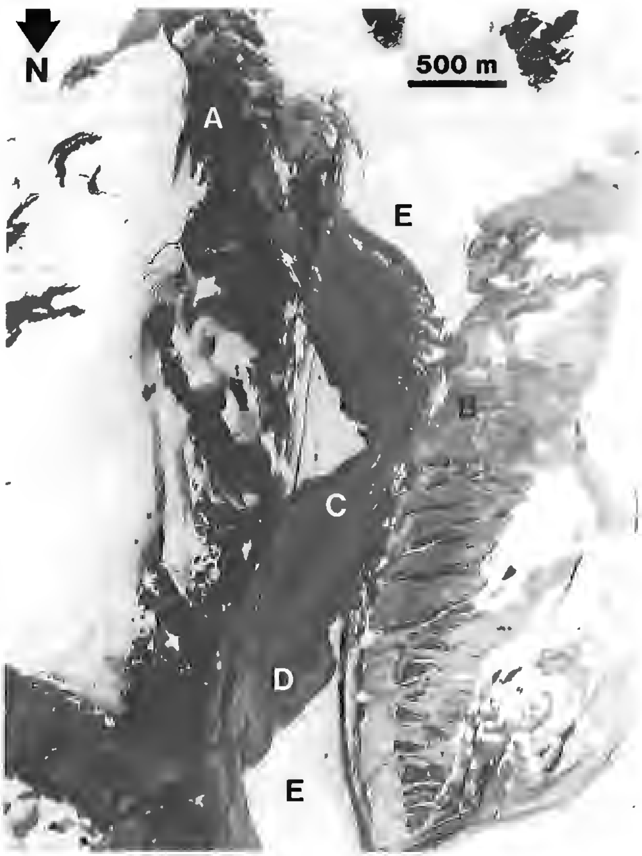

Figure 2. Vertical aerial photograph of Tim Williams Glacier rock avalanche taken in 1956. A = Detachment zone; B = run-up on west valley wall; C = flowlines in debris; D = Transverse banding resulting from compositional variation in the debris; E = East Tim Williams Glacier. BC 2182-52.

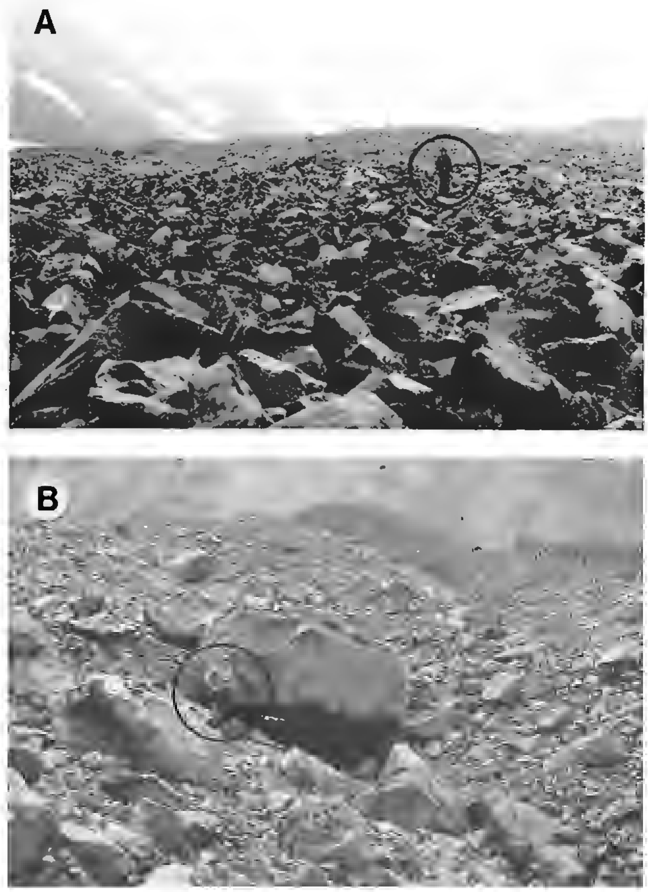

Figure 4A. Rock avalanche debris, viewed up glacier. Note figure (circled) for scale. Irregularities in the debris surface are caused by the melting of underlying ice and by glacier flow and crevasse formation. **B.** Large block in debris. Note figure for scale.

were noted in the debris of the classic Sherman Glacier landslide, triggered by the 1964 Alaska earthquake (McSaveney, 1978). Some of the flowlines cut across others, indicating a sequence of movements. There is also banding transverse to the direction of movement (Fig. 2). This reflects compositional variation within the debris; many of the darker bands are dominated by argillite, whereas light bands consist of tuff. Field observations in 1989 indicate that the debris sheet is relatively thin (for example, crevasses expose glacier ice at 1-3 m depth). The maximum thickness of the debris is probably 5-10 m.

The debris sheet has been transported 1 km northward by East Tim Williams Glacier since 1956. This movement, in combination with ablation of ice beneath the debris, has altered the surface morphology of the deposit (Fig. 4A,B).

IMPLICATIONS

The presence of deep open crack and downward-displaced blocks shows that the slope adjacent to the detachment zone of the Tim Williams Glacier rock avalanche is under considerable stress. One or more major rock avalanches may be expected to occur on this slope in the future.

The discovery of the Tim Williams Glacier rock avalanche indicates that the potential exists for large catastrophic landslides in this part of northwest British Columbia. The Hazelton Group may be particularly landslide-prone since it is dominated by weak rocks, such as pyroclastic and argillite. The presence of glacially oversteepened slopes and abundant meltwater further increases the possibility of catastrophic slope failures in this region. Finally, glacier downwasting and retreat during the last 100 years may have decreased the stability of some slopes, making them more susceptible to catastrophic failure in the future. These factors have significant implications for mining development in the region and highlight the need for caution in the siting of key facilities.

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Nonglacial sediments at Meadow Creek, British Columbia

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Fulton, R.J. and Warner, B.G., Nonglacial sediments at Meadow Creek, British Columbia, in Current Research, Part E, Geological Survey of Canada, Paper 90-1E, p. 355-358, 1990.

Abstract

Almost 10 m of subtill peat and wood-bearing nonglacial sediments occurs in the vicinity of Meadow Creek. Finite radiocarbon ages indicate that this material is Middle Wisconsinan in age. The deposits are dominantly medium to fine grained sand with interbedded gravel. This succession was deposited as overbank and channel sediments on an aggrading floodplain.

Résumé

On trouve, au voisinage du ruisseau Meadow, environ 10 m de tourbe s'étant accumélée sous du till et de sédiments non glaciaires renfermant du bois. Un nombre limité de datations au carbone radioactif indique que ce matériau remonte au Wisconsinien moyen. Les sédiments sont principalement constitués de sable à grain moyen à fin avec des couches de gravier interstratifiés. Cette succession a été mise en place sous forme de sédiments alluviaux d'inondation et de sédiments de chenaux sur une plaine d'inondation soumise aux effets de l'alluvionnement.

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INTRODUCTION

During construction of the Duncan Dam, nonglacial sediments deposited between 32 700 ± 800 BP (GSC-493) and 43 800 ± 800 BP (GSC-740) were exposed at a site near Meadow Creek, 8 km north of the north end of Kootenay Lake (50° 15′ 05″ N, 116° 59′ 00″ W; Fig. 1). The sediments exposed were described by Fulton (1968). Later, Alley et al. (1986) provided an interpretation of the paleoenvironment under which the upper part of the succession was deposited. During summer 1989, a backhoe was used to reopen exposures so paleoenvironmental studies could be started on the lower part of the succession (work to be done as part of a research agreement with B.G. Warner, University of Waterloo). This study is important because little information is available on paleoenvironmental conditions 40 000 years ago. What information is available seems to suggest that climatic conditions in south-central British Columbia at that time were similar to those at present, whereas conditions in the rest of Canada at that time apparently were cooler than they are at present. This preliminary paper presents the lithostratigraphic information obtained from this season's fieldwork.

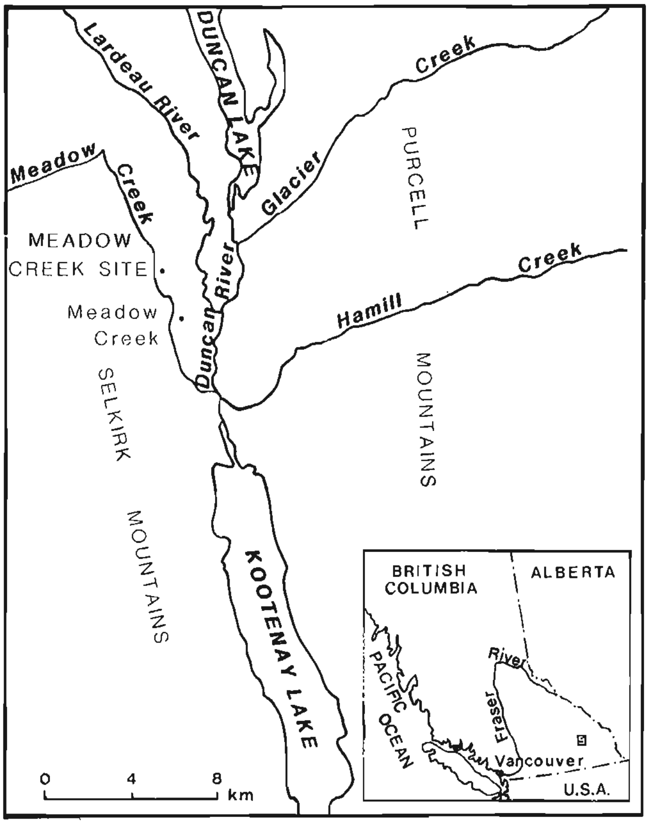

Figure 1. Location map.

LITHOSTRATIGRAPHIC FRAMEWORK

Five excavations were made to obtain samples for paleoecological and paleomagnetic analyses and to substantiate stratigraphic correlations made earlier. The lithostratigraphic information obtained is summarized in Figure 2. Correlation is based on lithology, stratigraphic succession, and unit thickness. Even though considerable local variation might be expected over short distances in a floodplain succession of this type, the main units appear to be continuous enough to permit reliable section to section correlation.

Bottom gravel

A pebbly gravel as thick as 130 cm occurs at the bottom of sites W and Y (Fig. 2). The gravel is well sorted and contains a large component of resistant quartzite and plutonic rocks and a relatively small component of local phyllite. The clasts in this unit are generally coarser than those in all except the highest gravels seen in these sections. Channel cut-and-fill structures are present. Study of this unit was difficult because it contained abundant water which quickly filled the bottom of the holes. The lower part of the gravel was very dense. This raises the possibility that the bottom gravel rests on a gravelly till, possibly the same unit as that described as exposed at the bottom of site V.

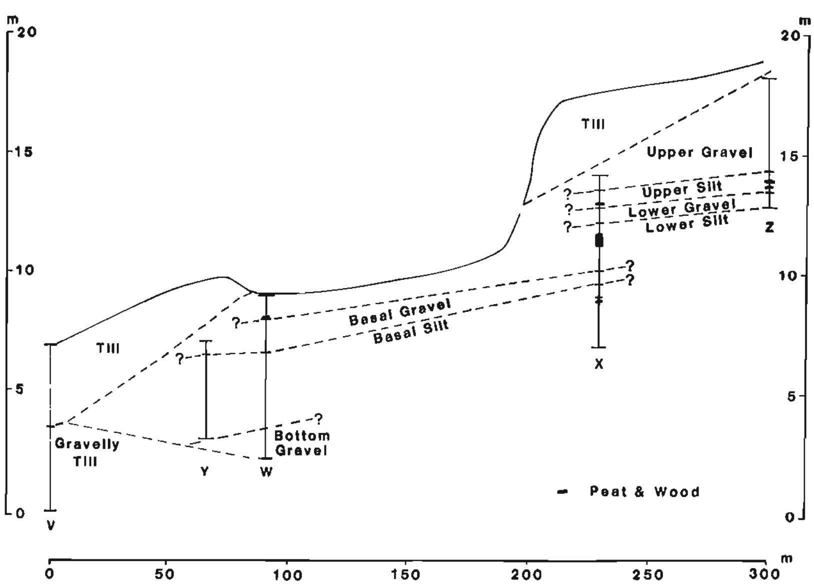

Figure 2. Cross-section showing the succession of deposits at Meadow Creek.

Basal silt

The unit referred to as basal silt is 340 cm thick at site Y, 300 cm at W, and at least 330 cm at X. In general it consists of fine grained sand and silt which occur as a series of roughly horizontal beds. Contacts between beds are irregular to undulating, and in detail the internal structure of the beds has been tilted, contorted, and disrupted. In some places silt and fine grained sand occur as rough couplets with sand units from 15 to 20 cm thick and silt 10 to 15 cm. The deformed bedding generally bends upwards as though the disruption was caused by loading and dewatering. Disseminated organic material appears to be most common in the upper part of the unit and at site X little organic material could be seen in the lower 200 cm. Compact peat and woody peat occurs in lenses up to 5 cm thick 60 cm below the top of the unit at site X. A stump 12 cm in diameter in growth position was collected from this part of the unit (Fig. 3). Sites Y and W contain several thin (approximately 5 mm) but distinct undulating seams of peat that extend across the entire exposure (Fig. 4). Even though no organic material could be seen near the base of the exposure at site X, one of the peat seams at site W is within 30 cm of the base of the unit.

Basal gravel

The basal gravel unit is interpreted as occurring at sites W, X, and Y (Fig. 2). It ranges in thickness from 150 cm at W to 45 cm at X. It is a pebbly, moderately well sorted gravel that consists dominantly of local quartzite and phyllite.

Figure 3. Stump in growth position at site X (cf. Fig. 2).

Lower silt

This unit was exposed at site X, where it is 182 cm thick; at site Z, where only the upper 20 cm was exposed; and at W, where the bottom 100 cm is assumed to be present. It is dominantly silt and fine grained sand and at site X contains abundant peat and woody peat. The very compact peat generally consists of well preserved mosses and contains seeds, leaves, including *Carex*, and some insects. The wood in the peat occurs as small rounded sticks and as larger compressed logs (up to 5 cm thick, 30 cm wide, and more than 1 m long). Pollen assemblages in the peat are dominated by *Picea*. Silt and sand occur as beds and seams within the peat and in places small lenses of fine gravel are present (Fig. 5). The base of the peat is not horizontal and possibly marks the floor of an abandoned channel. Individual peat and woody peat units are variable in thickness but were recorded as ranging from 7 to 16 cm thick. The main peat units are located 30 to 75 cm below the top of the unit.

Figure 4. Thin folded peat bed at site W (cf. Fig. 2).

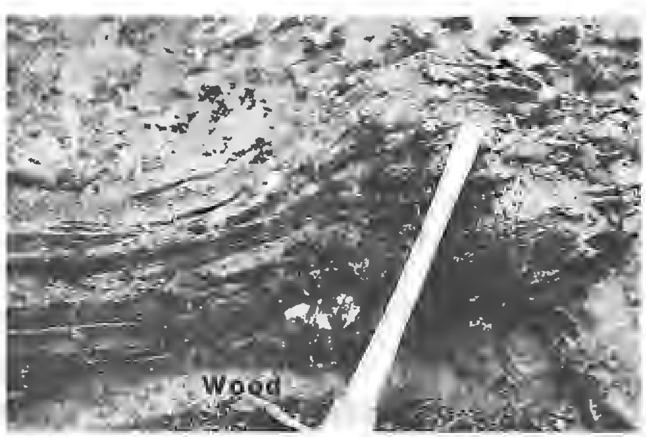

Figure 5. Woody peat containing lenses of sand and silt at site X (cf. Fig. 2).

Lower gravel

The lower gravel is 50-65 cm thick and consists dominantly of coarse sand and fine (granule) gravel which is mainly local quartzite and phyllite in composition. It occurs at sites Z and X (Fig. 2). At site Z a channel fill consisting of 25 cm of fine grained sand and silt occurs in the middle of the unit.

Upper silt

Approximately 90 cm of sand and silt underlies the upper gravel at site X and at site Z (Fig. 2). The upper part of this unit consists largely of reverse graded beds composed of silt or fine grained sand at the base and grading upwards to coarse grained and medium grained sand. The individual beds are 3-5 cm thick. The lower 30 cm contains 3 interstratified peat units that are 2-6 cm thick. The peat is mainly fine grained, partially humified and very compact. No wood was found in this peat.

Upper gravel

At site Z (Fig. 2) the highest unit exposed is a cobble gravel consisting largely of disc-shaped cobbles of green phyllite and fine grained quartzite. In earlier work this unit was referred to as "topset gravel and sand" (see unit 8 of Fulton, 1968). This unit is overlain by till which cuts out the gravel to the south. Only the lower part of this unit is present at the top of site X.

Till

Till was exposed in section V (Fig. 2). It consists of a yellow-olive sandy matrix containing abundant coarse clasts that in general consist of light green phyllite or fine grained quartzite, both of which are local. Two phases of till are present. The upper is as described above and the lower is an orthoconglomerate (clast supported framework) with a yellow-olive sand matrix in a framework consisting dominantly of rounded quartzite pebbles (this is referred to as gravelly till in Fig. 2). The gravelly till might be a deformation till (Dreimanis, 1982) developed from a gravel.

DISCUSSION

All units, with the exception of the bottom gravel and the gravelly till, were seen in previous exposures. Nothing that was seen during this investigation suggests that the earlier interpretation, that these deposits represent a succession of floodplain sediments (unit 7 of Fulton, 1968), requires modification.

It is suggested that the dense material that could not be penetrated at the base of the bottom gravel is a till. If this is the case, there is no evidence that a paleosol or other proof of a significant time break lies at the contact between the till and the overlying gravel. The bottom gravel in turn is a conformable part of a sequence of nonglacial Middle Wisconsinan deposits. This rather tenuous evidence might suggest that the Middle Wisconsinan sequence rests on Early Wisconsinan glacial deposits which were deposited a short time before 43 800±800 BP.

One item which warrants further consideration is the sporadic occurrence of wood in the floodplain sediments. Wood was found only in certain peat beds and even though finely disseminated organics were present throughout most of the fine grained sediment, pieces of wood were not found. Under natural modern conditions, stream floodplains in this region are thickly forested and overbank sediments include roots and in places buried pieces of wood. The presence of stumps in growth position and of sticks of all sizes in certain of the peat beds indicates that forest occupied the region during deposition of this succession. The lack of wood or evidence of tree growth in the fine sediments might indicate: 1) that the fine grained sediments were mainly deposited in semipermanent flood basins or 2) that the area was periodically flooded by a shallow lake. An alternative explanation might be that the apparent periodic disappearance of forest from the site was due to climatic change. The paleoecological studies should provide an answer to this puzzle.

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Revised stratigraphic and structural interpretation of folded décollements, southern Fraser River Antiform, Selwyn Range, British Columbia

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Mountjoy, E.W. and Grasby, S.E., Revised stratigraphic and structural interpretation of folded décollements, southern Fraser River Antiform, Selwyn Range, British Columbia, in Current Research, Part E, Geological Survey of Canada, Paper 90-1E, p. 359-367, 1990.

Abstract

The lower Miette Group consists of about 500 m of silver-grey to dark grey silty pelites with minor siltstone interbeds on the east limb of the Fraser River Antiform. The middle Miette Group is about 2700 m thick and contains the distinctive 70 m thick Old Fort Point Formation. The upper Miette Group is over 1000 m thick and consists of light to dark grey, rusty weathering pelites with mappable quartzites in its upper part.

A thrust fault, previously interpreted as a normal fault, is the southern extension of the Selwyn Range Décollement. On the southwest limb of the Fraser River Antiform, it follows a stratigraphic level near the base of the Old Fort Point Formation, and eastward cuts downsection until about 500 m of lower Miette Group strata occur in its hanging wall. In the window 500 to 700 m beneath the Selwyn Range Décollement, a thrust zone duplicating conglomerate and quartzite units in upper Miette Group pelites also cuts downsection northeastward. These relationships suggest that a regional fold was truncated by early thrusting.

Résumé

La partie inférieure du groupe de Miette est constituée d'environ 500 m de pélites silteuses dont la couleur varie de gris-argent à gris foncé, avec quelques interstratifications de siltstone sur le flanc est de l'antiforme de Fraser River. La partie moyenne mesure environ 2700 m d'épaisseur et renferme la formation caractéristique d'Old Fort Point, de 70 m d'épaisseur. La partie supérieur mesure plus de 1000 m d'épaisseur et est constituée de pélites gris clair à gris foncé prenant une couleur rouille à l'altération et qui sont accompagnées, au sommet du groupe, d'unités de quartzites que l'on peut cartographier.

Une faille inverse, interprétée antérieurement comme faille normale, constitue le prolongement sud du décollement de la chaîne de Selwyn. Sur le flanc sud-ouest de l'antiforme de Fraser River, cette faille suit le niveau stratigraphique situé à la base de la formation d'Old Fort Point et, en direction est, s'enfonce vers le bas jusqu'à ce qu'environ 500 m des couches inférieures du groupe de Miette apparaissent à la lèvre supérieure. Dans la fenêtre située 500 à 700 m au-dessous du décollement de la chaîne de Selwyn, une zone de chevauchement répétant des unités de conglomérat et de quartzite dans les pélites de la partie supérieure du groupe de Miette s'enfonce également vers le bas en direction nord-est. Ces relations semblent indiquer qu'un pli régional a été tronqué par un premier chevauchement.

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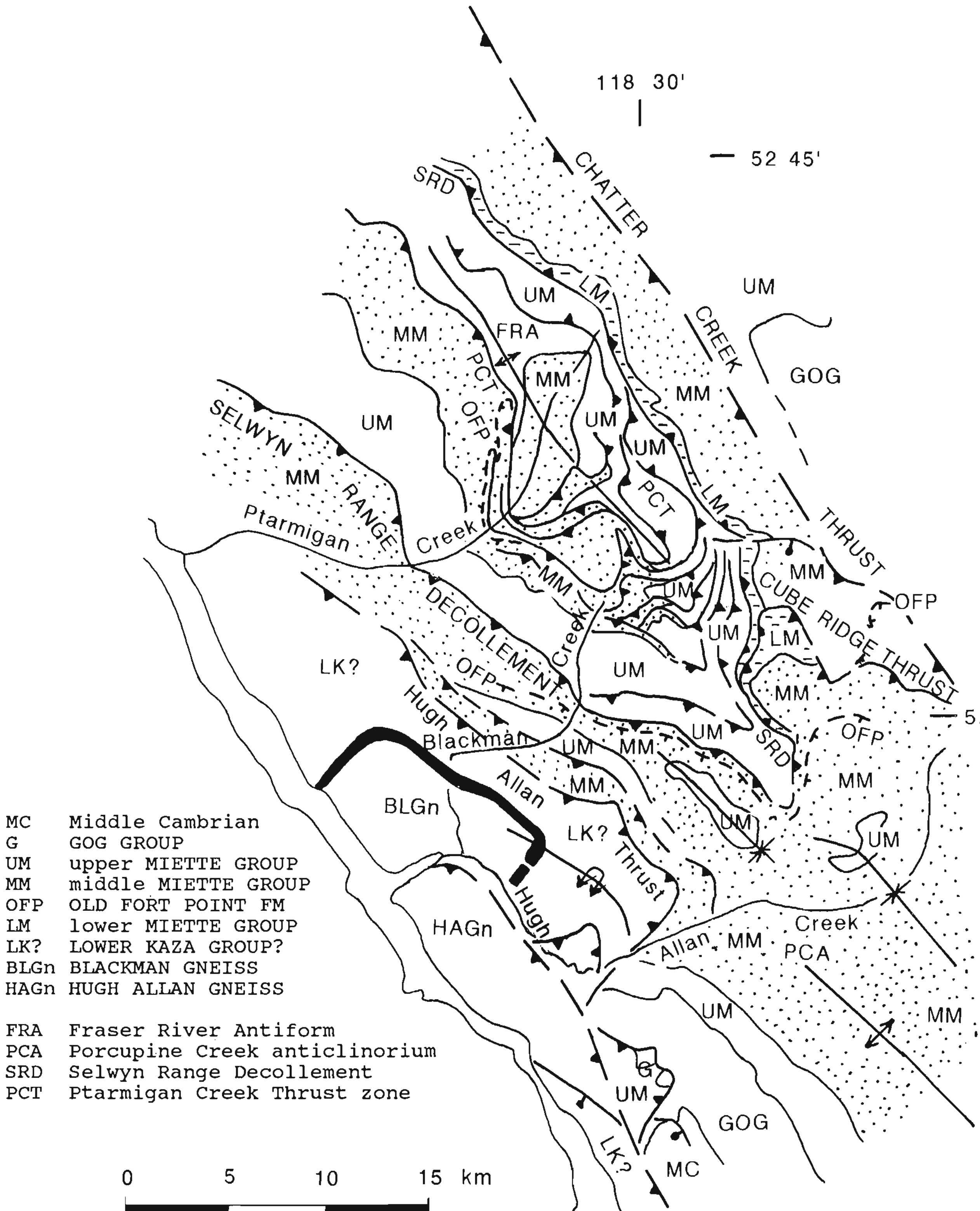

Figure 1. Regional geological map of the Hugh Allan — Ptarmigan Creek region, southern Selwyn Range. (Revised from Mountjoy and Forest, 1986).

INTRODUCTION

During the summer of 1989, portions of the southern Fraser River Antiform, containing the Selwyn Range Décollement (Dechesne, 1990; Dechesne and Mountjoy, in press) in its upper part, were examined and mapped between Blackman and Hugh Allan creeks in parts of map areas 83 D/8, 9, and 10. The stratigraphy of the Miette Group as well as fault zone structures were studied in detail.

This preliminary report describes the results of the mapping south to the main fork of Hugh Allan Creek and revises some of the stratigraphy and structure reported in Mountjoy et al. (1985) and Mountjoy and Forest (1986). The normal fault on the west limb of the Fraser River Antiform, which they and Leonard (1985) mapped, is herein shown to be a thrust fault (Selwyn Range Décollement of Dechesne (1990) and Dechesne and Mountjoy (in press), and strata in the footwall to the east are assigned to the upper Miette Group rather than the lower Miette Group pelites (Figs. 1, 2).

Farther northwest, an extension of this fault may link with the D_2 Packsaddle Thrust of McDonough and Simony (1986). The décollement also occurs on the east side of the Fraser River Antiform (Dechesne and Mountjoy, 1990), but was not recognized in the intervening area where McDonough and Simony (1988) mapped a conformable sequence. If our assessment is correct, the décollement should occur within their lower Miette Group strata where lower Miette Group pelites are faulted over upper Miette Group pelites (Dechesne, 1990; Dechesne and Mountjoy, in press).

STRATIGRAPHY

Miette Group

The Miette Group is divided into three informal map units: lower, middle and upper Miette Group (Campbell et al. 1973; Carey and Simony, 1985).

Lower Miette Group

Strata assigned to the lower Miette Group are dominantly silver-grey to dark grey pelites with abundant fine silt laminae (1-5 mm). Local 2 to 20 cm thick sand beds weather light brown to rust. The lower Miette Group occurs only on the east flank of the Fraser River Antiform in the hanging wall of the Selwyn Range Décollement. Only the upper part of the lower Miette Group occurs above the detachment, giving a minimum thickness of about 500 m.

Previously, Leonard (1985), Mountjoy et al. (1985), and Mountjoy and Forest (1986) placed in the lower Miette Group the dark grey to black pelites and associated underlying grits that occur in the footwall of the fault herein assigned to the Selwyn Range Décollement (Figs. 1, 3). These pelites differ in lithology and colour from those of the lower Miette Group and are now assigned to the upper Miette Group (Fig. 2) on the basis of distinctive carbonates and conglomerates occurring near the base. The underlying grits are assigned to the middle Miette Group, instead of lower Miette Group grits, on the bases of their occurring conformably below upper Miette strata and the presence of

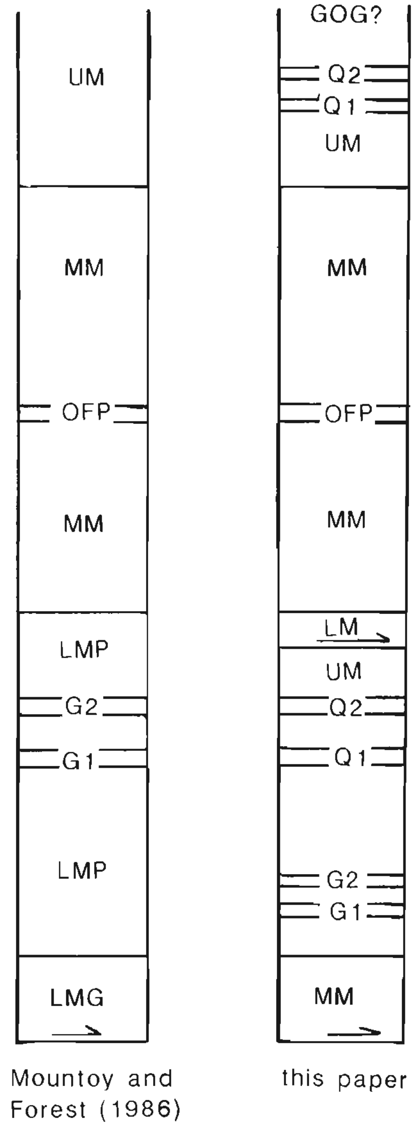

Figure 2. Comparison of revised stratigraphy with earlier stratigraphy of Mountjoy and Forest (1986), southern Selwyn Range.

the Old Fort Point Formation in the hanging wall of the Ptarmigan Thrust Zone (Dechesne, 1990) (Fig. 1). The only unequivocal section of lower Miette Group occurs to the north in the Cushing Creek area (Carey and Simony, 1985). It lacks mappable quartz sandstones and grits seen in the footwall pelites of the Selwyn Range Décollement and is feldspar poor, which adds support to our revised stratigraphic assignments of pelites in the southern Fraser River Antiform (see below).

Middle Miette Group

More than 2500 m of middle Miette Group are exposed in the study area. The middle Miette Group can be divided into three sequences: two thick sequences of dominantly pebbly sandstones (grits) divided by the distinctive thin carbonate and pelite marker, the Old Fort Point Formation (Fig. 2).

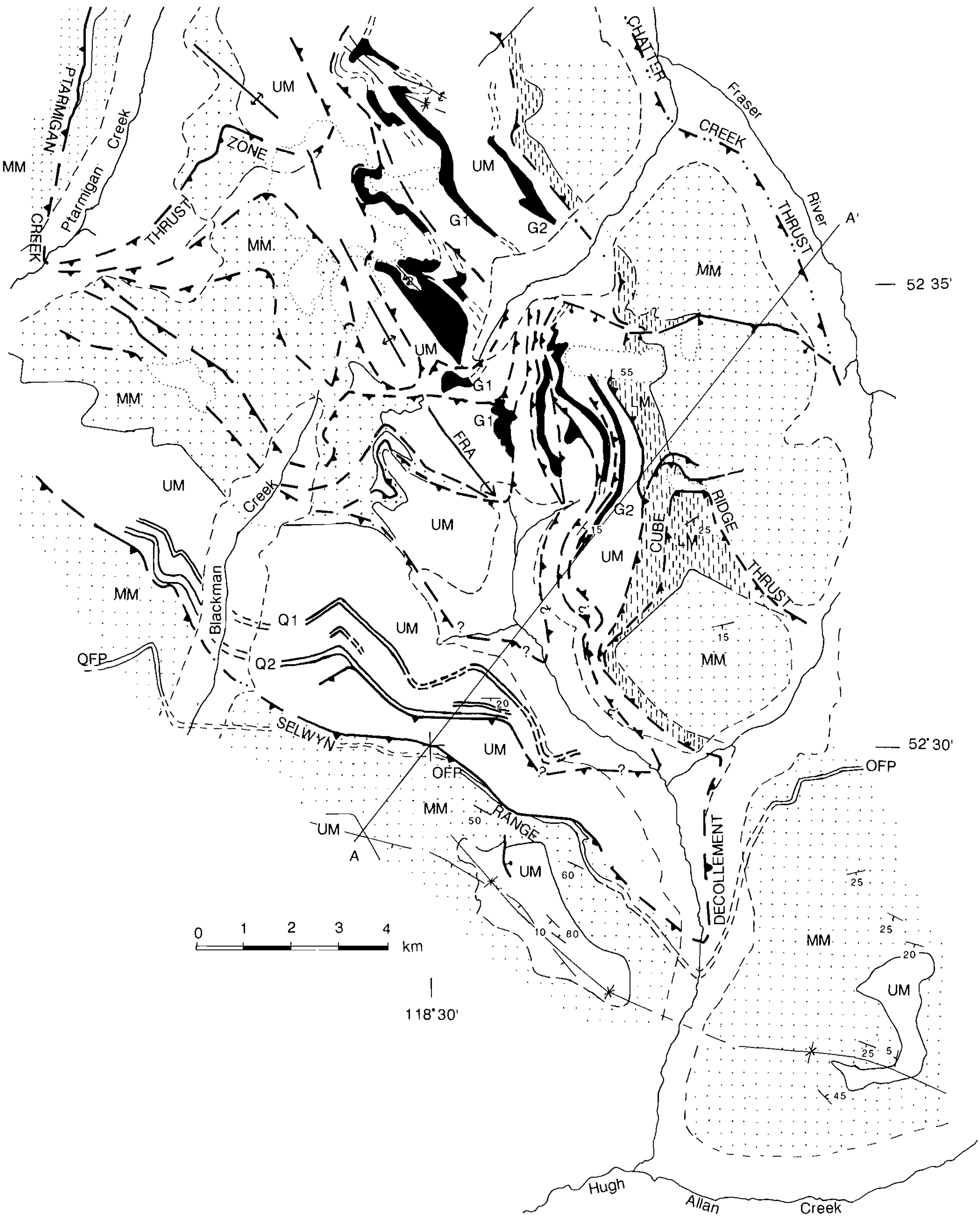

Figure 3. Revised geological map of Hugh Allan and Ptarmigan Creek areas (parts of 83 D/8, 9, and 10). Legend: FRA, Fraser River Antiform; UM, upper Miette Group; Q1 and Q2, quartzite units in upper part of upper Miette; G1 and G2, conglomerate and grit units in basal part of upper Miette; MM, middle Miette; OFP, Old Fort Point Formation; LM, lower Miette Group. Chloritoid-in isograd shown on extreme left. A-A', location of cross-section in Figure 4.

The lower middle Miette Group consists of pale to dark grey, poorly sorted, composite, graded pebble conglomerates 10 to 15 m thick, interbedded with green-grey to emerald green pelites 5 to 10 m thick. The lower middle Miette Group has a gradational contact with the underlying lower Miette Group. The contact is placed at the base of the lowest thick, massive, mappable grit unit, above which pelites are dominantly green-grey.

The Old Fort Point Formation is about 70 m thick in the hanging wall of the Selwyn Range Décollement. It conformably overlies the lower middle Miette Group and forms a distinctive tripartite sequence of basal bright green pelite, brown carbonate and upper black pelite. The lower 40 m of the Old Fort Point consist of maroon to emerald green silty pelites with local sandstone beds up to 15 cm thick. This is overlain by 5 m of green pelite with light brown, quartz rich, carbonate laminae (1-5 mm), grading into 14 m of light brown carbonate. The top 4 m of the carbonate consist of rhythmically bedded, gold-brown weathering carbonate (0.5-4 cm) and green pelite (1-5 mm). The carbonate is overlain by 5 m of black pelite. The Old Fort Point Formation can be traced around the southern Fraser River Antiform in the upper tectonic unit and also occurs north of Ptarmigan Creek in the lower tectonic unit (Dechesne, 1990, and pers. comm.). This sequence is similar to that described by McDonough and Simony (1988) in the Northern Selwyn Range and by Ross and Murphy (1988) in the Cariboo Mountains, giving it regional significance.

A continuous section of upper middle Miette strata occurs above the Old Fort Point Formation along both the western and eastern flanks of the Fraser River Antiform. A cross-section thickness estimate of 1500 m was obtained for the upper middle Miette Group on the east flank and about 1200 m on the west flank (Figs. 3, 4C). The upper middle Miette Group is dominated by thick (10-50 m), graded, composite pebble conglomerates interbedded with pale green to medium grey pelites 7 to 50 m thick, similar to the lower middle Miette Group sequence.

Upper Miette Group

The upper Miette Group is exposed throughout the area, showing considerable variation between the upper tectonic unit and the lower tectonic unit of the Fraser River Antiform (Figs. 1, 3). Along the western flank of the Antiform, in the hanging wall of the Selwyn Range Décollement, the upper Miette Group consists predominantly of light to dark grey, rusty weathering pelites with abundant fine silt laminae and occasional 10 to 20 cm thick brown quartzite beds. No carbonates were observed. A number of thin quartzite beds (1-2 m) occur 200 m above the base of the upper Miette Group. By contrast, in the footwall of the Selwyn Range Décollement, mappable quartzite units (Q1 and Q2) occur in the upper part, about 1000 m above the base of the upper Miette Group (Figs. 2, 3). Q1 and Q2 consist of relatively pure, fine to coarse grained, finely laminated to cross-bedded quartz sandstones. Locally, quartz pebble conglomerates, 2 to 5 cm thick, and containing some white feldspar, occur at the base of some of these quartzites.

A distinctive boulder conglomerate bed that occurs near the base of the upper Miette Group in different parts of the region is absent on the west flank, but occurs on the east flank in thrust imbricates (G1, G2) in the footwall of the Selwyn Range Décollement (Fig. 3). In the hanging wall of the Selwyn Range Décollement, approximately 300 m of the basal part of the upper Miette are exposed along the southeastern flank of the antiform. Access is difficult and it was only possible to examine the base. The gradational contact is placed at the top of the last thick (10 m) continuous middle Miette pebble conglomerate and grit, above which pelites are predominantly black with abundant silt laminae. At least four resistant quartzite beds (2-5 m) occur above the contact, separated by thick (20-25 m) sequences of black pelite. As on the west flank, no carbonates and no conglomerates or debris flows were observed at this locality. Our mapping and that of Dechesne (1990, and pers. comm.) indicate that the distinctive basal boulder conglomerate does not occur in the hanging wall of the Selwyn Range Décollement south of Blackman Creek. It is also absent south of Hugh Allan Creek (Klein and Mountjoy, 1988) and as far south as the Solitude Range (Gal et al., 1989).

On the eastern flank of the Fraser River Antiform, in the footwall of the Selwyn Range Décollement, the upper Miette Group consists predominantly of brown to black pelites with abundant silt laminae. A 10 m thick, light brown weathering quartzite unit outcrops along a ridge on the eastern flank (Fig. 3, 52°31′, 118°26′). Medium grained, light brown quartzite beds (20-60 cm) are interbedded with darker brown, more pelitic calcareous quartzite beds exhibiting large-scale soft sediment deformation structures. In the lower part of the quartzite, a series of boulder conglomerate beds occur with clasts (predominantly pelite and carbonate, plus quartz, rock fragments and feldspar) up to 1.2 m long and 60 cm in diameter. These conglomerates are similar to upper Miette Group basal conglomerates north of Ptarmigan Creek in the Ptarmigan thrust sheet (Dechesne, 1990). Underlying this quartzite is a sequence of dark pelites with occasional, thin, brown weathering sandstone beds. This is underlain in turn by a series of imbricate fault slices of upper Miette Group strata containing two mappable, lenticular, coarse conglomerate grit units 20 to 100 m thick (Fig. 3, G1 and G2 of Mountjoy and Forest, 1986).

In summary, the criteria that can be used in the southern Selwyn Range to distinguish between the two thick pelite sequences of the lower and upper Miette Group are as follows. The lower Miette Group: 1) contains no thick conglomerate or grit beds; 2) contains no thick, mature quartzites; and 3) colours range from dark grey to silvery grey. On the other hand, the upper Miette Group: 1) near the base includes locally mappable lenticular conglomerate and grit units (e.g., G1 and G2), along with scattered thin quartzite beds up to 5 m thick; 2) in the middle and upper parts contains abundant thin quartzites, some of which form units (about 80 m) that can be mapped over long distances, especially in the upper part, beneath the Selwyn Range Décollement (e.g., Q1 and Q2); and 3) colours are dark grey to black and the pelites weather a strong rusty colour. Mountjoy and Forest (1986) mistakenly correlated G1 with Q1 and G2 with Q2, which, as noted above, are very different lithotypes. In some sections it appears that quartzites may be a lateral facies of the coarse conglomerates (R. Dechesne and M. McDonough, pers. comm., 1989).

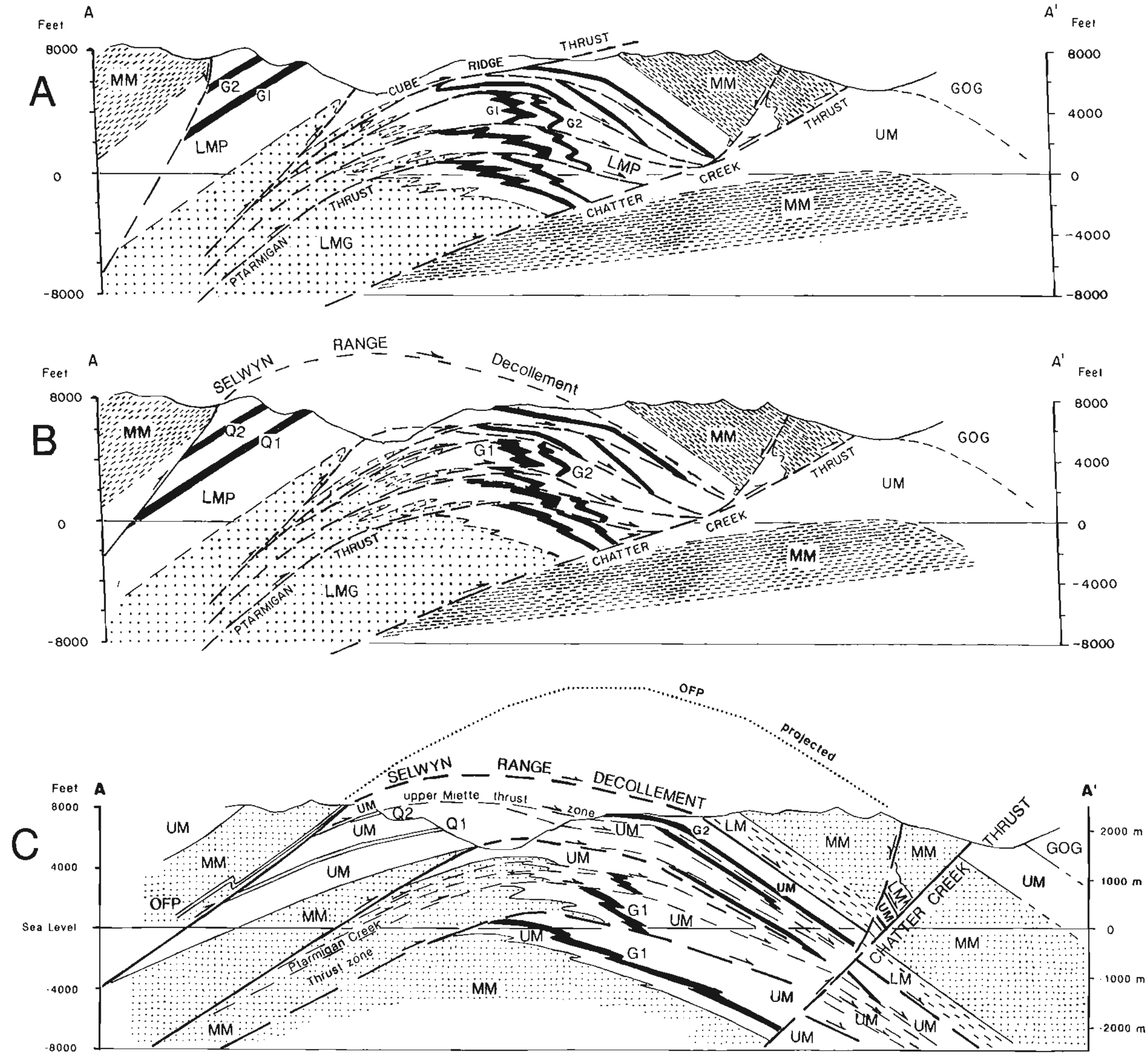

Figure 4. Cross-sections of southern Fraser River Antiform along A-A', Figure 3. **A.** Cross-section of Mountjoy and Forest (1986, Fig. 20.3). **B.** Modified cross-section of A. changing normal fault to Selwyn Range Décollement (SRD) but not changing the stratigraphy. Note that SRD cuts downsection northeastward in the hanging wall. **C.** Revised cross-section utilizing revised stratigraphy mapped in Figure 3. Both the SRD and the upper Miette Thrust Zone cut downsection to the northeast.

Uppermost upper Miette and/or basal (?)Gog Group

Clean quartzites generally do not form mappable units in much of the upper Miette of the Jasper-Yellowhead region, except near Yellowhead Pass and Astoria Creek (Teitz and Mountjoy, 1985, 1988). Where the quartzites form mappable, relatively pure quartzite units in the order of 40 to 80 m thick, these strata may represent the base of the Gog Group. This is especially true for Q2, which has been tectonically thickened to more than 300 m. It consists of relatively pure, medium to coarse grained, white, quartz sandstones containing at the base of some units quartz pebble (up to 4 cm) conglomerates with white feldspar fragments. Planar cross-bedding occurs locally. These quartzites form massive, resistant cliffs that weather light to dark grey when lichen covered. They are similar to the basal Gog Group, except that they lack the red hematitic colouring that is so prominent in much, but not all, of the basal Gog Group from Red Pass, British Columbia to Jasper. The top of Q2 is truncated by one of the thrust faults in the upper Miette Thrust Zone, and is overthrust by upper Miette pelites. Hence Q2 could be basal Gog (Figs. 3, 4C).

STRATIGRAPHIC CORRELATION AND INTERPRETATION

The suggested correlations between the northern and southern Selwyn Range of McDonough and Simony (1988, p. 108) concerning the lower Miette and the base of the middle Miette Group, are interpreted by us to be incorrect, since we now assign both G1 and G2 to the upper Miette Group.

Their (ibid.) mappable quartzite that is up to 80 m thick appears to correlate with Q2, which we assign to either the uppermost upper Miette Group or basal (?)Gog Group. So far we have not observed black silty limestones associated with these quartzites. These strata require more detailed study. If this suggested correlation is correct, it means that a window of upper Miette Group strata also occurs in the northern Selwyn Range beneath the middle Miette Group and a thin lower Miette pelite. As noted below under structure, the hanging wall has moved eastward relative to its footwall and has cut an earlier fold.

STRUCTURE

Fraser River Antiform

The dominant structure of the area is the Fraser River Antiform (Figs. 1, 2) which is a major F_2 fold that has folded the earlier F_1 thrust structures: 1) the lower Ptarmigan Creek Thrust Zone, 2) a middle thrust zone that duplicates the upper part of the upper Miette, and 3) the Selwyn Range Décollement (Figs. 3, 4C). The Ptarmigan Creek Thrust Zone outcrops to the north and its structure has been outlined by Mountjoy et al. (1985), Forest (1985), Mountjoy and Forest (1986), and Dechesne (1990).

Selwyn Range Décollement

Along the west limb of the Fraser River Antiform, the well exposed Old Fort Point Formation and some underlying middle Miette Group grits are locally cut out, indicating a fault riding just below this stratigraphic level (Fig. 3). Earlier, this was mapped as a normal fault by Leonard (1985) and Mountjoy and Forest (1986). Over an interval of about 100 m, strata above and below the fault are intensely deformed, with deformation increasing toward the fault. Minor folds outlined by bedding (S_0) have a northeast vergence with strongly developed, axial planar, southwest dipping shear zones locally crenulated by S_2 (Fig. 5). Thus the hanging wall rocks were thrust to the east prior to the development of the Fraser River Antiform that crenulated the shear zones. The southern and eastern flanks of the Fraser River area were traversed in detail and no faults were observed. Thus the fault must underlie the middle Miette and be folded by the Fraser River Antiform (Fig. 3). This thrust is correlated with the Selwyn Range Décollement mapped to the north by Dechesne (1990).

The Selwyn Range Décollement is also exposed along the east flank; however, it is more difficult to trace as it places pelite over pelite in what may appear to be conformable sequences. Where exposed beneath the Cube Ridge Thrust (52°31′, 118°26′; Fig. 3), the fault zone is characterized by a highly deformed area with relatively undeformed pelites above and below. Hanging wall rocks comprise about 500 m of light grey, silty pelites of the lower Miette Group. These pelites grade upward over a few metres into thick grit sequences of the middle Miette Group, whereas footwall rocks are black to brown pelites with thin quartzite and conglomerate units characteristic of the upper Miette Group.

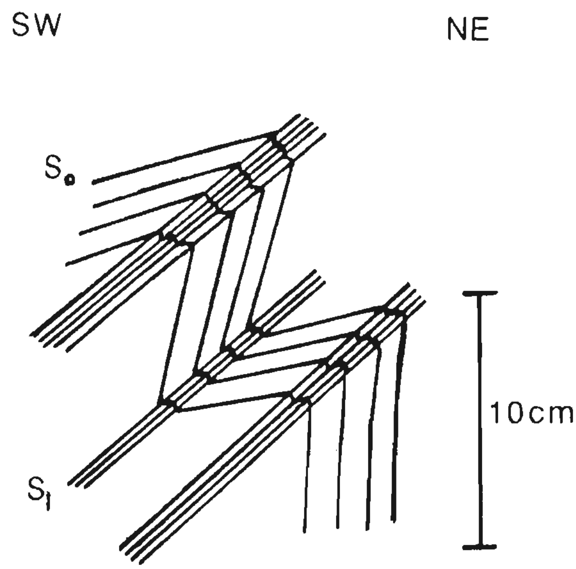

Figure 5. Sketch of microstructures from a hand specimen, illustrating structures related to the hanging wall of the Selwyn Range Décollement west of the north fork of Hugh Allan Creek. Folds outlined by bedding S_0 verge northeast with strongly developed, axial planar, southeast-dipping shear zones that are locally crenulated by S_2, which is related to development of the Fraser River Antiform.

On the east flank, the Selwyn Range Décollement is situated approximately 500 m below the base of the middle Miette Group. Thus the décollement cuts downsection from west to east across the range (Fig. 4), contrary to the "foothills rules" outlined by Dahlstrom (1970). This configuration holds for both stratigraphic interpretations (thin and thick stratigraphic successions, Fig. 4B, C), and, if it is correct, indicates that the Miette Group was broadly folded prior to initiation of the Selwyn Range Décollement. In the footwall, this early fold must occur somewhere downdip to the southwest and be of regional scale (Dechesne, 1990; Dechesne and Mountjoy, 1990; Dechesne and Mountjoy, in press).

Cube Ridge Thrust

The Cube Ridge Thrust is a low-angle fault (35°) of minor displacement (Fig. 3). With more detailed mapping, part of the trace of the southwestern limit of the thrust has been shifted about 1 km eastward from the position reported by Mountjoy and Forest (1986). The fault parallels hanging wall strata and cuts out some middle Miette grits in the footwall. Thin mylonite zones associated with the fault are crenulated by an S2 cleavage; therefore, the fault must be pre-Fraser River Antiform and not a late stage fault as interpreted by Mountjoy and Forest (1986). Although it is difficult to trace the fault southwestward, it is inferred to merge with and therefore be a splay from the Selwyn Range Décollement (Figs. 3, 4C).

Upper Miette Thrust Zone

The upper Miette is strongly imbricated in the footwall of the Selwyn Range Décollement on both the west and east flanks of the Fraser River Antiform (Fig. 3). On the west flank the uppermost upper Miette Group (or base of (?)Gog Group) Q1 and Q2 units are imbricated, whereas on the east flank the lower conglomerate and conglomeratic grits (G1 and G2) of the lower upper Miette Group are imbricated, suggesting that the faults cut downsection eastward through most of the upper Miette Group (Fig. 4C). This is similar to the cutting-down-eastward of the overlying Selwyn Range Décollement. Both of these structures suggest the presence of earlier, large, open F_0 folds.

On the east flank of the Fraser River Antiform, the upper Miette Group imbricates cannot be traced into or connected with the underlying middle Miette Group strata (Fig. 3); there are no faults of sufficient displacement with which to connect them. Rather they apparently remain within the upper Miette Group, and the only faults with which they can be connected are those that duplicate the most upper Miette (or base of (?)Gog Group) across the north fork of Hugh Allan Creek on the west flank of the Fraser River Antiform (Fig. 3).

The upper Miette Thrust Zone may reflect accretion of material or underplating beneath the Selwyn Range Décollement as a result of the downward migration of the basal décollement (Dechesne and Mountjoy, in press).

Chatter Creek Thrust

The Chatter Creek Thrust is a steep fault of small displacement and therefore must cut across the Selwyn Range Décollement, upper Miette and Ptarmigan Creek Thrust zones at depth (Fig. 4C), rather than partly merging with and underlying the Ptarmigan Creek Thrust Zone as intrepreted by Mountjoy and Forest (1986) (Fig. 4A). The total displacement on the three fault zones is large and, therefore, these faults must connect with thrust faults in the eastern Main Ranges and possibly the western Front Ranges (Dechesne and Mountjoy, in press). The Chatter Creek Fault is interpreted as a late stage, out-of-sequence fault (Dechesne and Mountjoy, 1988, 1990).

DISCUSSION AND INTERPRETATION

Three different structural interpretations are contrasted in Figure 4; the first two (A and B) using the old stratigraphic assignments (thick stratigraphy), and the third (Fig. 4C) using the revised stratigraphy (thin stratigraphy) shown on the geological map (Fig. 3). The earlier interpretation of Mountjoy and Forest (1986; Fig. 4A) is clearly incorrect because the Selwyn Range Décollement is definitely a thrust fault and not a normal fault. The thick stratigraphy option involving a lower Miette pelite and grit sequence is untenable for a number of reasons:

1. It does not explain the thick succession of pelites along much of the core of the Fraser River Antiform. These pelites have been thickened enormously both by folding and faulting.
2. The stratigraphy of the 'lower Miette' pelite contains elements that occur in the upper Miette, such as mappable quartzite units near the top and lenticular conglomerates associated with carbonate rocks near the base.
3. The Old Fort Point Formation occurs in the 'lower Miette' grits above the Ptarmigan Creek Thrust Zone (Fig. 1), thus making them truly middle Miette grits.

Currently we are left with what to us is a viable interpretation (Fig. 4C), the one involving thin stratigraphy. This means that the major thrust faults exhibit the unusual relationship of cutting downsection northeastward in the direction of transport. Such phenomena are relatively rare in the Canadian Rocky Mountains but occur locally in the Front Ranges along the Boule Thrust (Mountjoy, 1960, 1976). Early phase, westerly directed, small-scale folds occur east of the Rocky Mountain Trench (Leonard, 1985; Mountjoy et al., 1985), and large-scale nappes occur west of the Trench (Raeside and Simony, 1983). Such an early, large fold could have been truncated by the Selwyn Range Décollement and the upper Miette Thrust Zone. Truncation of an early fold structure would satisfactorily explain the stratigraphic and structural relationships shown in Figures 3 and 4C.

What formed the Fraser River Antiform is not clear as it depends on the origin of the structures that underlie the antiform at depth. Interpretation of the structure at depth is difficult. Dechesne and Mountjoy (in press) discuss three different options in filling the "hole" between the basement décollement and the surface Fraser River Antiform. Our mapping rules out the thick stratigraphy option, which involves the presence of a thick lower Miette Group pelite and grit unit (Mountjoy and Forest, 1986). Filling the interval from the surface to the basement requires the presence of a stack of thrust faults. Whether basement is involved in this structure at depth cannot yet be decided. Regardless of basement involvement, the shortening involved in the Fraser River Antiform is considerable and is more than the width of the present Main ranges outcrop. Dechesne and Mountjoy (in press) suggest that the antiform and the Jasper-Yellowhead structural culmination are related to thickening of the Miette Group along a basal décollement and progressive décollement abandonment.

METAMORPHISM

Most of the rocks in the area mapped appear to be subgreenschist facies. Leonard (1985) mapped the chloritoid-in isograd (Fig. 3) within the upper Miette Group along the west flank of the Fraser River Antiform in the hanging wall of the Selwyn Range Décollement.

CONCLUSIONS

The Selwyn Range Décollement and the upper Miette Thrust Zone were mapped around the southern plunge of the Fraser River Antiform, causing previous structural interpretations of this area to be modified. Stratigraphic constraints indicate that these fault zones cut downsection to the east and to the north, implying that a regional-scale folding event preceded these thrusts. Including the Ptarmigan Creek Thrust Zone to the north (Fig. 1), three structural detachments are exposed in the core of the Fraser River Antiform.

Lower Miette Group pelites are distinguished from upper Miette Group by the lack of mappable conglomerate and quartzite beds, as well as by their lighter colours. The stratigraphy and structure within the upper Miette Group beneath the Selwyn Range Décollement requires more detailed investigation.

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Preliminary biostratigraphic determinations for Middle Cambrian strata in the Dezaiko Range, east-central British Columbia

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Abstract

About 1250 m of strata previously assumed to be of Middle Cambrian age outcrop in the Dezaiko Range. Recovery of 5 trilobite-bearing collections provides the first biostratigraphic evidence of this. Lithological study and preliminary fossil identifications suggest that deep-water conditions prevailed during the earliest Middle Cambrian in this area, while a unit correlative with the basal Chancellor Formation was deposited. There followed a gradual shoaling to a well differentiated system of shallow subtidal and peritidal environments with progradation of the Snake Indian carbonate platform, which persisted through the ***Plagiura-Poliella*** *and* ***Glossopleura*** *Zones. A more uniform limestone and dolostone sequence, correlative with the Eldon and Pika formations, was deposited during the latter part of the Middle Cambrian.*

Résumé

Environ 1250 m de roches sédimentaires datant présumément du Cambrien moyen sont conservées dans la chaîne Dezaiko. La découverte de cinq niveaux contenant des trilobites fournit en fait la première preuve biostratigraphique à cet effet. Une étude lithologique et l'identification préliminaire des trilobites semblent indiquer que des conditions de mise en place en eau profonde prévalaient au début du Cambrien moyen, lors de la sédimentation d'une unité corrélative à la formation basale de Chancellor. Avec la progradation de la plateforme de roches carbonatées de Snake Indian, les conditions sont devenues de moins en moins profondes et un système différencié de milieux péritidaux et subtidaux s'est developpé. Ces conditions ont persisté durant la période s'étendant des zones ***Plagiura-Poliella*** *à* ***Glossopleura****. Par la suite, des calcaires et dolomies plus uniformes, corrélatifs aux formations d'Eldon et de Pika, ont été mis en place au cours de la dernière partie du Cambrien Moyen.*

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INTRODUCTION

Large tracts of Cambrian strata have been mapped in the Dezaiko Range of east-central British Columbia (Campbell et al., 1973; Taylor and Stott, 1981; McMechan and Thompson, 1985; McMechan, 1986). One generalized stratigraphic section for the northern part of the Dezaiko Range was published by Slind and Perkins (1966). This part of the Rocky Mountains, some 100 km northwest of McBride (Fig. 1), is more or less on strike with better known Cambrian units in the Mt. Robson-Jasper and Banff-Field areas. Correlation of the Middle and Upper Cambrian portions between these regions, however, has been hampered because of lithological variation and the lack (hitherto) of reliable biostratigraphic data from the Dezaiko Range. This paper reports preliminary identifications of trilobites collected from the Middle Cambrian interval of one composite section in the central part of the Dezaiko Range (Fig. 2).

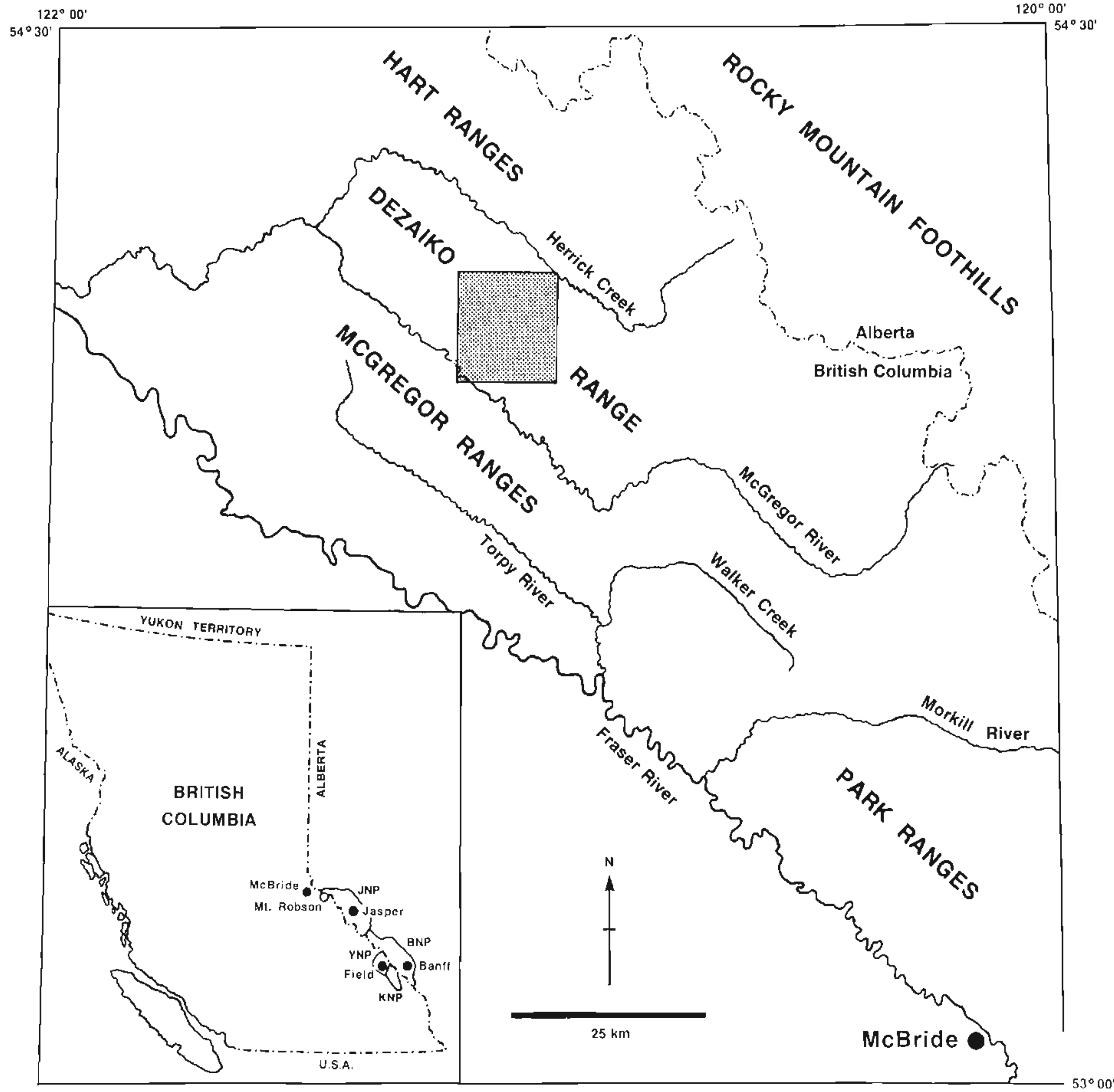

Figure 1. Location map of the study area in the Dezaiko Range, northwest of McBride, British Columbia. (Initials in the inset map, JNP, BNP, YNP and KNP, denote Jasper, Banff, Yoho and Kootenay National Parks, respectively.)

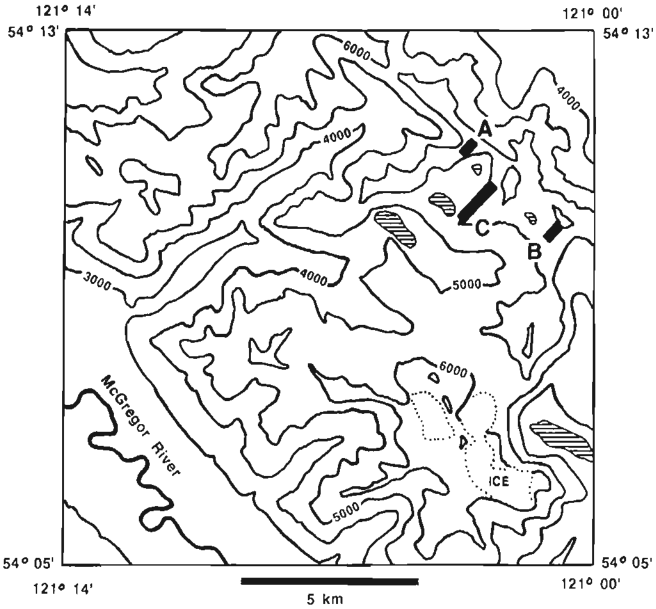

Figure 2. Simplified topographic map of the study area northeast of MacGregor River, showing location of the three measured sections used to produce the composite section depicted in Figure 3. Contour interval is in feet.

GEOLOGY OF THE STUDY AREA

Lithologies and formations

Middle Cambrian strata in the Dezaiko Range were subdivided into three thick units by Slind and Perkins (1966). Recent mapping (McMechan and Thompson, 1985; McMechan, 1986) and observations during this study indicate that the sequence lying above the Lower Cambrian Gog Group and below the Upper Cambrian Lynx Group is composed of a series of distinct units (Fig. 3) that correspond well lithologically with formations known in the national parks corridor to the southeast.

In the central part of the Dezaiko Range, the unit above the Gog is a heterogeneous, mainly carbonate sequence, referable to the Hota Formation, as defined in the Mt. Robson area (Fritz and Mountjoy, 1975). This unit is abruptly overlain by a resistant unit composed of parallel laminated siltstone and fine grained sandstone, 120 m thick, which grades upward into 175 m of interbedded, variably calcareous siltstone and sandstone. This sequence has a deep water, slope aspect, and can therefore be correlated with the Naiset Member of the Chancellor Formation (Stewart, 1989). The succeeding unit is about 215 m thick and consists of shallow, subtidal and peritidal, locally rippled or burrow-mottled, dolomitic, and variably silty limestone with rare oolitic beds and horizons of thrombolite patch reefs, capped by a distinctive, red, mudcracked siltstone. This unit appears similar to the Snake Indian Formation at its type section east of Jasper (Mountjoy and Aitken, 1978). Comparisons between it and the Chetang and Tatei formations, defined near Mt. Robson (Fritz and Mountjoy, 1975), are difficult to make at present and are under study (Pratt, in preparation).

Above the mudcracked siltstone is a 385 m thick, relatively homogeneous unit, dominated by lime mudstone but with peritidal cycles at the base containing oolitic grainstone, which correlates with the Eldon Formation from the Banff-Field area (Aitken et al., 1972) and the Titkana Formation of Mt. Robson (Fritz and Mountjoy, 1975). This is overlain by another relatively monotonous sequence, about 325 m thick, of largely dolomitized, variably silty, bioturbated lime mudstone, laminated grainstone and rare mudcracked beds and thrombolite patch reefs. These facies are arranged in a more or less cyclic fashion and a correlation with the Pika Formation to the southeast (Aitken et al., 1972) seems reasonable. Above these strata are 30 m of mudcracked, dolomitic siltstone and silty dolostone that can be assigned to the Arctomys Formation, albeit much thiner than at the type section (Aitken and Greggs, 1967).

Middle Cambrian strata have been reported to pass southeastward into a 500 m thick, "Chancellor-like" unit of silty limestone, overlain by about 250 m of limestones of the Titkana Formation (Campbell et al., 1973), which are exposed at Walker Creek, 35 km from the Dezaiko section. A brief visit to this locality as part of this study revealed that these limestones exhibit characteristics suggestive of moderately shallow to intermediate subtidal depths, rather than the deeper slope environment envisaged for the Chancellor (Stewart, 1989).

Trilobite collections

Trilobites were recovered from numerous horizons, in the Hota Formation, in the middle Chancellor, in the lower and upper Snake Indian, upper Eldon and lower Pika (Fig. 3). These trilobites are the first fossils collected from pre-Lynx strata in the Dezaiko Range. Raasch (*in* Slind and Perkins, 1966) reported a small Upper Cambrian trilobite faunule from the Lynx Group in a section about 20 km to the northwest of the section measured here.

BIOSTRATIGRAPHY

The current trilobite-based biostratigraphic zonation of the North American Lower and Middle Cambrian (Palmer, 1977) is still comparatively unrefined. Furthermore, it has been recognized, from species distributions in the Great Basin of the United States, that at least two concurrent Middle Cambrian zonal schemes are needed to take into account depth-related biofacies differentiation across the carbonate shelf and adjacent slope and basin (Palmer and Campbell, 1976; Robison, 1976).

Trilobite collections from the Hota Formation of the Dezaiko Range contain olenellids indicating the *Bonnia-Olenellus* Zone of late Early Cambrian age, which is consistent with the determinations in Fritz and Mountjoy (1975) for this unit at Mt. Robson. *Syspacephalus* sp., recovered from the middle Chancellor, suggests an earliest Middle Cambrian age, in the *Plagiura-Poliella* Zone, or its deep water equivalent, the lowest *Oryctocephalus* Zone. The fauna containing *Plagiura* sp., *Poliella* sp., *Vanuxemella* sp., and *Mexicella* sp. in the overlying basal Snake Indian is a shallow-water assemblage belonging to the undivided *Plagiura-Poliella* and *Albertella* Zones, and therefore supports correlation with the Mount Whyte and lower Cathedral formations in the Banff-Field area (Rasetti, 1951; Fritz, 1971; Aitken et al., 1972).

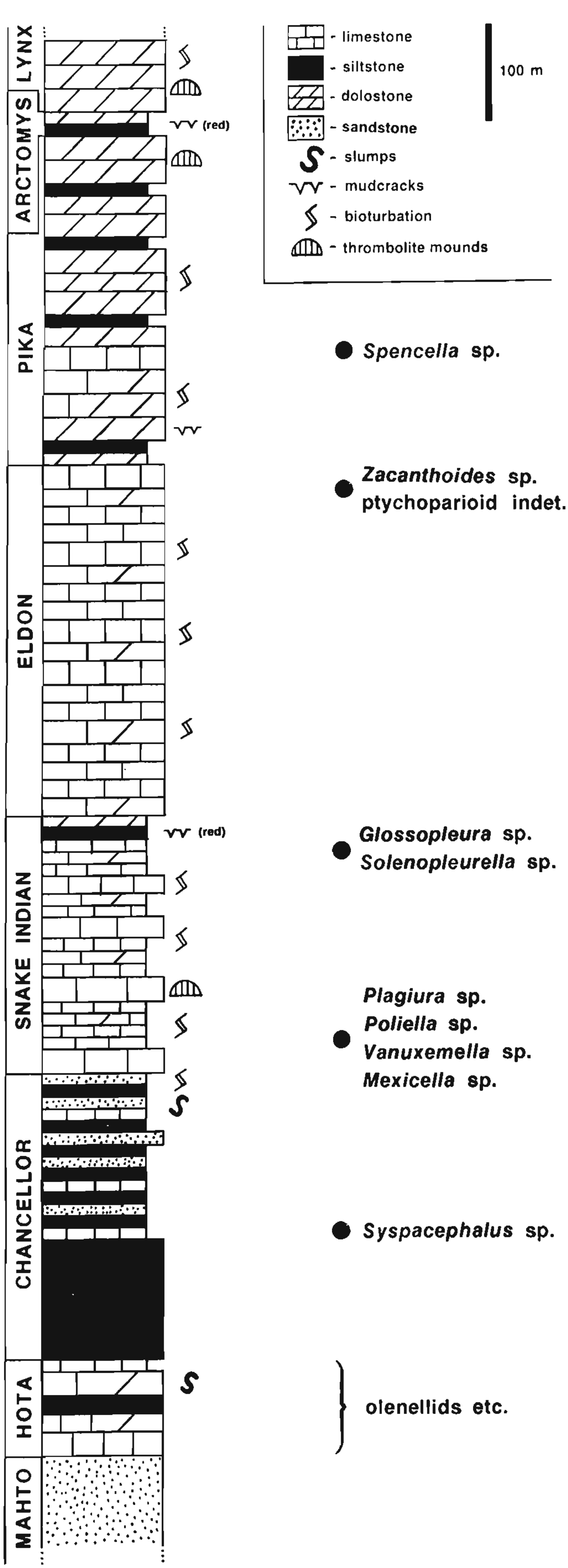

Figure 3. Composite measured section, showing generalized lithologies and weathering profile, suggested formation names, location of trilobite-bearing samples, and provisional generic identifications. Lithologies and thicknesses are based in part on observations by M.E. McMechan (unpublished).
The Hota Formation was measured at Section A; the Chancellor and Snake Indian formations were measured at Section B; and the Snake Indian, Eldon, Pika and Arctomys formations were measured at Section C.

The fauna composed of *Glossopleura* sp. and *Solenopleurella* sp., at the top of the Snake Indian Formation, indicates the *Glossopleura* Zone and correlation with the lower Stephen and uppermost Cathedral formations. This may imply that the base of the Eldon in the Dezaiko Range is slightly older than it is in the Banff-Field area. Collection of *Zacanthoides* sp. and an unidentified ptychoparioid from the upper Eldon, and *Spencella* sp. from the lower Pika, are not sufficiently diagnostic for precise comparison; especially given the generally poor documentation of trilobite distributions in the *Bathyuriscus-Elrathina* Zone and most of the *Bolaspidella* Zone in the North American Cordillera. Further work is under way to address the biofacies control on faunal patterns and refine Middle Cambrian trilobite-based biostratigraphy in Western Canada.

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