

LITHOPROBE

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and

CORDILLERAN TECTONICS WORKSHOP

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**LITHOPROBE SLAVE/NORTHERN CORDILLERA
LITHOSPHERIC EVOLUTION WORKSHOP
and
CORDILLERAN TECTONICS WORKSHOP**

Pacific Geoscience Centre/Institute of Ocean Sciences,
Sidney, British Columbia
February 21-23, 2002

FRIDAY, February 22, 2002

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Preface

This volume includes the abstracts of and short articles submitted for the 2002 LITHOPROBE Slave – Northern Cordillera Lithospheric Evolution (SNORCLE) transect meeting and Cordilleran Tectonics Workshop. These meetings were held concurrently at the Pacific Geoscience Centre. Accordingly, this volume includes several contributions related to the southern Cordillera, the northern Cordillera and the Slave province. We sincerely thank Elissa Lynn for her assistance in organizing the meeting and compiling this volume.

Frederick A. Cook and Philippe Erdmer

***ABSTRACTS of
ORAL
PRESENTATIONS***

Investigations along the Slave portion of the SNORCLE Transect

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Introduction

I will report on a variety of investigations in the Slave that are relevant to the SNORCLE transect.

Rock property studies

Circa 60 samples of major and minor rock types along the Slave portion of the SNORCLE transect were submitted to Matt Salisbury's rock property lab for density and acoustic property measurements. Density determinations are to complement earlier determinations that are used to support the gravity modelling of profiles across the Yellowknife Domain, particularly the Sleepy Dragon Complex (Bleeker et al., 1998; Hendry et al., 1999).

Results provided by Matt Salisbury show the standard Nave-Drake relationship (V_p vs. density) with some interesting outliers. The main distribution is bimodal: mafic versus felsic rocks. The "felsic" group comprises most if not all of the granitoid samples, moderately to highly deformed basement gneisses and granitoids, intermediate volcanic porphyries, and a broad selection of metasedimentary rocks, including Burwash Formation. Although Burwash Formation samples show slightly higher densities than the granitoids, and stronger anisotropies, their impedances are comparable and thus explaining the general transparency of the upper crust.

Mafic volcanics have high V_p 's (ca. 7 km/sec, at 200 MPa) and high impedances. Melanocratic gabbro of the Duck Lake sill plots at the high impedance range of the mafic group, thus confirming that this sill should be among the most prominent reflectors east of Yellowknife (Bleeker, 1999c). Outliers at the high impedance range are banded iron formation and calc-silicate rocks of the Central Slave Cover Group (Bleeker et al., 1999a,b), which would tend to enhance the reflectivity along the basement-cover contact. A kimberlitic sample (Snap Lake) is a distinct outlier at the low end of the range (density: 2.52 gram/cc, V_p : 4.87 km/sec).

These data support the conclusion that the main reflectors in the Yellowknife area (Y1 and Y2; names from Cook et al., 1999) are imaging the Duck Lake sill (Y1) and the mafic volcanic stratigraphy (plus Central Slave Cover Group) overlying felsic basement at depth (Y2), rather than major faults (cf. Cook et al., 1999). This is further supported by velocity modelling of Dave Snyder that shows that Y1 and Y2 are major positive velocity anomalies (Snyder et al., 1999; Snyder and Bleeker, in prep.).

Regional structural geology

On another front, further structural and stratigraphic analysis of first-order geological features of the Yellowknife Domain lead to the following conclusions (expanded from GAC abstract, Bleeker, submitted):

Archean granite-greenstone terrains are characterized by basalt dominated volcanic packages (ca. 2.9-3.1 gram/cc) overlying a middle to lower crust dominated by gneisses and granitoids of felsic to intermediate composition (ca. 2.6-2.8 gram/cc). This broad stratification may arise from voluminous basalt extrusion across extended sialic basement (e.g., Slave, Dharwar, Zimbabwe cratons) or, in the absence (?) of older sialic basement, by syn- to late-volcanic intrusion of voluminous tonalite-trondhjemite-granodiorite (TTG) plutons (e.g., parts of the Wabigoon subprovince, Superior craton). Once established, this stratification represents a significant density inversion (0.1-0.3 gram/cc).

Unlike regional tectonic processes, which can be “turned on or off” depending on external boundary conditions, buoyancy-driven processes can merely be retarded by low temperature (i.e. low viscosity); they cannot be “turned off” as long as a significant density inversion remains. Higher heat production in the Archean (2-4 times present values, e.g. Pollack, 1997) dictates that conditions for partial convective overturn were reached more readily in the Archean than in the Phanerozoic. Consequently, buoyant ascent of composite granitoid-gneiss domes and relative sinking of greenstones was ubiquitous during the Archean, irrespective of the nature of externally imposed tectonic processes (e.g., Macgregor, 1951; Anhaeusser, et al., 1969; Anhaeusser, 1971, 1984; Mareschal and West, 1980; Hickman, 1983, 1984; Collins, 1989; Collins et al., 1988; Bleeker, submitted).

The tell-tale signs of this process are composite granitoid-gneiss domes, 40-60 km in diameter (or somewhat larger if viscosities in the lower crustal feeding zone are greatly reduced), and intervening synclines or synclinoria with dense greenstones. This dome diameter, which fits observations across several cratons, is a straightforward prediction from Rayleigh-Taylor instability theory and scales primarily with thickness of the crust. Examples are, among many others, the Kenogamissi and Round Lake batholiths in the Abitibi greenstone belt, the Anton and Sleepy Dragon domal complexes in the Slave craton, and the spectacular domes of the eastern Pilbara craton (e.g. Hickman, 1983, 1984). Each likely developed through an interplay between buoyancy-driven processes and regional deformation (Bleeker, submitted). Dome amplification is likely to be most dramatic during intrusion of late-stage, crustally derived granitoid magmas, when the lower crust was hot, partially molten, thus characterized by very low viscosities and able to maintain lateral flow to the growing domes.

Modeling shows that rapid dome ascent should be accompanied by significant surface uplift (e.g. Mareschal and West, 1980), and sinking of denser supracrustals in flanking synclines. Increased erosion is thus expected above the domes, coupled to coarse clastic sedimentation in flanking basins.

Because “Timiskaming-type” clastic basins 1) are restricted in volume, 2) typically occur as tight to isoclinally folded synclinal panels on the flanks of high-amplitude domes, 3) were deposited broadly coeval with late-stage plutonism, 4) are filled with locally derived conglomerates and cross-bedded sandstones, 5) contain prominent granitoid clasts and detrital zircons indicative of erosional unroofing of nearby granitoid complexes, and 6) were deposited generally late in the evolution of the greenstone belt but just in time to experience the last thermal metamorphic pulse (heat conducting out from the rising domes), I propose that these deposits represent the sedimentary response to the ascent of Archean granitoid-gneiss domes. Hence, these deposits are as characteristic for granite-greenstone terrains as are large composite granitoid-gneiss domes. Regional strike-slip faulting may be involved but largely overprints and truncates the clastic basins (e.g., Bleeker, 1999).

In a Canadian context, prominent examples are Timiskaming Group rocks in the Abitibi greenstone belt, the Cross Lake Group in the Cross Lake greenstone belt (e.g., Corkery et al., 1992), northwestern Superior, and the Jackson Lake and Beaulieu Rapids formations in the Slave craton. Another well-known but long debated example that fits this pattern is the Moodies Group in the Barberton greenstone belt (e.g., Anhaeusser, 1984; Kamo and Davis, 1994), South Africa.

Structures resulting from the buoyant ascent include contractional structures in adjacent greenstone belts. Particularly, where buoyant domes develop laterally spreading mushroom heads, granitoids will come to overlie mafic rocks with moderate dips and, consequently, will create reflectors that are likely to be mistaken for accretionary structure in seismic profiles. Ductile reverse shear zones may develop on the flanks of the domes that will be ideal conduits for the large fluid pulse that must evolve from the sinking supracrustals. This may explain the attitude, kinematics, timing, and gold mineralization of the Campbell shear zones and related shears of the Con-Giant system in Yellowknife.

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Architecture of an Archean submarine fan, Burwash Formation, Slave Province, NWT.

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The Burwash Formation (ca 2661 Ma) represents an extensive sheet, 5-10 km thick, of metamorphosed sandstone and mudstone that lies above greenstone belts of the Slave Province in the Yellowknife area. The formation represents an unusually thick, well preserved unit of Archean turbidites, providing the opportunity to compare Archean deep-water sedimentary processes with Phanerozoic and modern analogues.

Field relations at the base of the Burwash Formation and in the underlying Raquette Lake Formation suggest an environment of active rifting, followed by a transgression during which condensed black argillites were deposited. Clastic sedimentation was punctuated in the overlying Burwash Formation by felsic tuff deposition. These tuff layers provide an opportunity for U/Pb zircon dating, thus placing constraints on sedimentation rates, and for lateral correlation within the sediment pile, allowing for architectural analysis of sedimentary facies. Zircon dates obtained from near the base and near the top of the formation are almost identical within error, indicating sedimentation rates that must have been of the order of 1 m per million years or higher, comparable to rates reported from modern submarine fan environments.

Sedimentary structures within the turbidites are closely comparable with those of Phanerozoic deposits; thinly bedded sandstones display partial to complete Bouma sequences. More thickly bedded intervals display internal scours superficially resembling hummocky and swaly cross stratification, but having a distinct unidirectional character. These structures are typical of high concentration sediment gravity flows.

Correlation of tuff layers and adjacent sediment packages has allowed the recognition of distinct architectural elements within the succession (i) channel-fill sandstones and rare conglomerates, typically with little interbedded slate; (ii) levee deposits, displaying significant lateral facies changes when traced hundreds of metres along strike; and (iii) sheet sands that onlap against levee deposits. The latter are interpreted as sands recording escape of turbidity currents from channels. A similar process in the Amazon and other modern fans has led to the deposition of high-amplitude reflection packages (HARPs) in interchannel locations.

The presence of abundant tuffs indicates that the Burwash Fans are not precisely equivalent to modern major passive-margin fans. However, a close analogy is suggested with fans developed in arc-related settings.

Regional-scale electrical structure of the Slave craton

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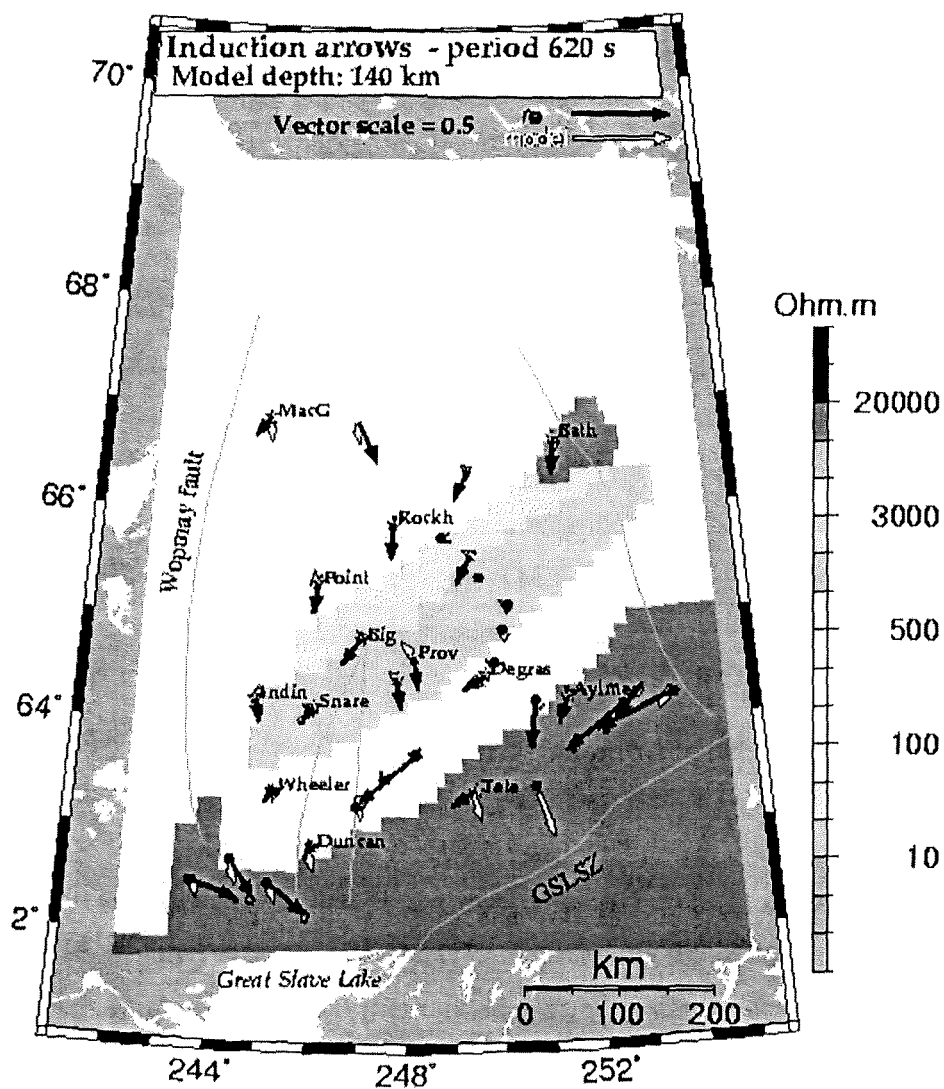
SUMMARY

The Slave craton in northwestern Canada is a relatively small craton (600 km x 400 km), but is ideal as a natural laboratory for investigating the formation and evolution of Mesoarchean and Neoarchean sub-continental lithospheric mantle. Excellent outcrop and the discovery of economic diamondiferous kimberlite pipes in the center of the craton during the early 1990s have led to an unparalleled amount of geoscientific information being available. Over the last five years four types of deep-probing electromagnetic surveys were conducted, using the natural-source magnetotelluric (MT) technique, as part of a variety of programs of study of the Slave craton to determine the regional-scale electrical structure. Two types of surveys involved novel acquisition; one through frozen lake ice along ice roads during winter, and the second deploying ocean-bottom instrumentation from a float plane onto the lake bottoms. MT measurements were made at over 150 locations across the craton, with the highest density along the winter roads and across the Yellowknife fault.

One principle goal for the MT surveys was to determine the geometry of the topography of the lithosphere-asthenosphere boundary (LAB) across the Slave craton. However, the MT responses revealed serendipitously a remarkable anomaly in electrical conductivity in the SCLM of the central Slave craton, named the Central Slave Mantle Conductor (CSMC). This anomaly is modeled as a spatially-confined region of extremely low resistivity (<30 $\Omega\cdot\text{m}$) beginning at depths of ~80-120 km and striking NE-SW. Where precisely located, it is spatially collocated with the Eocene-aged kimberlite field in the central part of the craton (the so-called "Corridor of Hope"), and also with a geochemically-defined ultra-depleted harzburgitic layer interpreted as oceanic or arc-related lithosphere emplaced during early tectonism. New three-dimensional (3-D) modelling of the CSMC shows that its top surface lies within the NE-SW striking G10

geochemical boundaries defined by Grütter, and that it dips to the NW (see below). This geometry is suggestive of lithospheric formation by imbrication of slabs from the SE (present-day coordinates), and is consistent with the G10-3 garnet trends described by Grütter.

The conductor may be interpreted in terms of either ionic conduction from hydrogen diffusion or electronic conduction due to the presence of carbon in graphite form, and neither explanation can be excluded based on existing data. The tectonic processes that emplaced this geophysical-geochemical body are likely related to the subduction of a craton of unknown provenance to the south during 2630-2620 Ma.



Three-dimensional electrical conductivity structure beneath the Slave Craton

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Nineteen long period magnetotelluric (MT) sites were collected on the floors of lakes throughout the Slave craton from 1998 to 2000 as an addition to Lithoprobe. After preliminary bounded influence data processing, high quality MT and magnetic Z/B responses have been obtained for periods ranging between 80 and 5000 s, thus allowing mantle penetration to depths as great as 400 km. The lake bottom data have been analyzed for the dimensionality of the deep conductivity structure, integrating in addition the data from other soundings collected along the winter road during 1996-2000.

A parameter that is a function of the longer period MT data indicates qualitatively the presence of anomalous conductivity zones. This has served to guide elucidation of the three-dimensional (3-D) structure in the deep crust and mantle. This 3-D induction strength parameter (see figure), reaches its largest values at Big and Providence Lakes and is still large to the SW at Snare Lake and to the NE at Contwyto Lake, suggesting that a conductor oriented NE-SW is traced at mantle depths beneath the center of the craton.

The data from the sites to the east, located near the Great Slave Lake shear zone and Bathurst fault (Healy in figure), are also affected by strong 3-D induction but at all periods. This suggests that the fault(s) have a high conductivity signature and reach substantial depths.

A 3-D conductivity model is in development, considering the qualitative information mentioned above, constructed by trial and error to fit the magnetic transfer and MT response functions. Interpretations of preliminary 2-D modeling of the MT data from the winter road state that the existence of a high conductivity zone encounters at depths of 80-120 km beneath the central Slave craton spatially coincides with the ultadepleted harzburgitic layer in the upper mantle, which may be related to an oceanic subduction slab emplaced during early tectonism. The 3-D model in preparation here traces this conductor as a NE-SW oriented mantle structure, with its centre located beneath Big and Providence lakes at a possible depth range of 50-250 km. The model includes also a more conducting layer in the northern Slave craton below the 40 km depth, subducting to the SSE and converging into the Slave mantle conductor at a depth of about 140 km. In the eastern craton, the model contains highly conductive crustal sheets beneath the Bathurst fault, steeping sub-vertically to the west and possibly converging into the Slave mantle conductor, suggesting the presence of conducting graphite films within a deep dipping fault zone.

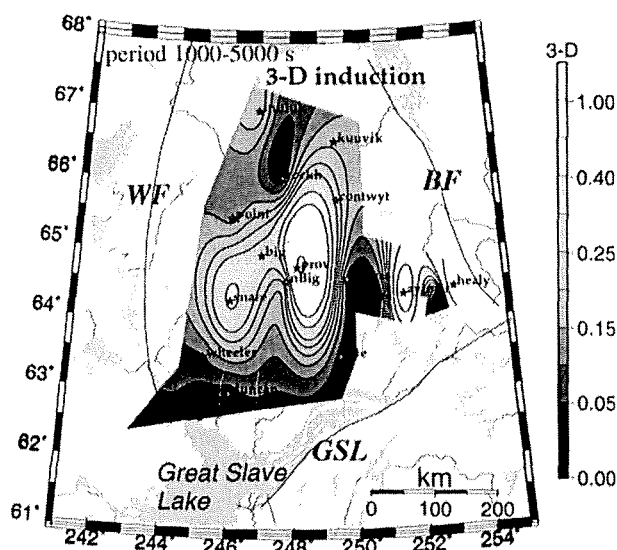


Fig. :
3-D induction strength parameter of magnetotelluric data from the Slave craton. Greater values (light grays) are indicative of anomalous conductivity zones.

WF : Wopmay fault zone.

BF : Bathurst fault zone.

GSL: Great Slave Lake shear zone.

Thermal Controls on Present Tectonics in the northern Canadian Cordillera: SNORCLE

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In the Canadian Cordillera, the average heat flow north of 59° North, 105 mW/m², is remarkably high. This is significantly higher than the average heat flow south of 59° North, 73 mW/m². Within the Cordillera, the higher heat flow is caused by higher heat generation north of 59° North, where the average heat generated in magmatic Cretaceous and younger rocks now exposed at the surface is 4.6 μ W/m³. The large variation in individual heat flow values can be explained by the observed geographical variation in upper crustal heat generation. However, since subduction stopped on this margin at 40 Ma, advective processes may be required to explain the extremely high heat flow and high, inferred lithospheric temperatures.

North of 59° North high heat flow extends eastward of the Cordillera into the foreland basin covering the Paleoproterozoic North American craton (Wopmay Orogen), where the average value is 92 mW/m². Most of the unexpected heat from this older lithosphere is thought to come from the upper crust, enriched in heat generation. This can produce deep lithospheric temperatures intermediate between those in the Cordillera and the low temperatures in the Slave craton. The moderate temperatures are confirmed by PN velocities and by isostatic models. The heat generated in samples from the Slave show the latest Archean magmatic rocks (7-15 μ W/m³) may be potential sources of the higher heat generation incorporated in the Wopmay orogen. The heat generated in 25 samples of well cuttings from the upper Wopmay basement averages 3.2 $\pm$ 1.8 μ W/m³. However, the drilling samples have preferentially lost the finest particles containing most of the enriched accessory minerals. An upper crustal heat generation of 6 μ W/m³ is used to calculate lithospheric temperatures beneath the Wopmay north of 59° North. Bottom hole temperatures indicate that the heat flow above the covered craton south of 59° North is more normal.

Lithospheric temperatures were calculated, using simple one-dimensional conductive models. Beneath the northern Cordillera, temperatures are so high (800-1000°C at the moho) that there is significant strength only in the upper crust. A mid-crustal detachment is inferred, allowing the obliquely converging Pacific plate to push the continental upper crust all the way to the Mackenzie Mountains, producing there the observed, enhanced seismicity. In the lower crust, high metamorphic grade rocks must be forming, truly the evolution of the lithosphere and a model of what has happened previously in the southern Canadian Cordillera. Some dipping structures identified along SNORCLE lines follow the trends of isotherms, and may be partially the result of differences in lithospheric temperature on the properties of the rock.

The difference in heat flow in the Cordillera occurring near 59° North produces a significantly warmer and weaker lithosphere to the north, similar to the lithosphere South of 51° North. With the Pacific plate moving obliquely toward the North American plate in a northeasterly direction, the offshore Queen Charlotte Transform Fault is probably caused by the significantly stronger and thicker continental lithosphere between 51 and 59° North. Underthrusting and local shortening must account for the component of convergence along this fault.

The modern Indus alluvial fan as an analog for the Paleoproterozoic Muskwa Assemblage

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Large sediment fans along continental margins at the mouths of major rivers such as the Indus, Ganges and Amazon are among the most impressively large geological features on Earth today. The new crustal seismic reflection profiles collected as part of SNORCLE reveal a wedge of layered rocks which may represent a similar apron of sediment formed between 1.85 and 0.54 Ga along more than 1000 km of the continental margin of ancient NW North America. This layered wedge was subsequently metamorphosed, tectonically thickened and covered by a series of exotic crustal and oceanic terranes (Gabrielse & Yorath, 1991). The terranes, thrust onto the wedge from the west between 200 and 50 Ma, thus largely hid its areal extent in the lower crust until probed by Lithoprobe.

The reflective, layered wedge observed on the SNORCLE reflection profiles is inferred to represent Proterozoic sediments primarily because of key correlations with outcrops near the eastern ends of lines 2 & 3. These correlations enable some of these reflective layers to be matched with well-studied Proterozoic sedimentary rocks such as the 0.8-0.54 Ga Windermere and 1.84-1.71 Ga Wernecke supergroups and the lesser known 1.8-0.8 Ga Muskwa assemblage (Cook & vander Velden 1993; Mitchelmore & Cook 1994; Cook & MacLean 1995; Ross et al. 2001; Thorkelson et al. 2001).

Stratigraphic studies of the exposed sedimentary sections document thicknesses of 7-21 kilometers, but correlations of seismic reflectors indicate a total sequence as thick as 30 km at the ancient continental slope or "hinge line" of Proterozoic North America (Clark & Cook 1992; Cook et al. 1999). It appears to taper to less than one kilometre at the toe

of the prism although metamorphism and a dynamic Moho may now mask its true thickness. Within the eastern half of the clastic wedge, we estimate that 1-5 km of Windermere Supergroup strata overlie a combined 5-25 km sequence of Muskwa and Wernecke strata and crystalline basement layers or tectonically inserted thrust slices. We are unable to resolve metamorphosed Proterozoic sedimentary rocks from gneisses or layered granitic intrusions in the mid- to lower crust with confidence using seismic-based methods alone.

Collectively, the SNORCLE reflective layers are here interpreted as a clastic apron built along the Proterozoic continental margin of North America. Analogy with the modern Indus or Ganges-Brahmaputra sediment budgets suggests that this prograding deep-sea fan probably represents about one to two thirds of the sediment load that originated in the interior of the continent (Goodbred & Kuehl 1999). Lateral equivalents represent the flood plain and deltaic depocenters of the same sedimentary units. In the modern systems flood plain deposits, prograding subaqueous delta deposits, and deep-sea turbidite deposits represent 30-40-30%, respectively, of the total sediment budget. In the Cordillera it thus becomes important to recognize not only the age of the sedimentary unit, but its depositional environment.

Currently, we hypothesize that much (70-80%?) of the layered reflective wedge is composed of Sequence A (Young et al. 1979) sedimentary rocks of the Muskwa Assemblage and Wernecke Supergroup (Thorkelsen et al. 1998; Ross et al. 2001; Thorkelson et al. 2001). Sequence B sedimentary rocks of the MacKenzie Mountains Supergroup are probably several kilometers thick on parts of the SNORCLE sections; the mix of carbonates, shales, evaporites and quartzites (Thorkelsen et al. 1998) suggests a shelf-deltaic depositional environment. Sequence C rocks of the Windemere Supergroup contain a wide range of sediment types. The units are generally thinner than those of the other sequences and variable in thickness across strike; syn-sedimentary block faulting suggest a protracted extensional or rift environment (Thorkelsen et al. 1998).

Dimensions of at least 1000 km along strike, 500-700 km in width and 30 km height indicate that the proposed wedge of Proterozoic sediments represents a minimum volume of $7.5 \times 10^6 \text{ km}^3$ of Proterozoic layered rocks. The wedge is thus similar in volume to the Indus fan (Clift et al. 2001), but the Cordilleran wedge represents a

compound feature deposited over a period up to ten times longer. If most of the wedge of layered reflections can be assumed to be composed of Muskwa Assemblage meta-sediments, then direct analogy with the modern Indus fan seems entirely appropriate. If so, careful study of the Muskwa Assemblage could provide important clues about the tectonics of the Proterozoic North America continent and also its climate and environment, much as the Indus has provided clues about the evolution of the Himalaya.

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SNORCLE CORRIDOR 3 MAGNETOTELLURIC EXPERIMENT

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Summary

The Canadian Cordillera comprises a region of oceanic and island-arc terranes accreted to western North America since the Neoproterozoic, with most of this westward growth occurring during the Mesozoic [Monger 1992]. These terranes have been since been displaced northward along large dextral strike-slip faults due to the motion between the Pacific ocean basin and the North American continent [Gabrielse and Yorath, 1991]. Morphogeologically and physiographically, these terranes can be divided into five major belts (Fig. 1) broadly representing fault-zone bounded geological units of regional extent. With the exception of the Foreland Belt, each belt comprises a collage of predominantly allochthonous terranes, and each terrane is characterized by its own geological and tectonic history different from its neighbors. Within the multidisciplinary study Slave-Northern Cordillera Lithospheric Evolution (SNORCLE) associated to LITHOPROPE (Canada's National Geoscience Project) [Clowes et al., 1992], the SNORCLE-Northern Cordillera MT experiment was carried during the summer-fall 1999. The SNORCLE-Northern Cordillera MT experiment (Fig. 1) included three regional-scale profiles crossing the main geological structures and a high resolution profile crossing the Tintina Fault along the Ketza River Mine road [Ledo et al., 2002].

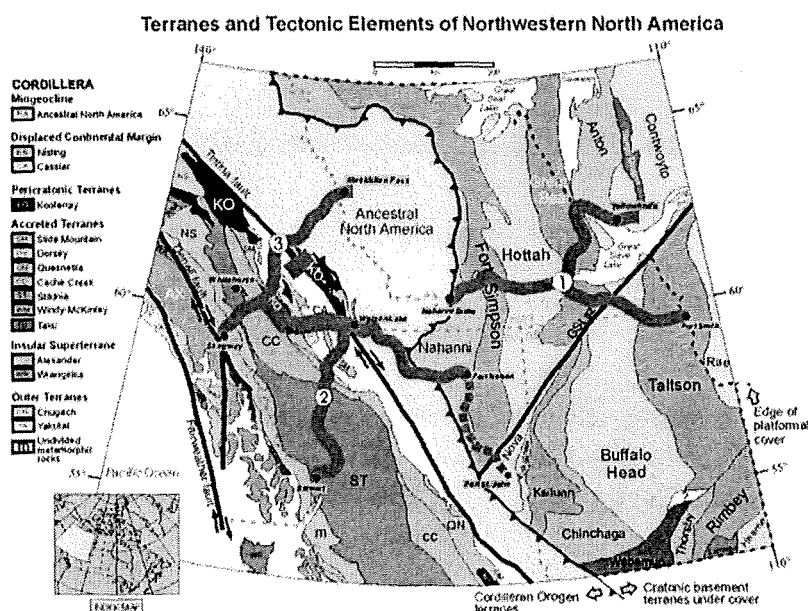


Figure 1. Terranes and tectonic elements of Northwestern North America and location of magnetotelluric profiles.

We focus on results for corridor 3, which is located mainly in the Yukon province and extends for 470 km from north of Skagway (Alaska), to the MacMillan Pass at the border between Yukon and the Northwestern Territories. The SNORCLE-Northern Cordillera corridor 3 MT experiment was planned to determine the electrical resistivity structure of the crust and the upper mantle across the accreted terranes and the ancestral North America beyond the Tintina Fault. From west to east, the profile starts in the Coastal Belt past through the contiguous Intermontane Belt and then crosses the Teslin fault that it is the limit between the Intermontane and Omineca Belts and after traversing the intracontinental strike-slip Tintina fault (TTF) samples the Ancestral North America rocks.

In total, forty-one MT soundings were acquired with a period range of 0.001 to 3000 seconds. In addition, long period data (LiMS) with a period range of 20-20,000 s were acquired at every other site. The maximum separation between sites was of 15 km and the minimum distance was of 5 km for the sites located close to the TTF. We used the Groom and Bailey MT tensor decomposition technique [Groom and Bailey, 1989] to identify and remove distortions caused by local near-surface features, and derive the regional two-dimensional (2-D) MT impedances in the most appropriate geoelectric strike directions. After that, a regional-scale 2D resistivity model is achieved through a minimum norm inversion scheme developed by Siripunvaraporn and Egbert [2000]. The resulting resistivity model is presented in figure 2.

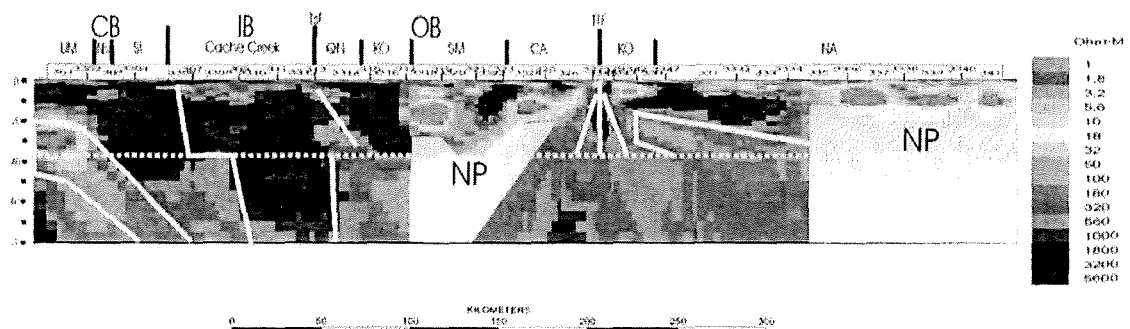


Figure 2. Magnetotelluric model of SNORCLE corridor 3, and its interpretation. NP: zones of no penetration.

2D model main structures and its geological interpretation

- Below the Coastal Belt (CB) at lower-crust and mantle depths (20-70 km) there is a northeast dipping conductive structure (3-100 Ohm.m) that could be related to plate subduction processes.
- The limit between the Coastal and Intermontane belts is imaged as a decrease of the resistivity. Within the Intermontane Belt, the upper crust is less resistive than the lower crust.

- The Intermontane Belt presents a lateral change at mantle depths that can be associated with the results obtained by Abraham et al. (2001) from the study of Tertiary to Recent alkaline lavas.
- The Teslin Fault is imaged as a northeast conductive dipping structure, this result agrees with the high-resolution study of the Teslin zone done by Snyder and Roberts (2001).
- The boundary between the Intermontane and Omineca belt is imaged as an important decrease of the conductivity of the Omineca belt. The presence of conductive structures below the Slide mountain terrane and the Cassiar terrane is consistent with the magnetotelluric model obtained for corridor 2 by Weenberg et al. (2002).
- The Tintina Fault is imaged at mid and lower crustal depths as a high resistivity feature. A more detailed study of the Tintina Fault at three different locations was done by Ledo et al. (2002).
- Within the ancestral North America the most significant feature is the presence of near surface high conductivity structures in the northeastern part of the model that may be associated with the presence of mineral deposits.

Acknowledgments

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Interrogating the Lithosphere of the Northern Canadian Cordilleran with Cretaceous to Recent Volcanics and Mantle Xenoliths.

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Since their assembly in the mid-Cretaceous, three episodes of volcanism have spanned of the disparate terranes of the northern Canadian Cordillera:

Recent Alkaline Volcanism (0-15 Ma)

Eocene Magmatism (40-50 Ma)

Late Cretaceous Carmacks Volcanics (70 Ma)

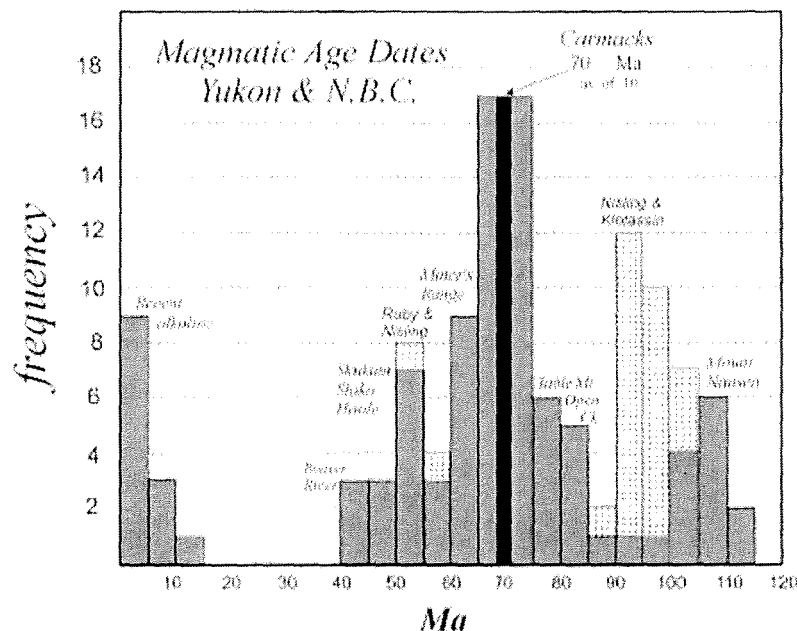


Figure 1: Frequency of magmatic age dates in the northern Canadian Cordillera.

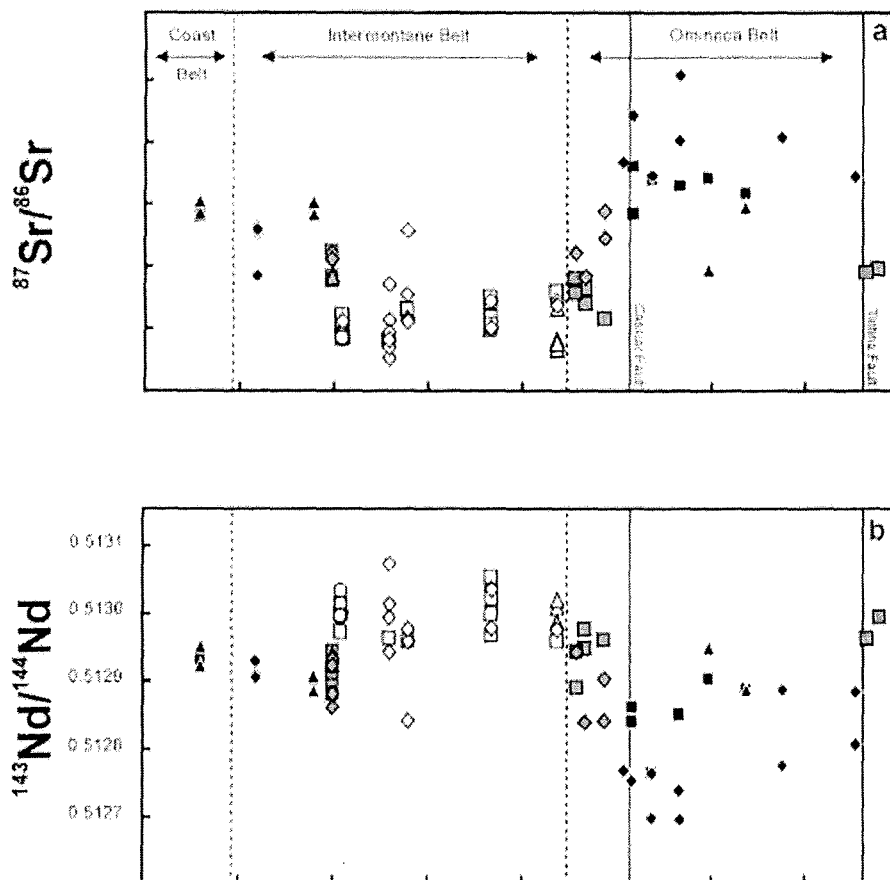
In each case, there are systematic chemical and isotopic changes in these volcanic suites across the Canadian Cordilleran that can be related to major terrane boundaries.

Recent Alkaline Volcanism (0-15 Ma)

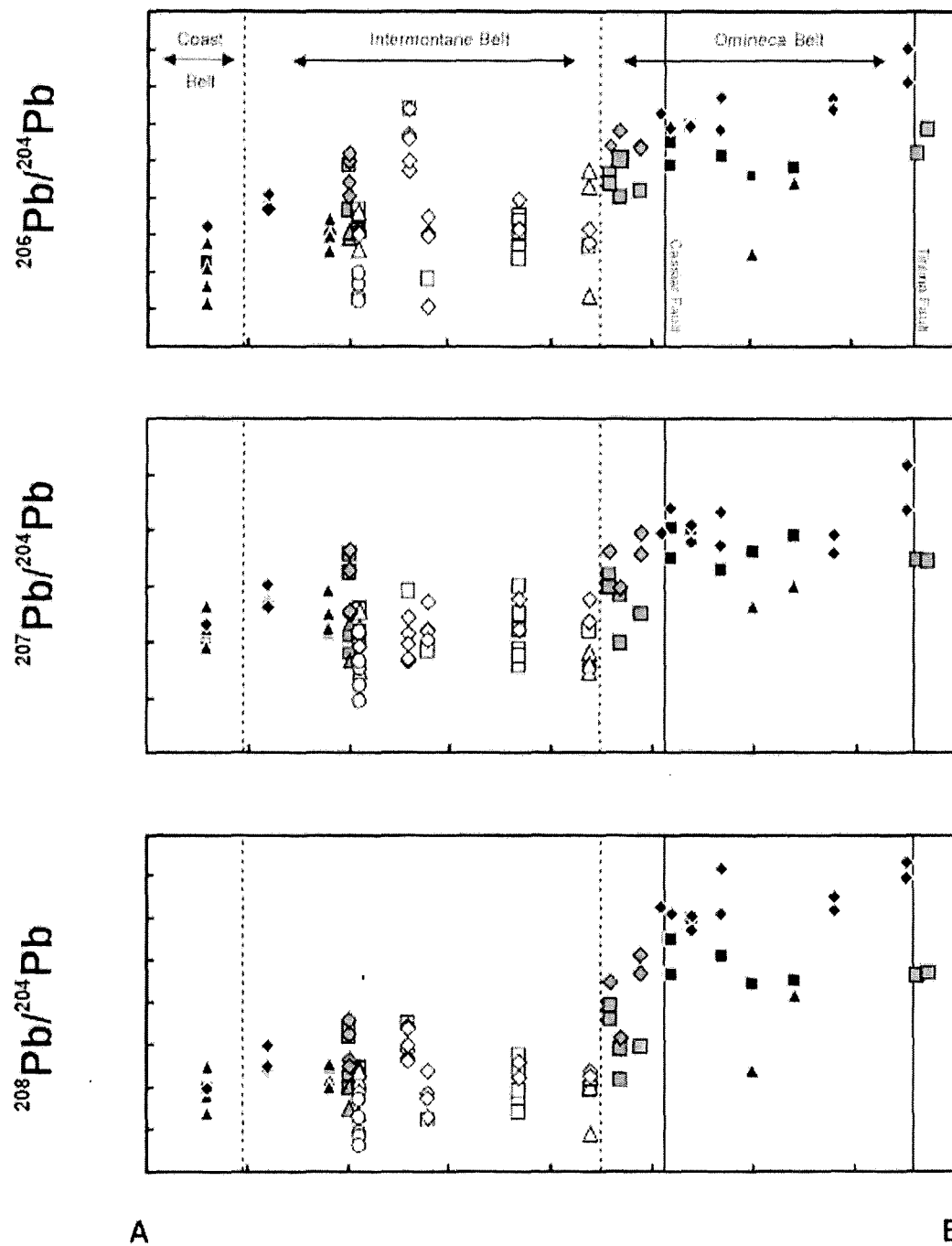
Numerous small alkaline volcanic centres of late Tertiary to Recent age occur along the length of the northern Canadian Cordillera that are thought to be associated with localised zones of trans-tension caused by the relative dextral motion between the Pacific and North American plates. Hypersthene-normative basalt is the volumetrically dominant lava type of recent alkaline volcanic centres in the Intermontane Belt (IMB), but lava compositions range through alkaline olivine basalt to basanite and olivine-nephelinite at a number of the smaller centres, in some cases in single volcanic vents. Suites of spinel lherzolite mantle xenoliths are relatively common in the smaller alkaline volcanic centers of the IMB, and have been found in primitive members of all 4 lava types. In contrast, more silica-understaurated alkaline olivine basalt is the dominant recent alkaline lava type in the Omineca Belt (OMB), and both olivine-nephelinite lavas and mantle xenoliths are absent.

There are three discontinuities in the isotopic and chemical signatures of the recent alkaline basalts across the northern Canadian Cordillera:

1. One occurs at the Tintina Fault, a major transcurrent fault separating the assembled terranes of the Cordillera from the North American Craton. Basaltic lavas of the east side of the fault are characterized by significant higher Ca, Zr, and $^{143}\text{Nd}/^{144}\text{Nd}$, and lower heavy rare-earth elements (HREE) and $^{87}\text{Sr}/^{86}\text{Sr}$ than their equivalents erupted on the west side of the fault. These results suggest that the Tintina Fault is a high angle fault that penetrates the upper mantle and separates 2 distinct lithospheric roots.
2. The second occurs between the Cassiar and Teslin Faults, approximately 20 km to the East of the surface expression of the boundary between the IMB and Omineca Belt (OMB), along the Stikine volcanic belt. Samples erupted to the East of the Cassiar Fault have higher Rb, $^{87}\text{Sr}/^{86}\text{Sr}$ and Pb isotopic ratios, but lower $^{143}\text{Nd}/^{144}\text{Nd}$ ratios than their equivalents in the westernmost OMB and the IMB.
3. A third occurs in the alkaline basalts of the IMB, to the east of its boundary with the Coast Plutonic Belt (CPB). The central IMB alkaline lavas have distinctly lower $^{87}\text{Sr}/^{86}\text{Sr}$, Nb/Zr, and Rb/Zr ratios than basalts erupted near its western margin or within the CPB to the West. The alkaline basalts of the western IMB and the CPB do not, however, share the high $^{207}\text{Pb}/^{204}\text{Pb}$ and $^{208}\text{Pb}/^{204}\text{Pb}$ ratios of the alkaline basalts of the OMB to the East. These results indicate that the relatively radiogenic source region of the CPB alkaline basalts extends beneath the western margin of the IMB.



In general, there is a striking overall correspondence between the regional variation in $^{87}\text{Sr}/^{86}\text{Sr}$ in the recent alkaline basalts and the Sr line defined by Armstrong on the basis of regional variations in the isotopic composition of Mesozoic and Tertiary granitoids. The Sr isotopic composition of the basalts, however, are systematically less radiogenic than adjacent Mesozoic granitoids. Modelling of the radiogenic alkaline lavas of the OMB, however, indicates that the differences between the recent alkaline basalts of the IMB and OMB cannot be simply the result of crustal contamination, but must lie deeper in the lithospheric mantle. This interpretation is supported by the evidence for the involvement of amphibole in the source region of these lavas, and the low Ca/Na ratios of the Cordilleran alkaline magmas in general, compared to oceanic equivalents, which indicates that they reflect relatively small degrees of partial melting of the lithospheric mantle. This interpretation, combined with the high heat flow in the northern Cordillera Cordillera, suggest that the lithospheric mantle beneath the Cordillera is undergoing active thinning. The absence of mantle xenoliths and olivine-nephelinite may indicate that this lithospheric thinning is more advanced under the OMB than the IMB.



Eocene Magmatism (40-50 Ma)

Eocene volcanic successions outcrop along the length of the western Intermontane Belt (IMB), adjacent to coeval granitoid plutons of the Coast Plutonic Belt (CPB) to the West, and isolated erosional remnants occur across the Cordillera, from west of Carmacks to the Tintina trench. The intensity of magmatism in the Eocene is believed to be related to the change in relative motion of the North American plate with respect to the Kula-Farallon plate, from dominantly orthogonal to oblique. Lava compositions range from basalt to rhyolite, and are characterized by chemical affinities that change from tholeiitic to calc-alkaline in both time and space across the northern Canadian Cordillera. Our chemical data indicates that the compositions of rocks from the Sloko - Skukum volcanics and coeval intrusives fall into two distinct groups. The first includes the Sloko Range and Sekulmun Lake volcanics, as well as the first cycle of the Bennett Lake volcanics (Fig. 1a & b). The second group includes all of the plutonic rocks, as well as the second cycle Bennett Lake volcanics, and the Skukum and Sifton Ranges. The first group is characterized by high Fe, Ti, Mn, P and low Si, with respect to the second group, and exhibits a fractionation trend that has transitional tholeiitic affinities. In contrast, the second volcanic group exhibits a more traditional calc-alkaline fractionation trend, with Fe, Ti, Mn, and P decreasing continuously with MgO. In comparison, Eocene lavas in terranes to the East of the CPB, in the IMB, OMB, and on ancestral North America tend to be basalts (Fig. 1b) with transitional tholeiitic affinity (Cinnamon Bun Strip, southwest of Dawson, and in the Ross-Hoole Rivers areas

There is a striking complimentary relationship between the two groups of compositional data. The population of the transitional tholeiitic lavas of the first group exhibits a continuous compositional range from basalt to dacite, but a pronounced population drop off at more Si-rich compositions (Fig. 1b). Field mapping suggests that there is probably a second mode at the composition of the more Si-rich rhyolites, but it is likely under represented because of the pyroclastic nature of many of the rhyolite units. The calc-alkaline lavas, however, are dominated by dacites and rhyolite, with relatively few basalts and andesites (Fig. 1b). They are compositionally indistinguishable from the plutonic rocks of the CPB which are dominated by biotite/hornblende granodiorites, but cut by younger granites. The vast majority of the compositions of both the calc-alkaline volcanics and the plutonic rocks fall in the population minimum that lies between rhyolite and the main mass of basalts and andesites of the transitional tholeiitic group.

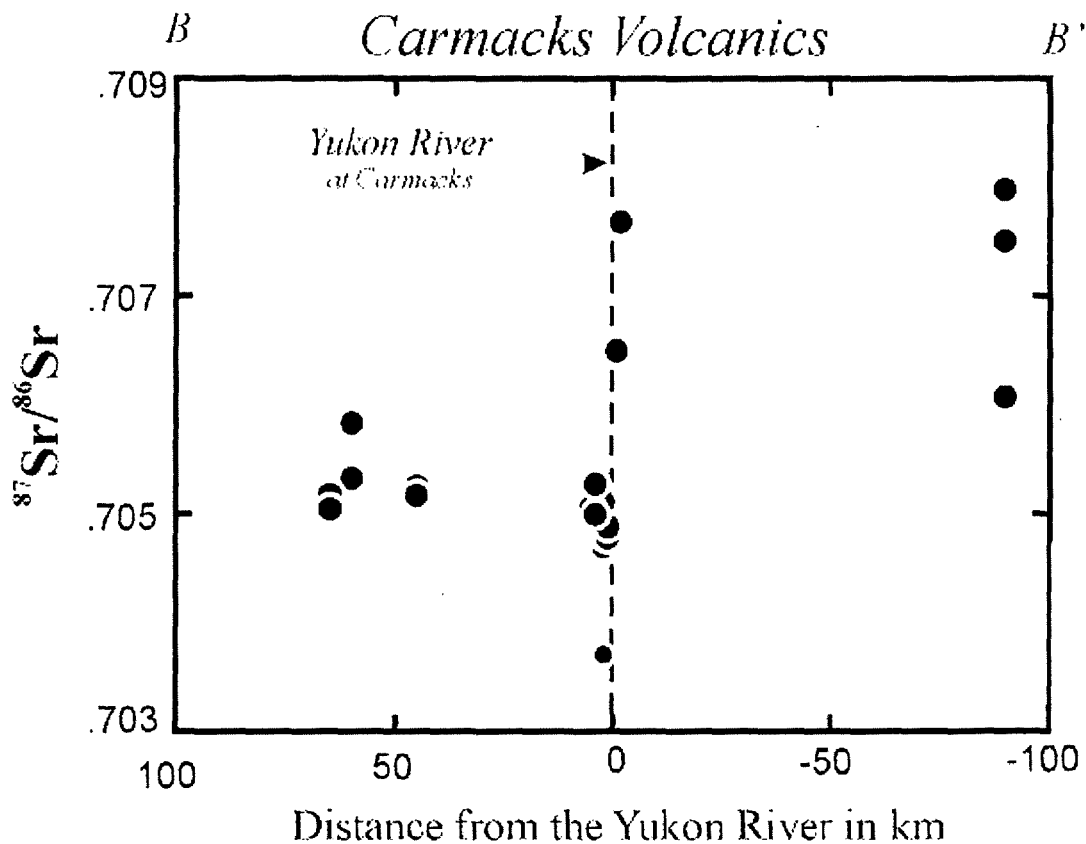
The evolution from early transitional tholeiitic lavas to late calc-alkaline lavas along the axis of the CPC during the Eocene, combined with the regional gradation to more primitive tholeiitic lavas with distance to the East of the CPB, suggests a model involving an evolving interaction between mantle-derived lavas and the continental crust in the CPB. According to one such model, early mantle-derived tholeiitic magmas caused only localized melting of the crust to produce the compositionally isolated rhyolitic volcanics. Along the CPB, however, the extent of interaction between mafic mantle-derived magmas and the crust increased with time, eventually producing the voluminous granodiorite plutons of the CPB and their calc-alkaline volcanic equivalents, which appear to represent hybrids of crustal and mantle-

derived magmas. Despite the clear evidence that calc-alkaline lavas succeed transitional tholeiitic lavas in the Bennett Lake Complex, however, the ubiquitous presence of both rhyolite and relatively primitive mafic dykes cutting all the volcanic and plutonic rocks, indicates that crustal and mantle-derived magmas have coexisted within this magmatic system throughout the Eocene. Unfortunately, there is not enough isotopic data at the moment to determine whether similar isotopic discontinuities corresponding to those observed in the recent alkaline and late Cretaceous Carmacks volcanics.

Late Cretaceous Carmacks Volcanics (70 Ma)

Remnants of thick successions of high-K to shoshonitic lavas stretch from Dawson to Whitehorse. These lavas have calc-alkaline chemical signatures, including: enrichment in large ion lithophile and light rare earth elements (REE), relative depletion in high field strength elements, and flat heavy REE patterns. In the central part of the Carmacks province, high-Mg ankaramites (15 wt.% MgO) have erupted on both sides of the Yukon River.

The lower Carmacks is dominated by localized accumulations of andesitic volcanoclastics that are overlain by more extensive flood basalts that range to primitive (15+ wt.% MgO) ankaramitic lavas containing olivine (Fo 93) megacrysts. All the lavas are relatively potassic, with calc-alkaline trace element signatures, including; enrichment in large ion lithophile and light rare earth elements (REE), relative depletion in high field strength elements, and flat heavy REE patterns. The Carmacks province is, however, chemically zoned on a regional scale, with shoshonitic lavas in the centre that are both more potassic, and more magnesian, than peripheral high-K lavas to the North and South. The presence of coeval shoshonitic ankaramites from Apex Mountain in the West to the easternmost Solitary Mountain occurrence implies that the primitive shoshonitic character of the central Carmacks magmas was insensitive to present Cordilleran terrane boundaries. The fact that the most magnesian lavas have the highest K/Na ratios and the lack of isotopic variation with degree of magmatic fractionation indicate the potassic nature of the Carmacks magmas is not a reflection of crustal contamination. High Si combined with low Al contents indicate that the primitive shoshonitic magmas in the centre of the Carmacks province formed at the shallowest depths and last equilibrated with refractory spinel-facies peridotite.



Isotopic data for the late Cretaceous Carmacks volcanics does, however, indicate the existence of an isotopic discontinuity across the boundary between the Intermontane and Omineca Belts along the Yukon River, near the town of Carmacks. From the West to the East sides of the Yukon River, there is a discontinuous increase in $^{87}\text{Sr}/^{86}\text{Sr}_i$ and decrease in $^{143}\text{Nd}/^{144}\text{Nd}_i$, which lines up with the northward projection of the Teslin fault to the South. Thus both the recent alkaline volcanics and the late Cretaceous Carmacks volcanics have detected a significant lithospheric change corresponding approximately to the IMB - OMB boundary that may correspond with the leading edge of ancestral North America.

The Great Alaskan Terrane Wreck

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Paleomagnetic studies place much of the Cordilleran orogen ≥ 3000 km south, relative to autochthonous North America at 85 Ma. This conflicts with geological studies which have been interpreted to indicate minor displacement. The inability to reconcile these interpretations has precluded understanding the tectonic development of the Cordillera, and has cast doubt on global paleogeographic reconstructions which rely upon paleomagnetic studies. I demonstrate that Alaska and Yukon are divisible into a series of geological belts, including the northerly east-west trending Arctic-Alaska, the central southwest-trending Ruby, the southerly east-trending Dillinger and the southeast-trending Yukon-Tanana belts. Each belt is characterized by four regularly arranged rock sequences; (1) a Paleozoic continental margin strata, (2) a Devonian-Mississippian arc assemblage, (3) an ophiolite, and (4) an Early to mid-Cretaceous arc. These correlations greatly simplify the terrane nomenclature of the northern Cordillera and provide a basis for a new palinspastic reconstruction of the Cordillera. Continuity of the belts can be demonstrated through connecting oroclinal arcs, and facing and structural vergence directions vary in a consistent and predictable fashion indicating that the belts previously formed part of a linear ribbon continent. This ribbon continent, referred to here as SAYBIA, originally extended south from eastern Siberia and was ~ 8000 km long. Northward translation in response to coupling with the Kula plate was accommodated by buckling (oroclinal orogeny) of SAYBIA, much like a derailing train, with scissored 'cars' forming much of Alaska. Oroclinal orogeny makes consistent interpretations of available geological and paleomagnetic data in the Cordillera, provides a means for the rapid construction of continent, and may have been an important process involved in the construction of ancient continental nuclei.

A Late Paleocene test of northwards terrane motions? Preliminary paleomagnetism of the 56 Ma Pattison Pluton, western Yukon

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1. Introduction

Paleomagnetism can be used to investigate past motions of Northern Cordilleran terranes on a regional scale. Such studies of the terranes underlying the Insular and Intermontane Belts (Fig. 1) have provided evidence for their systematic northwards translations and clockwise rotations relative to North America since mid-Cretaceous time (summarized by Irving and Wynne, 1990; Irving et al., 1996). Northwards translation of >1000 km has been confirmed for the northern Stikine terrane by several studies of mid-late Cretaceous units (Harris et al., 1997; Harris et al., 1999), although the ~2000 km displacement required by the 70 Ma Carmacks Group volcanics (Wynne et al., 1998) remains controversial.

Recent paleomagnetic studies under the auspices of the LITHOPROBE-SNorCLE transect have focused on the motion history of the pericratonic Yukon-Tanana terrane (YTT) and on the timing of the end of terrane motions relative to North America. Several paleomagnetic studies from units of the Coast Plutonic Complex and Intermontane Belt indicate that northwards motion of the accreted terranes had ceased by ~50 Ma (e.g. Symons et al., 2000a; Vandall and Palmer, 1990). Emplacement of the final, post-tectonic phases of the Ruby Range batholith and the Great Tonalite Sill at 58-50 Ma and their subsequent uplift also indicate consolidation of the accreted terranes in their present North American position by at least mid-Eocene time (Erdmer and Mortensen, 1993; Gehrels et al., 1991; Lowey, 2000).

Establishing the timing of the end of northwards motion is important for placing a minimum constraint on estimated terrane motion rates. Using 50 Ma as an 'arrival' time, northwards motion of ~10cm/yr is required to accommodate the displacement implied by the Carmacks Group result, involving ~100% coupling with the north tangential Kula plate motion with respect to North America from 70-50 Ma (Harris et al., 1999; Kelley, 1993). Here we report preliminary results for the Paleocene Pattison Pluton of the western YTT, in the interest of investigating the late northwards motion history of the accreted terranes.

2. Geology

The 225 km² Pattison Pluton (Fig. 2) is a roughly-circular undeformed alaskite alkali granite intrusion emplaced into the Paleozoic YTT Nasina quartzite and quartz-mica schist (Gordey and Makepeace, 2001; Lynch and Pride, 1984), some 260 km northwest of Whitehorse. It is part of the north-northwest-trending 57-54 Ma Nisling Range plutonic suite (Hart, 1997), and is associated with a series of north-trending dykes, perhaps collectively reflecting regional east-west crustal extension during Paleocene-Eocene time (Tempelman-Kluit, 1974).

The pluton features a roughly stratigraphic subdivision into coarse-, medium- and fine-grained phases, with gradational contacts between phases. The fine-grained phase is exposed at the highest elevations, and is interpreted to represent a rapidly chilled magma near the roof of the intrusion (Lynch and Pride, 1984). Mirolitic cavities

of centimetre size are present in the medium- and coarse-grained phases of the intrusion, further indicating a shallow crustal emplacement.

Late aplite sills cut all phases of the pluton and extend into the surrounding host rock. The sills are dominantly flat-lying, with slight dips to the northwest on average (Lynch, 1983). The sills, the roughly stratigraphic exposure of pluton phases, and the presence of generally flat-lying Carmacks Group flows to the northeast all indicate that the intrusion has likely not been significantly tilted since its emplacement.

A U-Pb zircon crystallization age of 56.3 ± 0.4 Ma has been determined for the Pattison Pluton (Hart and Mortensen, unpubl. data), placing it amongst the oldest of the Nisling Range intrusions. The pluton's size, shallow emplacement, and older age make it a useful subject for paleomagnetic study as part of the investigation of terminal Cordilleran terrane motions.

3. Paleomagnetism

The Pattison Pluton was sampled by gas-powered drill at 20 helicopter-accessible sites (Fig. 2), all from ridge exposures of the medium-grained alaskite phase of Lynch and Pride (1984). Care was taken to sample at the base of tors and away from local peaks, to lessen the chance of lightning strike-related remanence.

Sites were drilled in pairs, typically >50m apart. Each site consisted of six 2.54 cm diameter cores, oriented by sun and magnetic compass. The cores were later sliced into standard specimens of 2.2 cm length and stored in a room shielded from all but 0.2% of the Earth's magnetic field. Natural remanent magnetization (NRM) intensities measured for all 217 specimens had a median value of 2.2×10^0 A/m, and bulk volume magnetic susceptibilities had a median value of 3.2×10^{-3} SI.

The Pattison Pluton specimens underwent standard thermal or alternating field (AF) stepwise demagnetization in the shielded room at the University of Windsor Paleomagnetism Laboratory (for instrumentation and methods, see Symons et al., 2000b). Thermal demagnetization of two pilot specimens from each site revealed distributed unblocking with a sharp drop in the magnetite 500-580°C unblocking temperature range, indicating that the remanence carriers are likely a mixture of multidomain (MD) and pseudo-single domain (PSD) magnetite. Rock magnetic tests performed on 10 representative specimens confirmed the dominant presence of PSD-MD magnetite.

In most specimens, AF step demagnetization removed more than 80% of the NRM by ~20 mT, whereas thermal methods required ~450°C, consistent with the demagnetization behaviour of MD magnetite (Dunlop and Ozdemir, 1997). The above demagnetization ranges typically remove shallow-inclination remanence directions of various declinations from core to core. These scattered directions, carried by MD magnetite, are a viscous remanence that has no geological meaning. A steep characteristic remanent magnetization (ChRM) was recognized after removal of the viscous remanence in specimens from 13 sites. In the other 7 sites the shallow viscous remanence dominated even at high temperatures and coercivities, and no ChRM could be identified.

The steep ChRM is stable and present in both polarities. Nine sites carry an exclusively reversed, southeast-and-up ChRM whereas four sites carry both the reversed and a normal, northwest-and-down ChRM. The ChRM is carried by magnetite and subordinately by hematite, which occurs in the rock as staining associated with the miarolitic cavities. ChRM directions were obtained by least-squares line fitting (Kirschvink, 1980) where demagnetization trends could be clearly defined, and by sector-constrained remagnetization circles (McFadden and McElhinney, 1990).

1988) where some portion of the viscous remanence overlapped the ChRM.

Reversed- and mixed-polarity remanence directions are consistent with the behaviour of the Earth's field at around 56 Ma (Opdyke and Channell, 1996). The dual-polarity ChRM is most likely a primary remanence acquired as the pluton cooled through a span of time that encompassed at least one reversal of the Earth's magnetic field.

4. Discussion

Preliminary paleomagnetism of the Paleocene Pattison Pluton yields a paleopole of 87°N , 148°E ($dp=8^{\circ}$, $dm=8^{\circ}$), which is $6^{\circ}\pm 7^{\circ}$ far-sided and $24\pm 16^{\circ}$ clockwise-rotated in comparison with an interpolated 56 Ma reference pole from the North American apparent polar wander path of Besse and Courtillot (1991). Similar motion estimates arise from comparison of the Pattison Pluton pole with the 55 Ma reference pole of Torsvik et al. (2001).

Assuming the discordance between the pluton and reference poles to be entirely due to tectonic motions, the preliminary results from the Pattison Pluton imply that northwards translation had not quite ended by 56 Ma. This finding is consistent with the rapid motion rate required by results from the 70 Ma Carmacks Group volcanics (Wynne et al., 1998), but is also consistent with the lesser poleward motion rate required by the 75 Ma Mount Lorne stock of northern Stikine (Harris et al., 1999; Kelley, 1993). A polewards translation of 0° , consistent with results from 70 Ma YTT intrusions (McCausland et al., 2000; McCausland et al., 2001) lies within the 2-sigma confidence limits, albeit at lower probability.

Other paleomagnetic results of slightly younger age from plutons, dykes and volcanics of the Coast Plutonic Complex and the Intermontane Belt (e.g. Symons et al., 2000a; Vandall and Palmer, 1990) are on average $3^{\circ}\pm 5^{\circ}$ southwards translated, and $24^{\circ}\pm 10^{\circ}$ clockwise rotated, indicating that northwards translation had ended by 51-46 Ma, but also that some clockwise rotation had yet to occur. Pattison Pluton provides a rotation estimate that agrees with the Eocene studies. The ~ 1000 km geographic range of these clockwise-rotated units, from the Ootsa Lake volcanics (Vandall and Palmer, 1990) of the central Intermontane Belt to the Pattison Pluton of central-west YTT hints at a pervasive post-50 Ma rotation mechanism, such as distributed strain (Beck, 1998), or large-scale block rotations on sole faults (Symons et al., 2000a).

It is worth noting that the discordance between the observed Pattison Pluton and reference pole positions may be accounted for by post-emplacement NE-side up tilt of 6° . Such a tilt, or at least some lesser amount of it, cannot be ruled out.

If the aplite sills are assumed to be accurate recorders of paleohorizontal, a correction for their northwest tilt results in an unlikely tectonic motion estimate of $15^{\circ}\pm 7^{\circ}$ polewards translation and negligible clockwise rotation relative to North America. The sills are interpreted to have formed by late-stage filter pressing and lateral injection of the remnant melt, driven by increasing volatile pressure (Lynch, 1983; Lynch and Pride, 1984). Emplacement of such sills will have followed local pressure gradients in the high-level intrusion, and may not reflect paleohorizontal accurately enough to aid tectonic analysis.

Preliminary results from the Pattison Pluton marginally indicate that northwards terrane motion was still underway at 56 Ma, although a 'no translation' option cannot be ruled out. The consistency of Pattison Pluton's 24° clockwise rotation estimate with those of the Eocene studies noted above may indicate that no tilt correction of the Pattison Pluton result is required, and thus the $6^{\circ}\pm 7^{\circ}$ northwards translation is due to tectonic motion between 56 Ma and ~ 50 Ma.

5. Acknowledgements

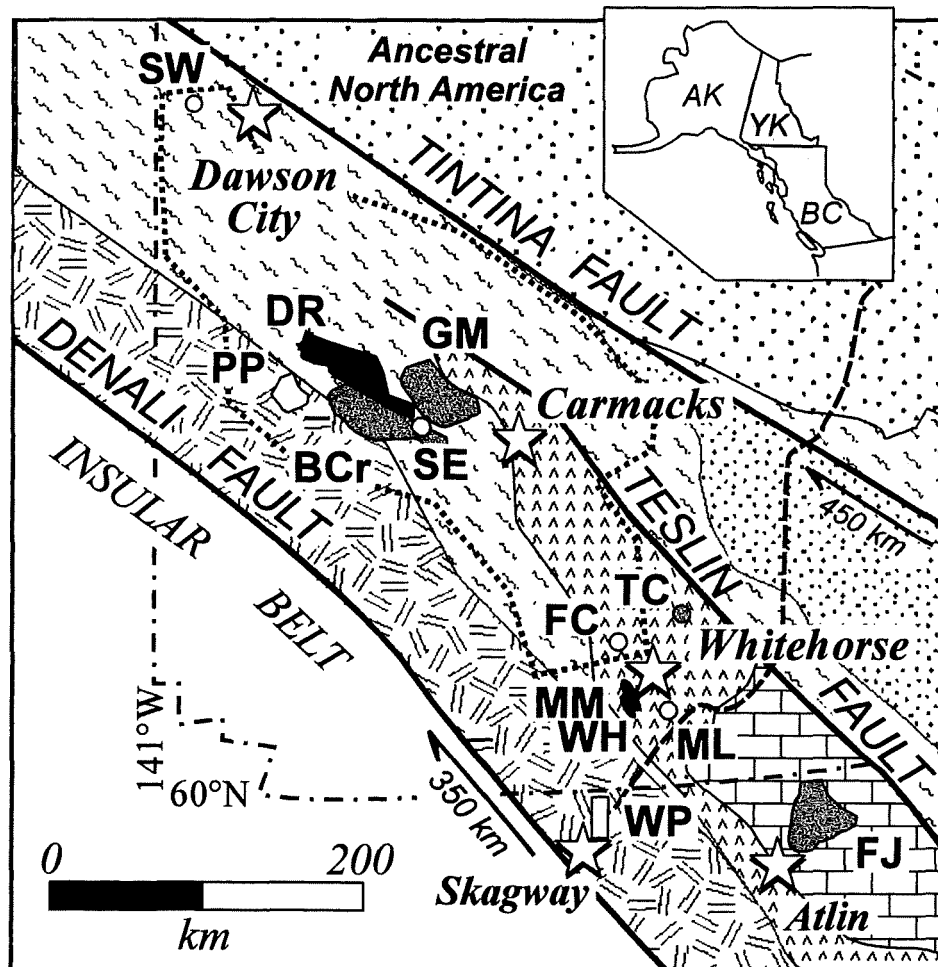
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Figure 1: Northern Cordillera, paleomagnetic studies



Intermontane Belt -YTT



Stikine



Cache Creek



Yukon-Tanana terrane

Post-accretionary/ overlap rocks



Coast Plutonic Complex



70 Ma Carmacks Group volcanics



**SNORCLE
Transect #3**



Fault zone

North American rocks



cratonic North America



Cassiar (transported)

WP-49 Ma White Pass dykes

FC-54 Ma Flat Creek pluton

PP-56 Ma Pattison pluton

SE-69 Ma Seymour Creek stock

SW-70 Ma Swede Dome stock

ML-75 Ma Mount Lorne stock

DR-100 Ma Dawson Range batholith

WH-110 Ma Whitehorse pluton

MM-112 Ma Mount McIntyre pluton

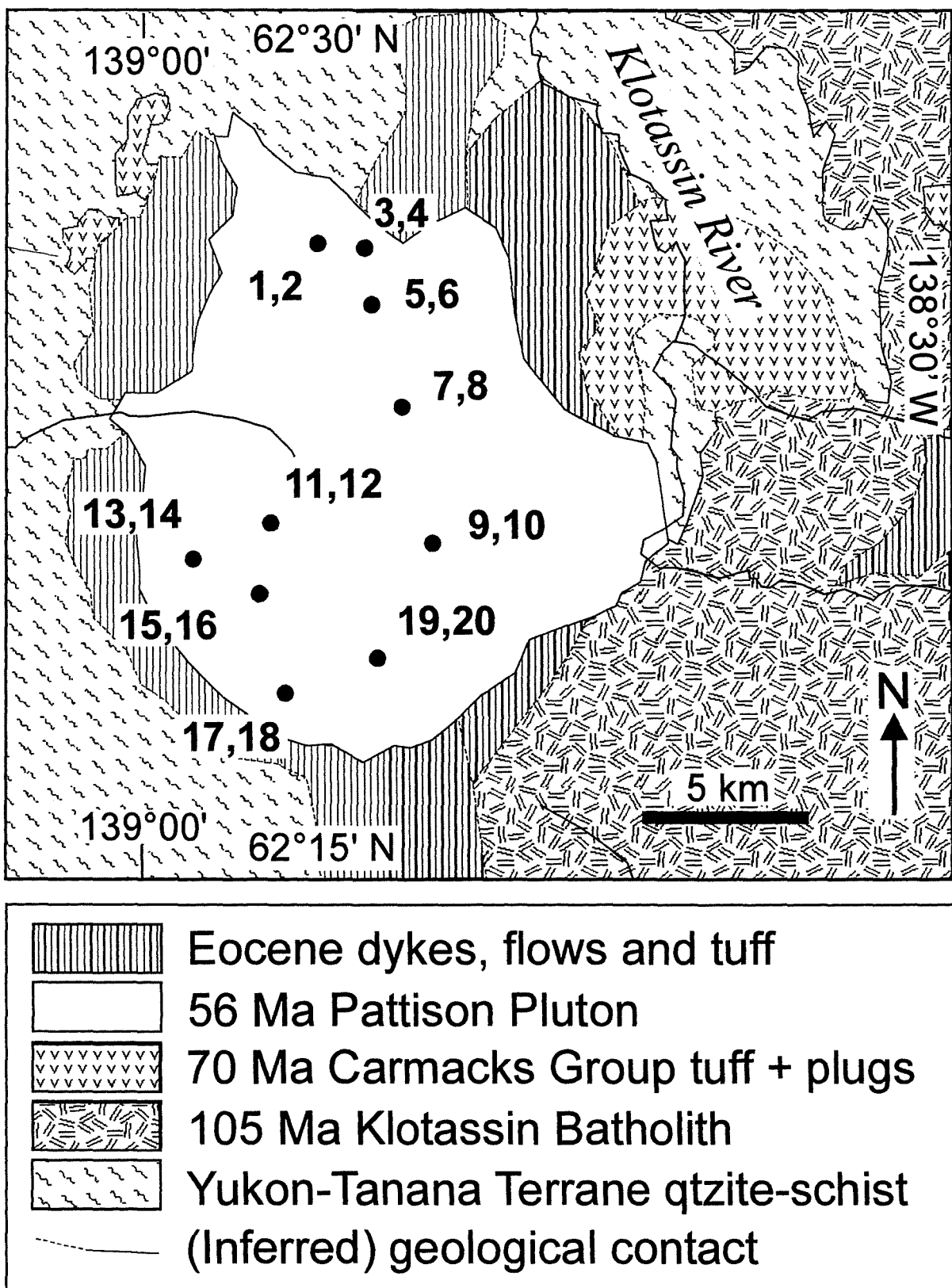
FJ-172 Ma Fourth of July batholith

TC-180 Ma Teslin Crossing stock

BCr-190 Ma Big Creek batholith

GM-190 Ma Granite Mtn batholith

Figure 2: Geology and sampling sites, Pattison Pluton



IS YUKON-TANANA TERRANE NORTH AMERICA?

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Speculation on the relationships of Yukon-Tanana Terrane with North America has ranged widely from original direct stratigraphic correlation, implying the terrane is relatively in place (Tempelman-Kluit, 1976) to the suggestion that it has been torn from the margin to the south and emplaced along dextral transcurrent faults (Mortensen and Jilson, 1985) to the idea that it is in part or whole, allochthonous (Tempelman-Kluit, 1979). Geophysical evidence from Alaska suggests that the terrane is thin-skinned (e.g. Beaudoin et al.). In the Finlayson Lake area of central Yukon (Fig. 1), elements of the ancestral margin and Yukon-Tanana Terrane are adjacent, therefore allowing a test of the above ideas through direct comparison of stratigraphy, structure and tectonic settings, as well as an examination of their mutual contact.

In this region, Late Precambrian to Triassic weakly metamorphosed sedimentary rocks that form the western part of the ancestral margin can be grouped into four distinct sequences: (1) Upper Proterozoic rift clastics comprising at least 3 km of turbiditic sandstone and shale (Hyland Group), (2) Lower Cambrian to Middle Devonian relatively deeper water limestone, shale and chert (Selwyn Basin) about 1600 m thick shoaling southwesterly to include shallow water quartz arenite and dolostone of Siluro-Devonian age (McEvoy Platform), (3) an Upper Devonian and Lower Mississippian rift succession comprising shale, turbiditic quartz-chert sandstone, and chert-pebble conglomerate (900?+ m) derived from elevated fault blocks of older basinal strata (Earn Group), and locally including thin accumulations of felsic volcanic rocks, and (4) a Lower Mississippian to Upper Triassic shallow shelf succession consisting of shale, chert, carbonate, minor sandstone, and siltstone (1700 m). Jura-Cretaceous regional deformation led to imbricate thrust faults of large overlap (50 km) and to local complex deformation above, extensive subhorizontal detachments. This deformation occurred prior to emplacement of widespread mid-Cretaceous granitic intrusions and their related thick, caldera-fill crystal tuff.

In contrast, the adjacent Yukon-Tanana Terrane consists of polydeformed meta-igneous and metasedimentary rocks (Murphy et al, 2002) including: (1) the Grass Lakes succession of Upper Devonian and Lower Mississippian psammatic schist, marble, mafic and felsic meta-volcanic rocks and mafic meta-igneous rocks, (2) the Wolverine succession of Lower Mississippian felsic meta-volcanic rocks, phyllite and exhalite, (3) the Tuchitua succession of Mississippian to Pennsylvanian intermediate volcanics, limestone and quartzite, (4) the Finlayson succession of Carboniferous(?) phyllite, chert,

clastic rocks, marble and intermediate and felsic volcanic rocks, (5) the Campbell Range succession of Permian basalt and chert, and (6) Middle Permian to Triassic conglomerate, siltstone and basalt. Lower parts of the terrane are intruded by voluminous Late Devonian to Early Mississippian granitic plutons. The terrane has been interpreted as a complexly evolving arc and back arc system, which underwent early Mississippian, mid- and latest Pennsylvanian deformation and shortening, before the Early Permian rifting that produced the Campbell Range succession (Murphy et al, 2002).

A comparison of the stratigraphy, ages of deformation and plutonism of Yukon-Tanana Terrane with that of the ancestral margin indicates nothing in common. The mid-Paleozoic plutonism and compressional deformation of Yukon-Tanana Terrane is not known from the adjacent margin, nor is extensive upper Paleozoic mafic volcanism. The extension-related Earn Group clastics of the margin derive from older basinal strata of that margin. There is no hint of any component of volcanic or arc-derived detritus within the Earn Group that may have originated from Yukon-Tanana Terrane. From these observations alone, it is clear that Yukon-Tanana Terrane is far-travelled relative to the North American strata to which it is now adjacent.

The structure separating Yukon-Tanana Terrane from the ancestral margin was identified in the Finlayson Lake area through fieldwork in 2001 carried out under the Targetted Geoscience Initiative (Murphy et al, 2002). Informally named the Inconnu thrust, it is a discrete, moderately southwest dipping, northwest trending, brittle thrust fault, in places juxtaposing upper Paleozoic Yukon-Tanana strata above Upper Triassic strata of the margin, demonstrating older over younger relationships. Along trend, the fault cuts up or down section in both hanging wall and footwall, so that in many places mid- to late Paleozoic Yukon-Tanana rocks may instead overlie older strata of the margin (i.e. younger over older). The age of emplacement of the terrane to its present position along this large thrust is post-Upper Triassic and pre-mid Cretaceous (i.e. the fault is locally cut by the mid-Cretaceous granitic suite), consistent with the age of deformation of ancestral margin strata of the footwall.

Although moderately dipping at its leading edge, geometrical relationships including the gentle dip of hanging wall rocks (i.e. Yukon-Tanana Terrane) indicates the Inconnu thrust to project southwesterly beneath Yukon-Tanana Terrane with relatively gentle dip. An implication is that the entire terrane northeast of Tintina Fault is detached and rests as a sheet above more distal strata of the ancestral margin, an overlap of about 60 km. This amount is in concert with the large overlaps on thrust faults and extensive detachments that have been demonstrated within footwall rocks of the ancestral margin. Accounting for dextral offset along Tintina Fault (Fig. 1), the Inconnu thrust would continue to project in the subsurface beneath the Stewart River area southwest of the Tintina. The persistence of gentle dips of hanging wall rocks (i.e. Yukon-Tanana Terrane) to the southern margin of this area, suggests that strata of the ancestral margin could project underneath the entire Stewart River region, and perhaps farther southwest to the Denali Fault.

Regionally the Inconnu thrust and equivalent segments (see below) also carries slices of mafic-ultramafic rocks and ribbon chert of the oceanic Slide Mountain Terrane, Permian eclogitic rocks, and proximal, immature conglomeratic rocks (accretionary wedge clastics) derived from them. These components were extensively imbricated with Yukon-Tanana Terrane in a forearc setting along its eastern (relative to present day) edge in Permo-Triassic time. The Slide Mountain Terrane should not be thought of as separate from Yukon-Tanana. Rather the two were already largely(?) assembled as a structural package and emplaced as a unit in their final position above North America in the Mesozoic.

In this context the Inconnu thrust in the Finlayson Lake area, is a segment of the largest, most regionally extensive thrust fault in the northern Cordillera. As indicated, it bounds the entire Yukon-Tanana and Slide Mountain terranes northeast of Tintina Fault (Fig 1, no. 1). Equivalents to the south include the sole of the Sylvester Allochthon in northern British Columbia (Fig 1, no. 2), the eastern limit of Yukon-Tanana rocks in south-central Yukon where they rest next to Cassiar Terrane (Fig. 1, no. 3) and the base of the St. Cyr (Fig 1, no. 4) and McNeil klippen (Fig. 1, no 5; Gordey, 1981) in the Pelly Mountains. Eliminating dextral duplication along Tintina Fault, the Inconnu thrust and its equivalents can be traced along strike for some 900 km.

The stratigraphic, tectonic and structural considerations described above clearly indicate the Yukon-Tanana Terrane is a thin-skinned terrane and far travelled relative to the North American strata with which it is now juxtaposed and overlies.

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Extension of Dorsey assemblage into the southern Sylvester allochthon: Farthest south Yukon-Tanana Terrane

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The Sylvester Allochthon comprises a thrust-imbricated set of assemblages of oceanic to pericratonic arc affinity, lying in structural juxtaposition above para-autochthonous strata of the Cassiar Terrane in far northern B.C. (Figure 1). From structural base to top they are: late Paleozoic marginal oceanic rocks of the Slide Mountain terrane; late Paleozoic arc and carbonate strata similar to the Harper Ranch subterrane of Quesnellia (Nelson, 1993); and a deformed and metamorphosed pericratonic unit in the southern part of the allochthon (Harms *et al.* 1993; Gabrielse, 1998; Harms, 2000). This pericratonic unit resembles the Dorsey assemblage, which forms part of the allochthonous suite (formerly Dorsey Terrane) west of the Cassiar Batholith (Harms and Stevens, 1996). Both the Dorsey assemblage and the pericratonic unit in the Sylvester Allochthon are intruded by Late Devonian to early Mississippian plutons (Gabrielse *et al.*, 1993; Nelson, 1999), a feature that also allies them to the vast Yukon-Tanana Terrane farther north. It has been suggested that the pericratonic assemblage in the Sylvester Allochthon is a direct southeastward extension of the Dorsey assemblage, interrupted by the Cassiar batholith (Harms, 2000; Nelson, 2001). This 2001 field project was intended to test correlations between the pericratonic assemblages east and west of the Cassiar Batholith via detailed mapping within the Sylvester allochthon in the Beale Lake area east of the batholith (Figure 1).

In the Beale Lake area, the Sylvester Allochthon can be divided into three distinct assemblages, a lower one of oceanic affinity, a middle one of island arc affinity, and an upper deformed and metamorphosed pericratonic unit (Figure 2). The structurally lowest slivers, truncated to the southwest by the Cassiar Batholith, are ultramafite, gabbro and basalt; they probably form a southern strike extension of Slide Mountain Terrane exposures along the Stewart-Cassiar Highway near Cassiar. They also correspond to Package I of Harms (1986). Above them, across a highly sheared contact, lies the Four Mile assemblage (informal name), a set of thrust slices of more diverse character, including widespread felsic to intermediate volcanic rocks of probable arc affinity. Interbedded sedimentary rocks have yielded Pennsylvanian-Permian and Permian conodonts (Gabrielse, 1998; M. Orchard, personal communication, 2002). The stratified rocks are intruded by the mid-Permian Meek and Nizi plutons. The Four Mile assemblage shows a similar tectonic environment to Harper Ranch subterrane panels exposed farther north, for instance the Huntergroup volcanics and an unnamed Permian pluton in the headwaters of Big Creek near the Yukon border (Nelson and Bradford, 1993).

The Beale Mountain Thrust, exposed for many kilometres on the western face of Beale Mountain above the Four Mile River valley, spectacularly marks the base of the deformed and metamorphosed pericratonic allochthon (Figure 2). It juxtaposes amphibolite and metasedimentary rocks, which were ductilely deformed in early Mississippian time, above a virtually undeformed Pennsylvanian-Permian volcanic/plutonic footwall. Above the thrust, the deformed rocks can be divided into two units. The lower unit consists of garnet grade basinal sedimentary rocks - metachert, meta-argillite, pelite and quartzite - with large lenses of amphibolite and metagabbro. The well-foliated but less tectonized upper unit contains phyllite, quartzite, metachert and most significantly, quartz-feldspar porphyry rhyolite and dacite flows and tuffs. These two units correspond closely to the lower and upper Dorsey assemblage in the southern Yukon and northern British Columbia west of the Cassiar Batholith (Stevens and

Harms, 1995; Roots and Heaman, 2001; Nelson, 1999, 2000). The lower unit, with its extensive cliff-forming amphibolites, is recognizable in localities from the Swift River valley in southern Yukon (Stevens and Harms, 1985) to the west face of Beale Mountain. The upper unit, with its felsic volcanic component, is probably equivalent to not only the upper Mississippian felsic units in the Dorsey assemblage in southern Yukon, but also to the Yukon-Tanana Terrane in the Finlayson Lake belt.

The lower Dorsey assemblage in the Beale Lake area is an interlayered sequence of siliceous metasedimentary strata, amphibolite and metagabbro. The metasedimentary rocks, now fine grained quartzites, quartz-biotite-muscovite(-garnet-plagioclase) schists, marble and calc-silicates, probably had chert, argillite impure limestone, and minor orthoquartzite protoliths. True pelites are rare; most micaceous rocks are dirty quartzites. Kyanite occurs in two samples from the lower Dorsey assemblage in the Sylvester Allochthon. In both cases it is relict, partly resorbed, and overgrown by muscovite. In one of these, from the west side of Beale Mountain, later fibrolitic sillimanite grows in the predominant fabric. This paragenetic sequence of kyanite to sillimanite mirrors aluminosilicate growth recorded in the lower Dorsey assemblage west of the Cassiar Batholith (Nelson, 2000). Harms *et al.* (1993) report amphibolite-grade assemblages of garnet-muscovite-zoisite and biotite-garnet-staurolite from pelites elsewhere in the deformed pericratonic assemblage.

Amphibolite occurs on all scales in the lower Dorsey assemblage, from fine layers alternating with metachert and siliceous pelite to mountain-scale, lensoid bodies. Mineralogy varies from simple hornblende-plagioclase-quartz-titanite to garnet- and garnet-diopside-calcite-bearing varieties. The latter have a calcareous component in excess of that for normal basaltic rocks. It could be due to admixture with carbonate sediments, or seafloor alteration. Metamorphic fabrics in the amphibolites range from extremely foliated to nearly isotropic. The most foliated varieties consist of ribbons of very well aligned hornblende prisms, alternating with more felsic quartz-plagioclase-bearing laminae. Fine titanite trains parallel the layering. More equant and randomly oriented hornblende overgrows these textures, suggesting that peak metamorphism succeeded peak deformation.

The upper unit of the Dorsey assemblage contrasts strongly with the lower unit. On both sides of Beale Lake, amphibolites abruptly disappear upsection, and well-foliated, quartz-feldspar phyric felsic metavolcanic rocks become important. They are accompanied by dark grey, grey, tan and green phyllite, quartzite with remnant quartz grains, and metachert. The identification of abundant felsic metavolcanic and epiclastic material in the upper part of the pericratonic unit is an important step in regional geological and metallogenetic correlations, since such rocks characterize both the metal-rich Finlayson Lake belt and, more locally, the early Mississippian upper Dorsey assemblage near Swift River (Roots and Heaman, 2001). U-Pb analyses are pending.

Previous work (Gabrielse, 1998) suggested that the Permian Nizi pluton was emplaced into the deformed pericratonic unit, constituting a common element with coeval plutons in the underlying panels (Gabrielse *et al.*, 1993). This study has documented a major fault, the Beale Mountain Thrust, between them. The existence of a Permian intrusive suite within the pericratonic assemblage, similar to the Ram Stock in the lower Dorsey Terrane near Swift River, Yukon (Stevens and Harms, 1995) and small pegmatites farther south in British Columbia (Nelson, 1999), has thus lost a major support. There is now only one known Permian intrusion within it, a small body between the Dease and Four Mile rivers in southern McDame map area (sample 89-SY-72A; Gabrielse *et al.*, 1993). However, a large, heterogeneous, unfoliated, diorite to tonalite

body cuts the lower Dorsey assemblage northeast of Nizi Creek; part of it is exposed in the northeastern corner of Figure 2. In composition and texture it strongly resembles the nearby Nizi pluton, and could be its displaced roof. It has been sampled for U/Pb dating. In addition, sparse, greenish-black amphibolite rafts in the Nizi pluton indicate that it penetrated a basement not unlike the amphibolite-bearing lower Dorsey assemblage.

The Dorsey assemblage in the Beale Lake area is characterized by at least two episodes of penetrative deformation and shearing. One episode occurred in the early Mississippian, as shown by late synkinematic plutons of that age (Gabrielse *et al.*, 1993). A second, later, ductile-brittle shearing event accompanied its emplacement on top of the Four Mile assemblage as the hanging wall of the Beale Mountain Thrust. This event must have postdated the youngest footwall rocks, the mid-Permian Meek and Nizi plutons. Distinct sets of fabrics and strain gradients can be related to these two events.

Ductile strain diminishes gradually upwards within the Dorsey assemblage. The lower Dorsey assemblage contains sheath folds and very strong quartz stretching (E1) and mineral streaking (M1) lineations, associated with shear bands and c-s fabrics in appropriate lithologies. These features decrease to uncommon occurrence in the upper Dorsey assemblage. Rootlessly folded small leucosomes, a common feature of lower Dorsey amphibolites, do not appear in the upper Dorsey assemblage.

The E1 stretching and M1 streaking lineations show a pronounced northeasterly maximum at 58/18, and a lesser southwesterly one (Figure 3a). These represent measurements on the two limbs of an upright, F3 syncline, which also folded transposed layering and schistosity (S1) in the Dorsey assemblage (Figure 3b), and gives rise to the tongue-shaped outcrop pattern of the Dorsey assemblage in Figure 2. Shear bands and asymmetric porphyroclasts both in outcrop and thin section indicate tops to the northeast. This early tectonic event is assigned to the early Mississippian, based on radiometric data on cross-cutting plutons (Gabrielse *et al.*, 1993). The top-to-the-northeast shear sense and well-developed E1 stretching lineations are identical to those observed in the Dorsey assemblage west of the Cassiar Batholith (Nelson, 1999).

Close-spaced shear zones are developed in rocks tens of metres from the post-mid-Permian Beale Mountain Thrust, including the lower Dorsey assemblage and the otherwise undeformed Four Mile assemblage. Shear bands indicate top-to-the-northeast motion on these subsidiary shears, and thus by implication on the main thrust fault. Interestingly, the emplacement of the Dorsey assemblage on top of the subadjacent late Paleozoic arc assemblage followed the same sense as its much earlier internal deformation. Either the same large-scale crustal architecture controlled both, or crustal geometries shaped during the early Mississippian event influenced later accretion.

This study makes explicit the correlation between the pericratonic rocks in the Sylvester allochthon and those west of the Cassiar Batholith. It is suggested that the name "Dorsey assemblage", originally proposed by Harms and Stevens (1996) for metamorphic rocks in the Swift River drainage, southern Yukon, be also used within the Sylvester Allochthon. Grounds for this correlation include the twofold lithologic division, presence of Devonian-Mississippian plutons, metamorphic grade, and style of deformation.

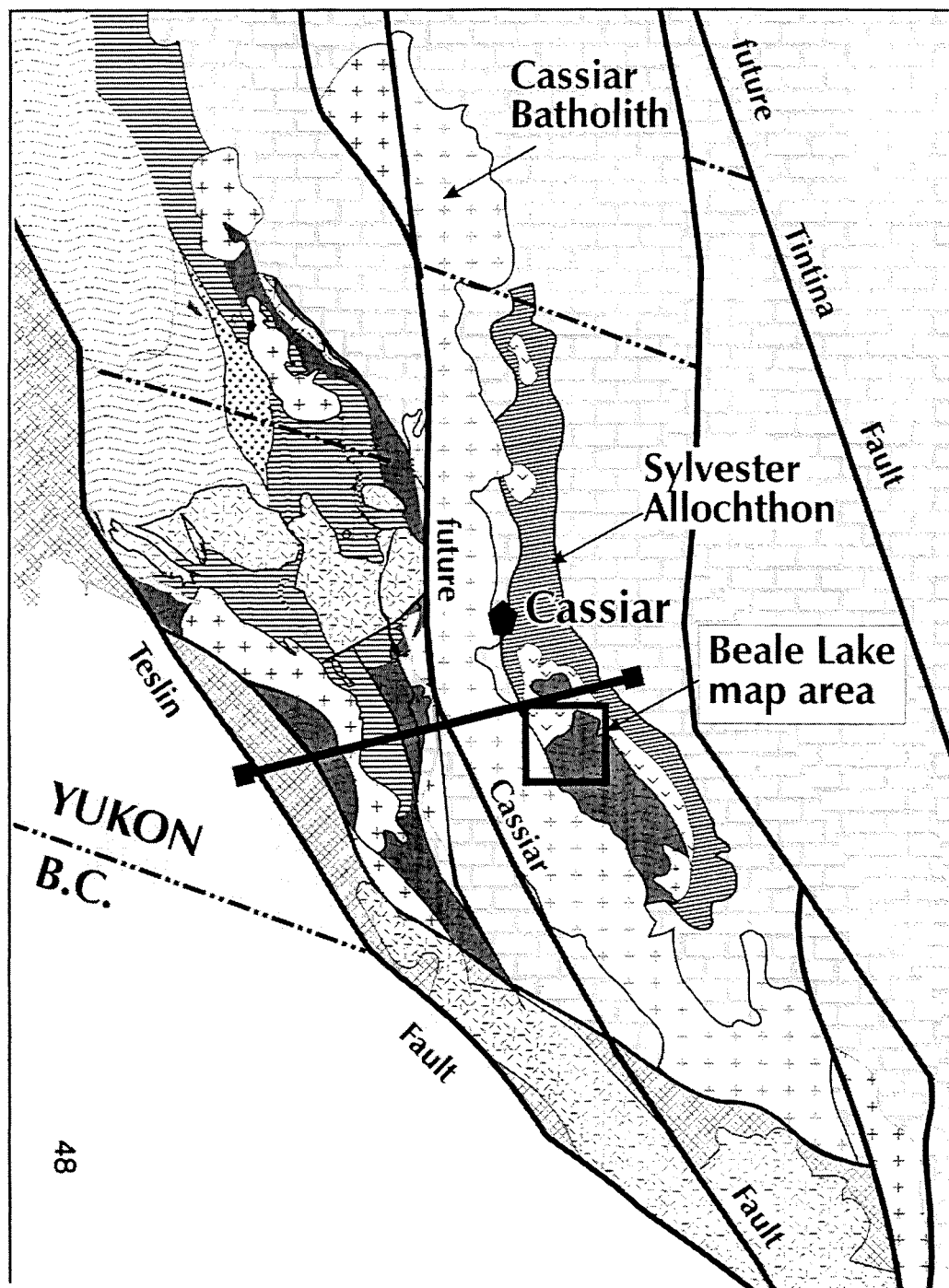
Continuity of the Dorsey assemblage across the Cassiar Batholith allows construction of a combined cross section from the Teslin fault eastward through the Cassiar Mountains. Restoration of 100 km of mid-Cretaceous dextral transcurrent motion on the Cassiar Fault provides a useful base map on which original accretionary relationships can be seen (Figure 1). In this section (Figure 4), the allochthonous units are exposed in two flat-bottomed synclinoria, with the region of the Cassiar Batholith an anticlinorium between. The combined stacking order is, from lowest and most inboard, to highest and most outboard: 1) Slide Mountain Terrane; 2) Paleozoic arc volcanic and carbonate strata (Ram Creek assemblage west of the batholith; late Mississippian-Permian Harper Ranch subterranean units in the Sylvester Allochthon) 3) Dorsey assemblage; and 4) late Paleozoic basinal, arc and carbonate strata that overlie the Dorsey assemblage west of the batholith. The Slide Mountain rocks disappear westward, cut out by the basal fault of the combined allochthons. Within unit (2) it is noteworthy that the oldest rocks are exposed in the Ram Creek assemblage to the west; this is consistent with fault ramping upsection towards the foreland.

The gentle structural geometry in this cross section is consistent with relationships observed elsewhere northeast of the Teslin and Big Salmon faults, for instance the gentle geometries of the McNeil and Ct. Cyr klippen traced out by Tempelman-Kluit (1979), and the interpretation of deKeijzer *et al.* (1999) that Cassiar Terrane strata reappear west of the allochthonous rocks in the Teslin zone near the Big Salmon River. It is also consistent with the gentle structural attitudes portrayed in the SNORCLE transcrustal sections (Cook *et al.*, 2001): a pleasing instance of harmony between geological and geophysical data sets!

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Legend

Post-accretionary units

- Eocene Major Hart pluton
- Cretaceous granite
- Jurassic intrusions

Pericratonic terranes (Yukon-Tanana equivalents)

- Triassic "Logjam clastics"
- Klinkit succession, Screw Cr. limestone,, Swift River succession
- Harper Ranch subterrane
- Dorsey assemblage
- Ram Stock (mid-Permian)
- Big Salmon Complex
- "Kootenay Terrane" in Teslin map area
- Ram Creek assemblage

Other terranes

- Slide Mountain Terrane
- Cache Creek Terrane
- Quesnellia: Triassic-Jurassic Shonoktau and Nazcha Formations
- Stikinia
- Ancestral North America (Cassiar Terrane)

Figure 1. Regional setting of southern Yukon-Tanana Terrane and Dorsey assemblage. Base reconstructed prior to mid-Cretaceous collision between Canada and South America.

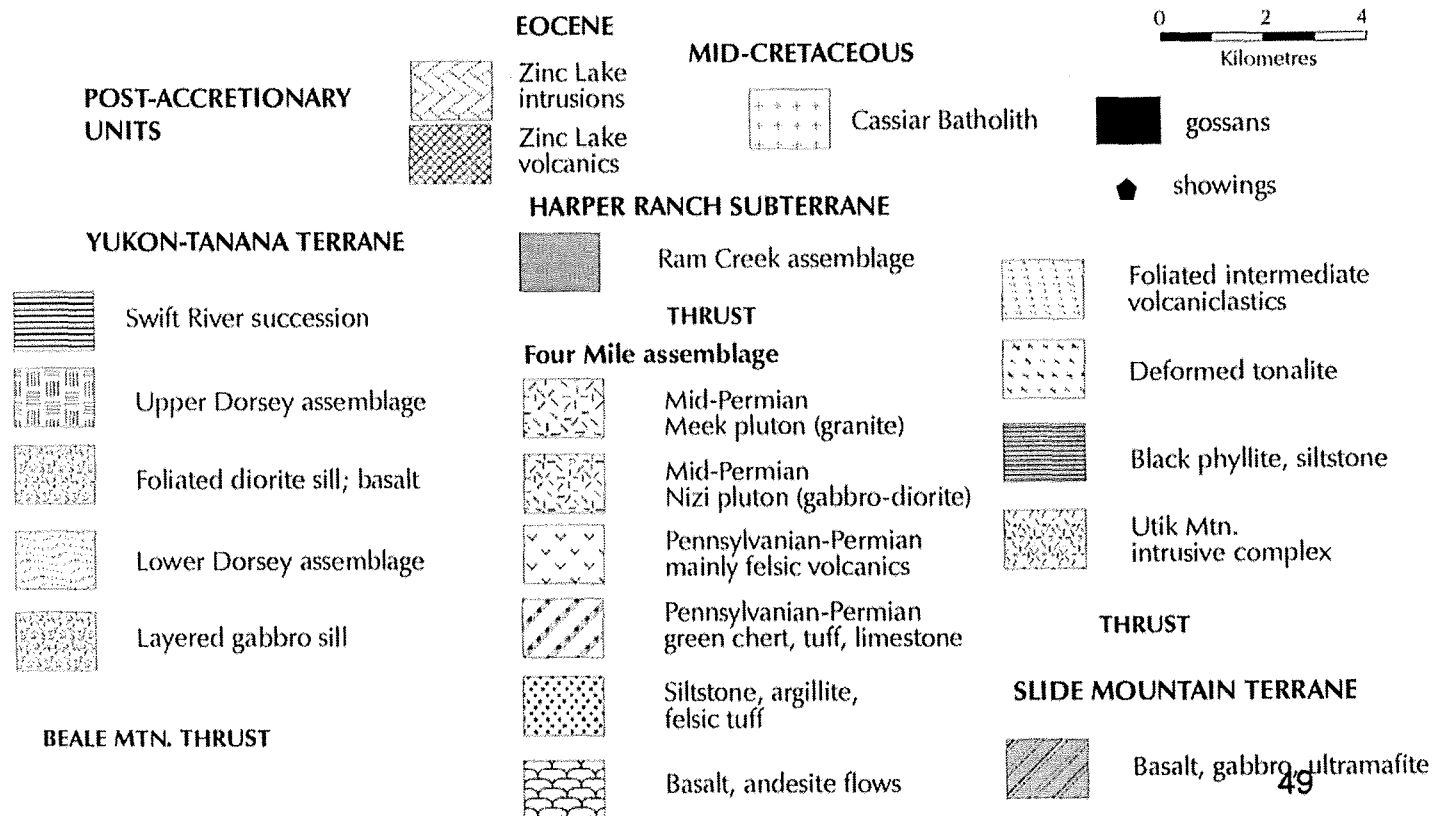
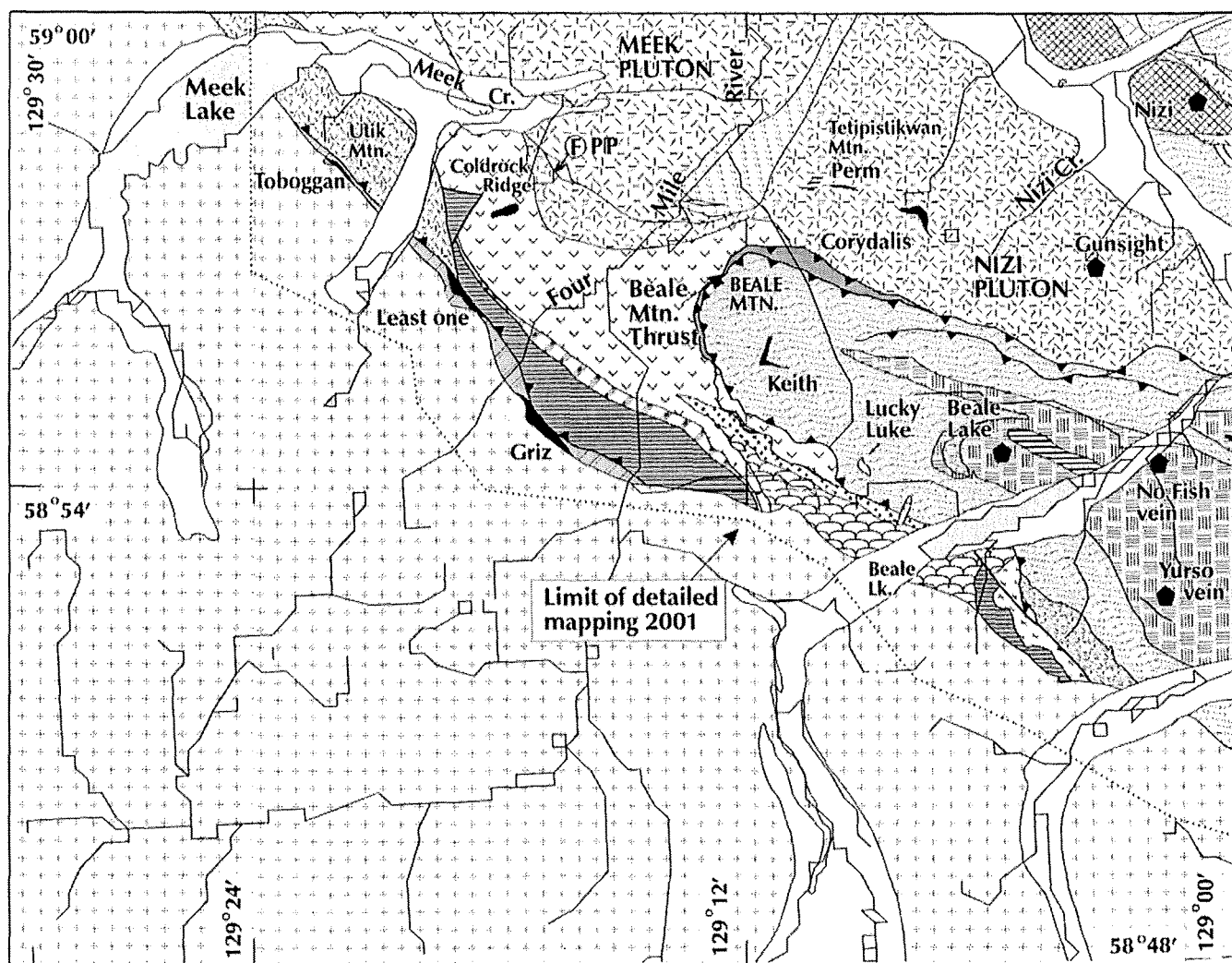


Figure 2. Geology of the Beale Lake (1041/N) map area

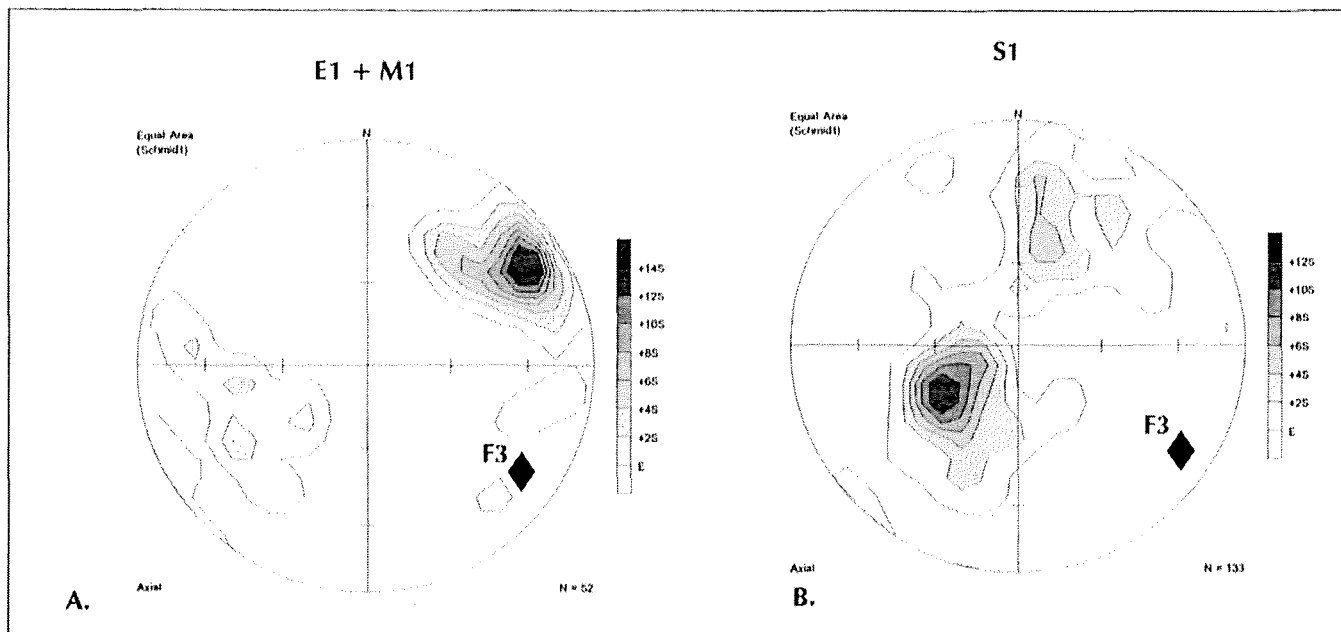


Figure 3. Structures in the Dorsey assemblage, Beale Lake map area.
A) E1, M1 lineations B) S1 foliations

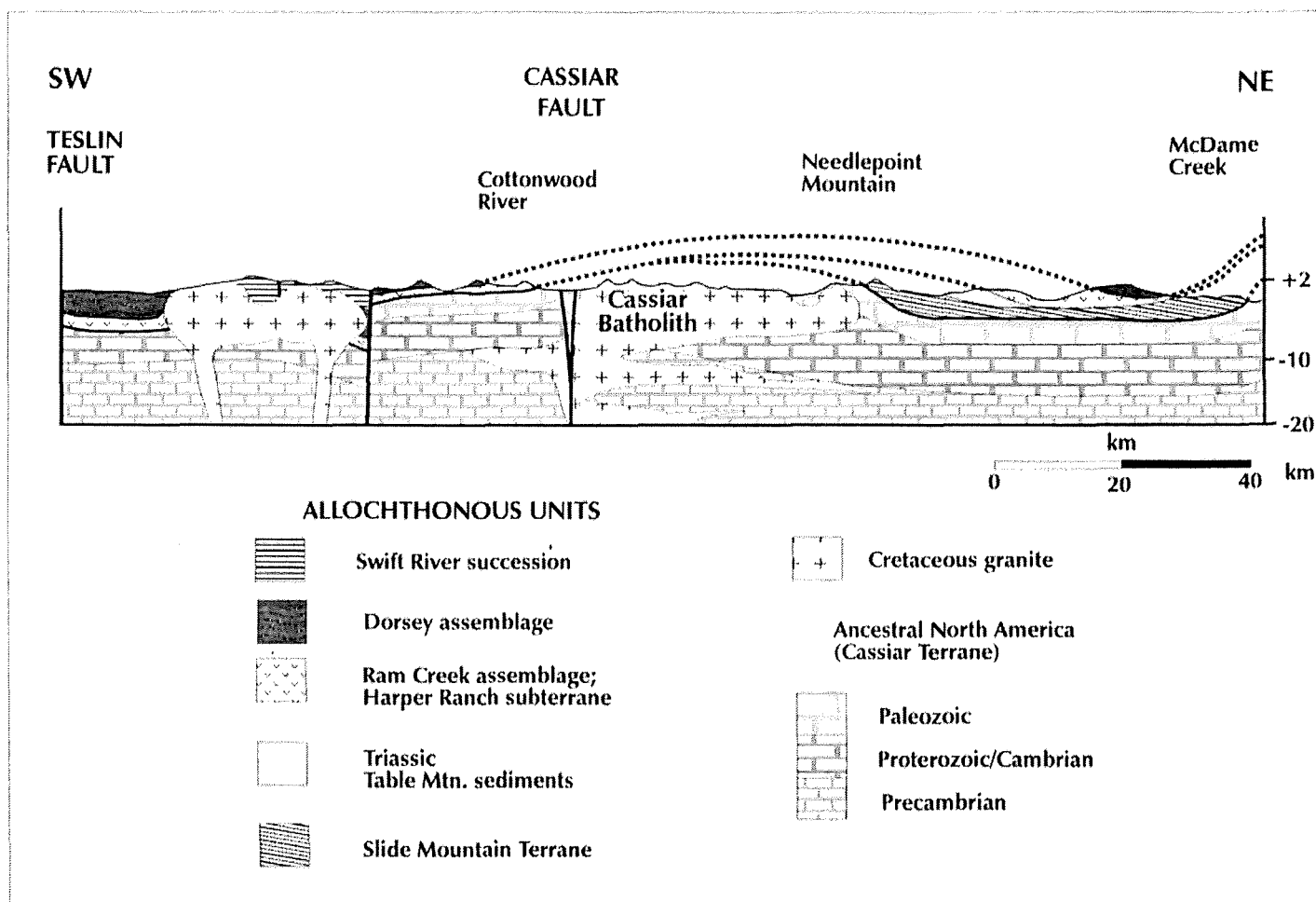


Figure 4. Cross section from Teslin Fault across Cassiar Batholith and Sylvester Allochthon, showing gentle attitudes of thrust-bounded panels

Mantle Exhumation in an Early Paleozoic Passive Margin, Yukon-Tanana Terrane, Yukon

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An orogenic peridotite body occurs as a megaboudin (200 m) structurally juxtaposed with smaller bounded masses (0.1 - 10 m) of corona-textured troctolite, garnet amphibolite and calc-silicate in quartzite and quartz-mica schist of the Yukon Tanana Terrane in the central Dawson Range, Yukon. The partially serpentinized peridotite shows well-developed plagioclase corona on spinel and records cooling and decompression from >1100°C 1.0 GPa, to below the spinel-plagioclase peridotite facies reaction at ~ 650°C. Pyroxenes occurring as coronae between olivine and plagioclase in the troctolite record cooling to 850°C through the same reaction. These rocks are interpreted as dismembered lower crustal mafic intrusions. Adjacent to the peridotite body, the quartzite country rock has layers of metapelite leucosome containing garnet and orthopyroxene, implying heating above the fluid-absent metapelite solidus (about 900°C) at pressures above cordierite stability (0.6 GPa). The structural juxtaposition of upper mantle peridotite and lower crustal mafic intrusions as boudins on various scales within continental margin metasediments heated to granulite grade, is explicable by the contrasting strength of upper mantle, lower and upper crustal lithologies during rifting of continental lithosphere at least 30 km thick. Re-Os isotopic compositions of the peridotite body show that it represents juvenile lithosphere that has a similar or slightly older age than the metasedimentary rocks in which it is boudinaged. The rocks in the study area bear striking geological, petrographical and geochemical similarities to those in the Zabargad massif in the Red Sea and are interpreted to have formed in an early Paleozoic pre-oceanic passive margin rift. Many other enigmatic mantle tectonite occurrences in continental margin metasediments along strike in the Yukon Tanana Terrane in Yukon and Alaska may have a similar origin.

Mid- to Late Paleozoic paleogeographic evolution of the pericratonic elements of the northern Cordillera: a hypothesis

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Ongoing regional mapping and related geochemical and geochronological studies in the northern Cordillera have established the regional stratigraphic framework and tectonic affinity of (some of) the pericratonic elements (Yukon-Tanana [YTT], Slide Mountain, and affiliates) which lay outboard of the Ancient Pacific Margin of North America in mid- to late Paleozoic time.

From Middle Devonian to Permian time, the pericratonic elements of the northern Cordillera formed as part of a system of arcs and related marginal basins. The geochemical and isotopic characteristics of igneous rocks of YTT consistently show that these arc complexes formed (at least in part) on older crust. We speculate that these crustal blocks were stranded after intermittent Late Proterozoic – Early Paleozoic extension along western Laurentia. Although the paleolatitudinal position of YTT with respect to northern Cordilleran miogeocline remains uncertain, the isotopic characteristics of igneous rocks of YTT are consistent with derivation of these crustal blocks from the basement terranes of northwestern Canada.

Arc magmatism in the northern Cordillera is first recorded in the Middle Devonian of the Alaska Range and Ecstall belt of northwest B.C. Magmatic activity in the Alaska Range-Ecstall belt coincides with alkaline magmatism and syn-depositional extension recorded in outer miogeoclinal strata of the Selwyn basin.

In Late Devonian, magmatic activity in YTT becomes more widespread and persists into Early Mississippian time. The distribution of arc (on the southwest) and back-arc (on the northeast) volcanic rocks indicates that these rocks formed above an east-dipping subducting slab. Late Devonian – Early Mississippian arc magmatism coincides with the onset of spreading of the Slide Mountain ocean.

Discrete Mississippian deformational events and the formation (and preservation) of Early Mississippian eclogites suggests a complex geodynamic evolution of the Early Mississippian arc system. These events are interpreted to indicate that YTT comprises an amalgamation of arc terranes which must have been separated by small ocean basins.

In Late Mississippian, arc magmatism occurs primarily in southern YTT. In northern YTT, arc magmatism was succeeded by carbonate deposition and the arc and back-arc regions were juxtaposed by thrust faults. Arc magmatism in the south persisted until the Early Permian. In Early Permian, the northern part of YTT evolved (rapidly) into a marginal basin (Campbell Range basalt). The overall evolution of YTT between Middle Devonian and Early Permian time can be explained as the result of multiple cycles of arc/back-arc construction and shortening which developed above an east-dipping subducting slab. These cycles are likely linked to changes in style of subduction (roll-back vs. shallowing) or overall plate configuration.

By mid- to Late Permian, the polarity of subduction changed, resulting in formation of the Klondike arc and consumption of the Slide Mountain ocean. West-dipping subduction is indicated by occurrences of forearc deposits (including a syn-orogenic conglomerate) and eclogites to the east of Klondike arc magmatism.

Racklan orogeny: deformational features and correlations in northwestern Canada

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The Racklan orogeny can be favourably correlated with the Forward orogeny in the Northwest Territories (Cook and MacLean, 1995) and the Fifteenmile orogeny in the Ogilvie Mountains (Mercier, 1989; Abbott, 1997; Brideau et al., 2002; Brideau, 2002). The region of orogenesis extends for at least 1000 km east-west, and 250 km north-south.

The term Racklan orogeny was originally used by Gabrielse (1967) to describe an angular unconformity between "Purcell-like" strata and the Rapitan Group in the North Rackla River area in the Wernecke Mountains of Yukon. Gabrielse's (1967) description was built on observations by Wheeler (1954), who recorded an angular unconformity between rocks of uncertain affinity. Subsequent work by Blusson (1974) and Eisbacher (1979) demonstrated that the unconformity actually lies between the Wernecke Supergroup (>1.71 Ga) and the Pinguicula Group (ca. 1.38 Ga; Abbott, 1997; Thorkelson, 2000). Most if not all of the orogenic activity within this interval appears to have occurred prior to emplacement of Wernecke Breccia at 1.60 Ga, which crosscuts schistosity and kink bands in the Wernecke Supergroup (Thorkelson, 2000; Thorkelson et al., 2001a).

Recent work in the Wernecke Mountains (Brideau et al., 2002; Brideau, 2002) has shown that the Racklan Orogeny consisted of at least three distinct deformation events. The first event produced a chloritoid +/- garnet and opaques porphyroblastic, chloritoid-chlorite-muscovite-quartz schist. P-T conditions have been estimated to lie between 3-6 kbar and 450-550 °C. The second event produced a crenulation and the third generated kink bands.

The Fifteenmile orogeny was recognized by Mercier (1989) on the basis of fabrics in weakly metamorphosed rocks beneath the Fifteenmile Group in the Coal Creek area of the western Ogilvie Mountains, Yukon. The deformed Proterozoic rocks described by Mercier (1989) were considered on the basis of microflora to be Paleoproterozoic in age (Mercier and Chauvel, 1987; Mercier, 1989), and were correlated with the Wernecke Supergroup by Roots and Thompson (1992) and Thompson (1995). The Fifteenmile Group is considered correlative, at least in part, with the Pinguicula Group (Cook and MacLean, 1995; Abbott, 1997), demonstrating that the Fifteenmile and Racklan orogenies belong to the same interval, and are likely to be manifestations of the same orogenic event.

The Forward orogeny (Cook and MacLean, 1995) was a contractional event that occurred during deposition of the ca. 1.66 Narakay Volcanic Complex (Bowring and Ross, 1985), part of the Hornby Bay Group in the Coppermine homocline near Great Bear Lake in the Northwest Territories. It is expressed as an angular unconformity between the Hornby Bay Group and the overlying Dismal Lakes Group (Cook and MacLean, 1995). The Forward orogeny's age correlates favourably with that of the pre-1.60 Ga Racklan events in the Wernecke Mountains.

The importance of the Racklan-Fifteenmile-Forward orogenic belt to continental and intercontinental tectonics is still not fully understood. Thorkelson et al. (2001b) drew attention to tectonic events of similar age in the southwestern United States (Yavapai and Mazatzal orogenies) and Australia (Isan and Kimban orogenies) and suggested the Racklan orogeny could be far-field expression of those orogenies. Whether these events and the Racklan orogeny are physically linked and genetically related remains uncertain.

he term "Racklan orogen" was used by Cook (1992) and Dredge Mitchelmore and Cook (1994) to describe a broad region of deformed rocks in Yukon and the Northwest Territories. Contractional structures post-dating the 1.27 Ga Coppermine River Group in the Coppermine homocline of the Northwest Territories were included as features of this orogen. Equivalent angular relations are evident between the Mackenzie Mountains Supergroup and the Tweed Lake basalts (Sevigny et al., 1991) in the Colville Hills (Anderson Plains) of the Northwest Territories, and between the Pinguicula Group and the Hematite Group in the Wernecke Mountains, and between units R1-R4 and unit R5 of the Lower Fifteenmile Group in the Ogilvie Mountains (Abbott, 1997; Thorkelson, 2000).

The structures ascribed by Cook (1992) to the "Racklan orogen" are now known to be >0.3 Ga younger than those which generated schistosity at lower greenschist grade in the Wernecke Mountains near the type area of the Racklan orogeny. For this reason, the Racklan orogeny as intended by Wheeler (1954) and Gabrielse (1967), advanced by Thorkelson (2000), and investigated by Brideau (2002) should be regarded as temporally and geologically distinct from the post-1.27 Ga deformation of Cook's (1992) "Racklan orogen." Consideration should be given to discontinuing the use of "Racklan orogen" for rocks deformed by events younger than the ca. 1.38 Ga Pinguicula Group.

Nomenclature aside, two periods of widespread contractional deformation in Early and Middle Proterozoic time (>1.60 Ga, and 1.27-1.0 Ga) are evident in western Canada. The significance of both events remains poorly understood.

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Preservation of Paleoproterozoic Slab volcanic megaclasts, Wernecke Mountains, Yukon

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The Paleoproterozoic Slab volcanics have been observed at one main and two minor localities in the Wernecke Mountains of northeastern Yukon. The main exposure is at the informally named "Slab Mountain," immediately north of the Bonnet Plume River (near Lat. 65°, Long. 134°). Minor localities are situated in the Bonnet Plume River valley, 10 and 15 km east of Slab Mountain, near Fairchild Lake. At all localities, exposed sides of the Slab volcanics are in contact with zones of 1.60 Ga Wernecke Breccia of hydrothermal origin. This field relation supports the hypothesis that the Slab volcanics have been preserved as megaclasts and were not emplaced into the country rock by another mechanism such as faulting.

The main volcanic block, measuring about 600 meters by 250 meters, dominates the eastern portion of Slab Mountain (Figure 1). The block consists of 35 thin (0.8-14 m), steeply dipping mafic to intermediate lava flows with rare interbedded volcanoclastic beds. The flows are dominantly massive with scoreaceous flow tops; ropey pahoehoe texture is visible locally. Boulder sized clasts of Slab volcanics which are surrounded by Wernecke Breccia crop out on "Slab Ridge" (Figure 1) and are included in this locality.

The minor localities are smaller and not as well exposed as the main locality, but show the same lithologic relationships as the main succession. Lava flow texture is obscured by a local conspicuous spotty hornfels texture caused by replacement of plagioclase and other minerals by scapolite.

The ca. 1.71 Ga dioritic and gabbroic Bonnet Plume River Intrusions (BPRI) are in direct contact with the Slab Volcanics at one locality and crop out throughout the study area. Preliminary geochemical analyses indicate a magmatic relationship between the Slab volcanics and the BPRI. It has been hypothesized that the BPRI were likely feeder dykes to the volcanism (Thorkelson et al, 2001a). Further analysis is necessary and ongoing to test this hypothesis.

The fortuitous preservation of the blocks of Slab volcanics is likely a result of clastic fall-back of the blocks into dilatant zones near the tops of the breccia zones during or immediately after explosive venting of fluids (Thorkelson 2000). Upward transport of the Slab volcanics seems unlikely due to the enormous size of the blocks, especially at Slab Mountain; the hydrothermal fluids involved in brecciation are not expected to have the ability to entrain blocks with such small surface/volume ratios.

Metasomatism during the 1.60 Ga hydrothermal-breccia activity caused the alteration and precipitation of minerals which is observed in the Wernecke Supergroup, Slab volcanics and BPRI (Thorkelson et al., 2001b). Metasomatic effects of the hydrothermal activity on the Slab volcanics include replacement of plagioclase by albite, potassium feldspar or scapolite, and replacement of augite by chlorite or actinolite.

The Slab volcanics are younger than the 1.71 Ga Wernecke Supergroup and older than the 1.60 Ga Wernecke Breccia. Erosion after eruption, during the 200 Ma years before deposition of the overlying Pinguicula Group, removed all evidence of the volcanic succession. The only remnant of this volcanic event are the three localities mentioned as there are no correlative proximal or distal volcanic rocks in the surrounding Wernecke and Ogilvie Mountains (Thorkelson 2000).

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Yakutat Block Collision in the Gulf of Alaska and Stress Transfer Across the NE Canadian Cordillera

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The collision of the Yakutat block with the North American plate in the SE corner of the Gulf of Alaska produces large uplift and deformation in the adjacent St Elias mountains. Surprisingly, this collision also results in strong deformation about 600-800 km to the northeast of the collision zone: thrust faulting seismicity in the MacKenzie mountains near the eastern Canadian Cordillera deformation front, and strike-slip faulting to the northeast in the Richardson mountains region. We propose that stress is transmitted from the Yakutat collision across the whole northern Cordillera with little intervening deformation, despite the high heat flow and the inferred thin mechanical lithosphere. To support this hypothesis, we have developed a kinematic and mechanical model of stress transfer from the collision zone to the Cordillera deformation front.

The kinematics is constrained by campaign GPS velocities recently acquired in the Yukon and Northwest Territory, and by the Whitehorse-Yellowknife permanent GPS baseline. GPS results indicate a northeast to north-northeast motion of the northern Cordillera at about 5 mm/yr, relative to the stable craton to the east. We also show seismic moment release statistics that indicate 3-5 mm/yr of shortening across the MacKenzie range and about 4 mm/yr of right-lateral strike-slip in the Richardson mountains to the north, in agreement with the GPS velocities.

This northeast motion of the northern Cordillera requires a horizontal quasi-rigid displacement of the upper crust accommodated by a detachment in the weak lowermost crust. Using a 2D Finite Element mechanical model, we show that the horizontal strain transfer over 600-800 km and the development of a lower crust detachment zone is made possible by the high temperature of the northern Cordillera lithosphere (up to 800-1000°C at the Moho based on very high heat flow). Our model also indicates that this strain transfer requires a weak eastern Cordillera deformation front where the lower crust shear zone rises to reach the basal detachment of thrusting in the foreland basin.

STRUCTURAL STYLE OF NORTHWEST BOWSER BASIN, NORTHERN BRITISH COLUMBIA: UPPER CRUSTAL ANALYSIS OF A PORTION OF SNORCLE REFLECTION LINE 2A AND STRUCTURAL FIELD MAPPING.

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In an attempt to clarify the structural history of the Skeena Fold Belt, seismic reflection data for SNORCLE Line 2A are used in combination with detailed structural mapping of the northwest Bowser Basin. The region studied lies along the Stewart-Cassiar Highway between Bell II and Iskut, a distance of 50 km.

The Skeena Fold Belt is a large, thin-skinned, fold and thrust belt dominating the width of the Stikine Terrane in northern British Columbia. Northwest trending folds dominate the central and eastern parts of the belt and are found in large areas of the western fold belt along with discrete areas of northeast trending folds [Evenchick, 1998]. The deformation history of the western region of the fold belt is contained within the relationship of these northeast and northwest trending structures. Structural mapping in three areas along the eastern side of the highway was undertaken to clarify the relationship between the two fold sets and investigate the deformation history of the western fold belt as suggested by Evenchick [2001]. The three areas studied are entirely within Bowser Basin sediments. Study areas 1 and 3 are situated in the centre and south of the region and comprise strata of the middle to upper Jurassic, Ritchie-Alger Assemblage [Evenchick *et al.*, 2001]. The areas studied were characterised by thick sequences of interbedded sandstones and siltstones and locally dominant conglomerate units. Study Area 2 in the north of the region comprises sediments of the Skelhorne Assemblage [Evenchick *et al.*, 2001], and in the area studied was characterised by dominant conglomerate units and minor interbedded sandstone units.

Study Area 1, located 25 km south of Iskut, contains both northwest and northeast trending folds. Northwest trending folds, having both southwest and northeast vergence and wavelengths varying from 100-800 m, dominate the area. Northeast trending folds identified in the area have wavelengths of >100 m and verge northwest. A disjunctive spaced cleavage is found with varying concentration throughout the area and has extremely variable strike orientation, in some cases >90°. In most cases, cleavage appears oblique to both fold sets. In rare cases, cleavage could be interpreted to be axial-planar to northwest trending folds; no

cleavage associated with northeast trending folds was identified. Common extensional fractures lying within the a-c and b-c tectonic planes [McClay, 1987] are associated with the northwest trending folds. Conjugate shear fractures normally associated with the northwest trending folds also are common. In a small number of cases, conjugate shear fractures situated on northwest trending folds that are oblique to this fold set have been identified and may indicate previous deformation. Faults and quartz veins are common throughout the area. Two main styles of veins are present. One style comprises concentrated areas (5 m²) of thick vuggy veins, with non-systematic orientation and occasional areas of tightly folded veins trending northwest. The second style of veins are more commonly distributed across the entire study area and comprise thin (<1 cm) straight, parallel vein systems, which strike northeast, interlinked by thinner veins at an oblique angle. Evidence from this area suggests initial discrete northeast trending folding followed by more common northwest trending folding.

Study Area 2, located 15 km south of Iskut, contains a large (>2 km diameter) basin. This is interpreted to be a Type 1 interference structure produced by the interaction of northeast and northwest trending structures [Evenchick, 2001]. A northeast trending fold associated with strongly oblique cleavage is located to the north of the basin. Bedding-cleavage intersection lineations throughout the area imply a northeast striking cleavage, which has subsequently been refolded.

Study Area 3, located 15 km north of Bell II contains northeast trending folding with axial planar cleavage. Cleavage data from this area display some variation in strike and bedding-cleavage intersection lineations indicate some degree of refolding. Fracture data from the area also show evidence of an oblique angle to northeast trending folds.

These structural observations were then compared with a seismic reflection data set acquired through the region. The SNORCLE Line 2A data were acquired along Highways 37A and 37 from Stewart, BC to Watson Lake, YT. From the final stacked seismic section produced by staff at the LITHOPROBE Seismic Processing Facility for SNORCLE, we took a 50 km segment. Since our study concentrates on the upper crust, we considered only the upper 8s (~25 km) of the data.

Our primary effort involved defining the velocity variation within the upper crust (<3 km) through analysis of the first-break refraction travel time picks on the shot gathers, as provided by the LSPF processors. To ensure reliable velocity determinations, we applied three different procedures to these data: 2-D ray-trace modelling using RAYINVR [Zelt and Smith, 1992], travel time inversion using FAST [Zelt and Barton, 1998] and GLI-3D (a commercial software package), which allows the crooked line geometry to be considered. The shallow

velocity variations are compared with known or inferred geology along the seismic line. Two velocity models to 25 km depths were generated for the 50 km line segment. One used the shallow velocity model merged with results derived from standard velocity analyses of the reflection data as provided by the processors. The second crustal model merged the shallow velocity structure with one derived from analyses of the SnoRE 97 refraction data [Hammer *et al.*, 2000]. The 8 s of stacked seismic data for the region were then migrated using the two crustal velocity models.

Previous work identified a significant change occurring at 3.5-4 km depth and suggested it represented the base of the Bowser Lake Group. This conclusion was based on a rapid increase in velocities at this depth [Hammer *et al.*, 2000] and magnetic properties [Lowe *et al.*, 1998]. However, the reflection profile in our study reveals a strong reflector at 3 s (~9 km depth) that we interpret to be the base of the Bowser Basin. The reflection data identify a degree of variation on the basement of the basin and the northern boundary of the basin with the volcanic Hazelton and Stuhini rocks (10 km south of Iskut). Structural data suggest a deformation history for this area of the Skeena Fold Belt that involves initial northeast trending folding followed by northwest trending folding. Cleavage may be indicating variable deformation histories on a local scale (10's of km). The style of deformation throughout the region indicates the depth at which deformation occurred was shallow, probably <6 km [Salvini and Storti, 2001]. The style and abundance of quartz veins throughout the study areas suggest these sediments were relatively wet when deformation occurred and fluid overpressuring was a strong component of deformation [Etheridge *et al.*, 1984]. The mesoscopic style of deformation present in the region suggests that deformation in the Skeena Fold Belt is more akin to accretionary prism deformation than classic fold and thrust belts such as the Rocky Mountains.

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Neogene Magmatism in Stikinia: Insights from Hoodoo Mountain volcano

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Introduction

The northern Cordilleran volcanic province (NCVP; Edwards and Russell 2000) (Fig. 1, inset) is one of the largest Neogene volcanic provinces in western North America. The NCVP comprises dominantly mafic, Neogene to Quaternary, alkaline volcanic rocks thought to result from extensional forces acting on the northern Cordilleran lithosphere (Souther 1977; Edwards and Russell 1999; 2000). Three large, distinctive volcanic centers in northwestern British Columbia form the central portion of the NCVP: Mount Edziza (Souther et al. 1984; Souther 1992), Level Mountain (Hamilton 1981) and Hoodoo Mountain (Edwards et al. 2000) (Fig. 1). Relative to most other NCVP volcanoes, these are long-lived edifices comprising, in part, large volumes of intermediate to felsic alkaline lavas that do not derive directly from asthenospheric mantle sources (Hamilton 1981; Souther and Hickson 1984; Souther 1992; Edwards et al. 2000). Another trait shared by these large-volume, long-lived, volcanic edifices is that they all overly rocks of the Stikine terrane (Fig. 2).

Our objectives in this paper are 2-fold. Firstly, we present new petrochemical data on the phonolites and trachytes that comprise the least studied of these large volcanic centres: Hoodoo Mountain volcano (HMV). Secondly, these data are used in conjunction with geophysical datasets (Hammer et al. 2000; Hyndman & Lewis 1999) to establish the physical-chemical origins of the magmas that fed HMV. Such peralkaline magmas are most efficiently produced where alkali olivine basaltic magma stalls in the mid-crust and assimilates felsic rocks (e.g., Souther and Hickson 1984; Edwards and Russell 1998; Russell and Hauksdóttir 2000).

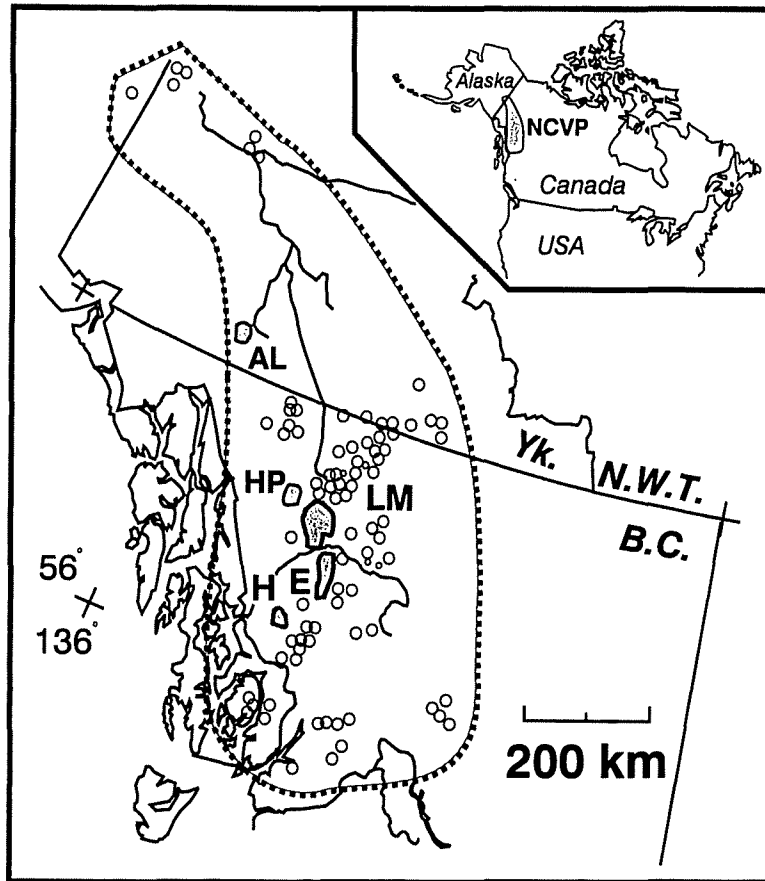


Figure 1. Inset shows location of northern Cordilleran volcanic province (NCVP; Edwards & Russell, 1998; 2000). Main figure locates specific Neogene and Quaternary volcanic centres within NCVP, including Hoodoo Mountain (H), Edziza (E), Level Mountain (LM), Hearts Peak (HP) and Alligator Lake (AL) (After Edwards et al. In press).

Geological Setting

Hoodoo Mountain volcano (HMV) (56°46' N, 131°17' W; UTM center 360000E 6294500N) is a Quaternary, phonolitic volcano situated immediately north of the Iskut River in the Coast Mountains of northwestern British Columbia, Canada (Fig. 1). Its activity spans the last 100 k.y., it may have erupted as recently as 9 ka, and it encompasses a volume of approximately 17 km³. Kerr (1948) briefly described volcanic deposits at Hoodoo Mountain during a study of the regional geology along the southern part of the Iskut River, and Souther (1991a) provided a short account of its striking morphology and some age-constraints from preliminary K-Ar dating. The striking, step-like topographic profile and bounding cliffs of

HMV led Kerr (1948) and Souther (1991a) to suggest that many lava flows at HMV were obstructed or “dammed” by glacial ice. Edwards et al. (2000) gave the first detailed summary of the Quaternary stratigraphy and petrology of the Hoodoo Mountain volcanic complex.

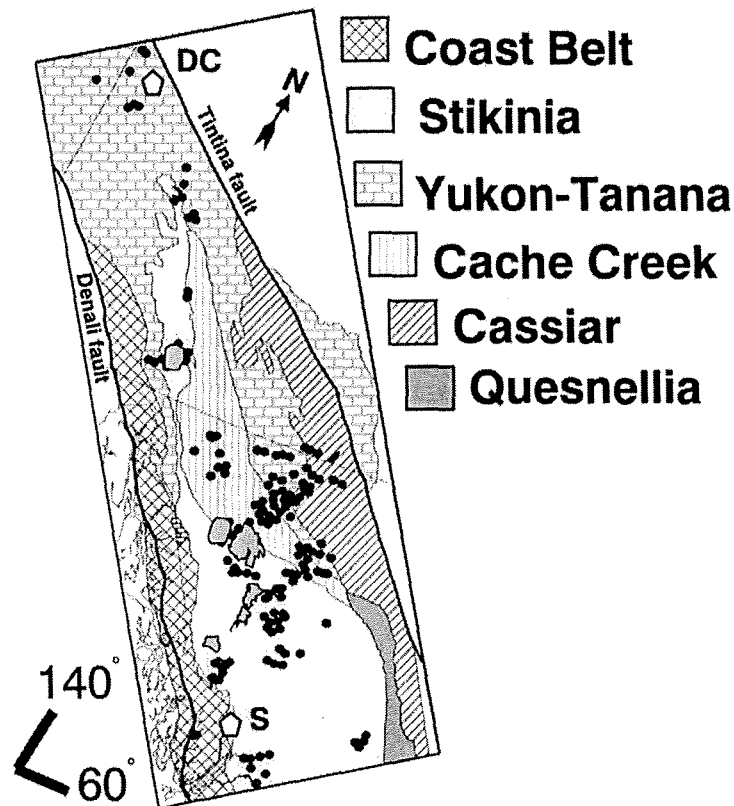


Figure 2. Distribution of all Neogene to Quaternary volcanic centres assigned to the NCVP draped on the tectonostratigraphic terranes of the northern British Columbia cordillera. Solid black dots denote relatively small centres or remnants of centres and are generally basalt, nephelinite and/or basanite in composition. Large, long-lived evolved centres situated in Stikinia are denoted by dark grey shading (see Fig. 1 for labels).

The volcano comprises fragmental and non-fragmental peralkaline volcanic rocks, including subglacial and subaerial lava flows, domes, spines, dykes, and pyroclastic deposits. The development of the edifice overlaps in space and time with formation of nine small basaltic centers in the Iskut volcanic field (Stasiuk and Russell 1989; Souther 1991b; Hauksdóttir et al. 1994; Russell and Hauksdóttir 2000). One of these basaltic centers, Little Bear Mountain (volume $<0.3 \text{ km}^3$), is located immediately north of Hoodoo Mountain and has been partly overridden by lava flows from HMV.

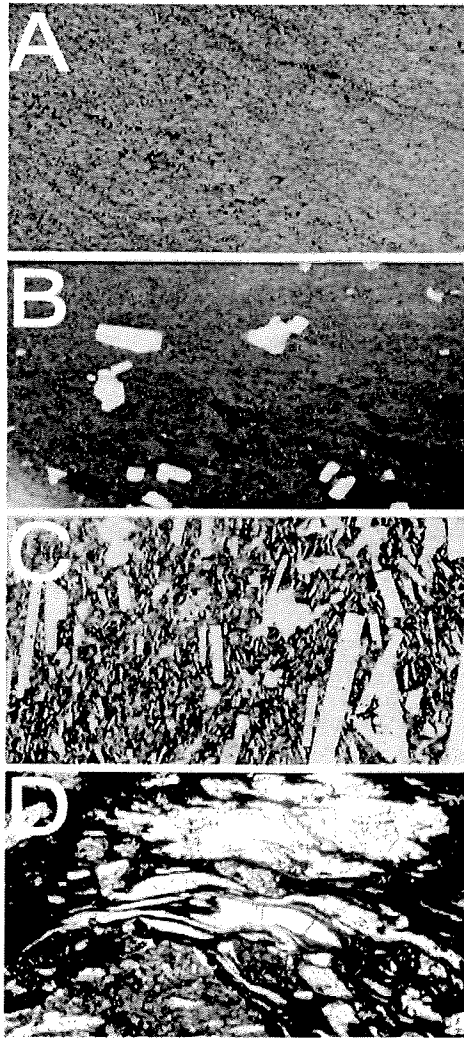


Figure 3. Plane-polarized scanned images of thin sections of volcanic rocks from HMV. Images show: (A) aphanitic character of stratigraphically lower lavas (Qvap₂); (B) weakly alkali-feldspar porphyritic character of lavas from upper stratigraphic levels (Qvap₅); (C) and highly porphyritic (alkali feldspar >> clinopyroxene > magnetite) and trachytic character of youngest lava flows (Qvpp); (D) eutaxitic texture in pyroclastic succession (Qvpy). Fields of view for A–C are 2.5 by 3.5 cm and for D is 1 by 2 cm.

Mineralogy

Mineralogically, all rocks from HMV are nepheline-bearing alkali-feldspar trachyte (IUGS classification, LeMaitre 1989) and contain euhedral to subhedral alkali feldspar (AF; dominantly anorthoclase), magnetite (Mt) and Fe-rich augite (Cpx), both as phenocrysts and in the groundmass. Nepheline (Ne) and a Na–Al–Si–Cl bearing phase (possibly cancrinite) are also found in the groundmass of some samples. Massive lava flows (e.g., Qvap and Qvpp) vary from aphyric (< 1 % phenocrysts) to abundantly porphyritic (> 15 % phenocrysts) (Fig. 3A–C), but all samples have an aphanitic matrix. Most lavas are holocrystalline to cryptocrystalline and

trachytic to pilotaxitic. Volcanic glass is only abundant in pyroclastic deposits (e.g., unit Qvpy; Fig. 3D). Higher stratigraphic units (e.g., unit Qvpp) have higher abundances (15 – 20 vol %) and larger sizes (> 2 cm) of AF phenocrysts. The youngest Qvpp lavas commonly contain glomerocrysts of AF, Mt, and Cpx, and locally inclusions of troctolite and gabbro.

Major element geochemistry

A complete suite (n=48) of major and trace element chemical compositions for HMV rocks is available on the internet at <http://perseus.eos.ubc.ca>. The compositional range of the HMV phonolites and trachytes is small relative to the two other large volcanic centers the Stikine subprovince, Mount Edziza and Level Mountain; all but one HMV samples contain between 56 and 60 weight % SiO₂ (Fig. 4A). Samples of volcanic glass from unit Qvpy and the majority of samples of Qvap₁₋₅ are phonolite or transitional between trachyte and phonolite. Samples of unit Qvpp plot as trachyte, near the phonolite boundary.

Samples of HMV lavas are nepheline normative except for a single quartz normative sample from the base of the volcano. Samples of HMV lavas also have normative acmite and are chemically peralkaline. However, samples of glass from pyroclastic units are the most sodic and agpaitic of all the samples, while samples of porphyritic Qvpp are slightly less agpaitic compared to samples of Qvap with similar values of K/Na.

The homogeneous and undersaturated nature of HMV lavas distinguishes them from the intermediate to felsic volcanic rocks found at Level Mountain and Mount Edziza. Samples from units at Level Mountain and Edziza include quartz-saturated peralkaline rhyolites in addition to quartz-free peralkaline and metaluminous trachytes (Hamilton 1981; Souther and Hickson 1984; Souther 1992). Compared to trachytes from Level Mountain and Edziza, the HMV trachytes and phonolites are more agpaitic (lower values of Al/(Na + K)) and also more sodic (lower values of K/Na). More than 90 % of the samples from Level Mountain, Edziza and HMV have K/Na ratios less than that for sanidine (0.59); this is consistent with the dominance of phenocrystic anorthoclase in HMV rocks. Virtually all HMV lavas are peralkaline whereas at Mount Edziza and at Level Mountain lavas are peralkaline and metaluminous in subequal proportions.

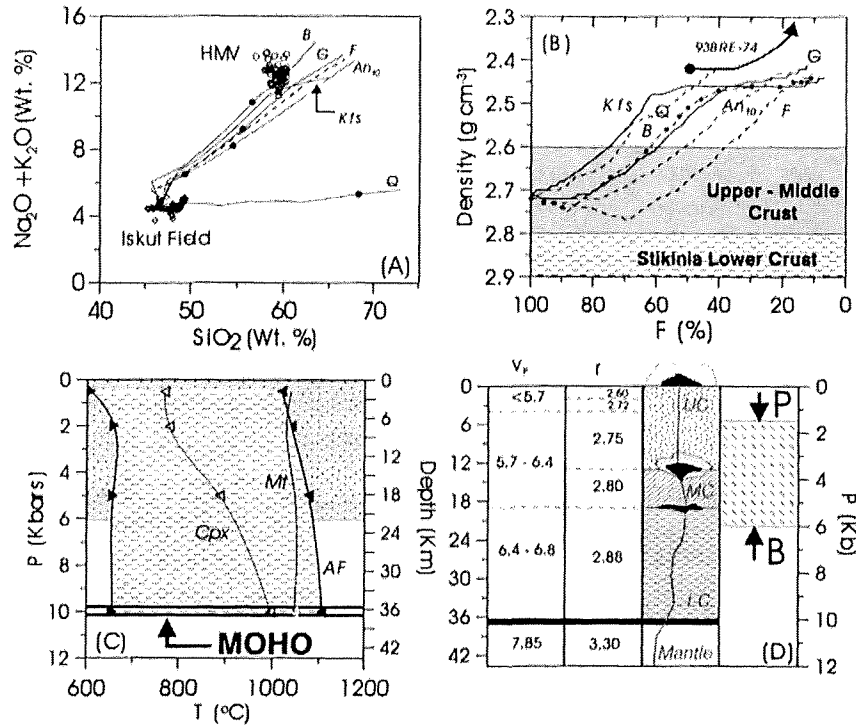


Figure 4. Chemical model for origins of phonolitic magmatism at HMV. **(A)** Compositions of HMV rocks and Iskut basalts shown as $\text{Na}_2\text{O} + \text{K}_2\text{O}$ vs. SiO_2 and compared to liquid lines of descent produced by AFC processes involving basalt and crustal components, including: quartz (Q), K-feldspar (Kfs), plagioclase (An₁₀), Mg-rich biotite (B), and granite (G). Closed system fractionation path (F) is also shown. Filled circles indicate where residual melt is 50 % of the original mass. **(B)** Same model paths plotted as melt density vs. percent melt (F %) in system. Density of Stikinia crustal rocks (shading) are from (Hammer et al. 2000). Heavy line shows low pressure fractionation path for HMV phonolite. **(C)** P-T crystallization phase diagram calculated for HMV phonolite. Dashed pattern shows the P-T region of stability for the phase assemblage (AF, alkali feldspar; Mt, magnetite; Cpx, hedenbergite) expected for HMV magmas. Lines with symbols denote near-solidus conditions (left with solid triangles) and first appearance for Cpx (open triangles), Mt (shaded triangles), and AF (right with solid triangles). Phenocryst assemblages in Iskut basalts (stippled pattern) suggest crystallization at pressures below 6 Kb. **(D)** The crustal properties of Stikinia as constrained by seismic data including V_p and density, used to define thicknesses of upper (UC), mid- (MC) and lower (LC) crust and depth to MOHO (Hammer et al. 2000; Hammer and Clowes 2000). Panel also portrays ascent of asthenospheric-sourced basaltic magmas (black) to mid to upper crustal depths where AFC processes produce peralkaline phonolitic magmas of HMV. The depth of AFC processes (right-hand panel) is constrained by phenocrysts in basaltic rocks of the Iskut field (B, maximum) and based on the interpretation that AF, instead of Mt, is the liquidus phase in HMV lavas (P, minimum; see panel C).

Chemical Evolution

Volcanism at HMTV has produced $>17 \text{ km}^3$ of compositionally restricted phonolite and trachyte over a 80 k.y. time interval. While the compositional characteristics of the contemporaneous Iskut basalts can be attributed to asthenospheric sources (e.g., Carignan et al. 1994; Cousens and Bevier 1995; Edwards and Russell 2000; Francis and Ludden 1990) and minor crustal contamination (Russell and Hauksdóttir 2000), the large volume of phonolite and trachyte at Hoodoo Mountain requires an alternative explanation.

Trachyte and phonolite melts can result from diverse processes, including: i) partial fusion of alkali olivine basalt (Hay and Wendlandt 1995; Kaszuba and Wendlandt 2000); ii) fractional crystallization of basanitic or alkali olivine basaltic melts (e.g., Le Roex et al. 1990; Kuntzmann and Blasse 2000; Kyle et al. 1992); and iii) coupled fractional crystallization and assimilation (e.g., Edwards and Russell 1998; Kyle et al. 1992; Wilson et al. 1995). Our analysis indicates that HMTV phonolites and trachytes derive from coupled assimilation–fractional crystallization (AFC) involving primary mantle–derived alkali olivine basaltic magmas and granitic crustal rocks.

This scenario is supported by a number of lines of evidence. At HMTV, the ratio of basalt (Little Bear Mountain) to phonolite is extremely low, 0.3 km^3 to $>17 \text{ km}^3$, particularly compared to the other two major centers in the Stikine subprovince, Mount Edziza (1:1; Souther 1992) and Level Mountain (4:1; Hamilton 1981). We argue that the apparent lack of large volumes of basalt within the Iskut region is most consistent with an origin by coupled fractional crystallization and assimilation. The alkali olivine basalt from the Iskut volcanic field is clear evidence of widespread, contemporaneous mantle–derived, mafic magmatism. These basaltic magmas carry abundant xenoliths and xenocrysts of crustal material (Stasiuk and Russell 1989; Edwards and Russell 2000; Russell and Hauksdóttir 2000); many show evidence of melting during transport and have chemically modified some basaltic magmas (Cousens and Bevier 1995; Russell and Hauksdóttir 2000). Similar direct evidence of magmatic assimilation is common at other volcanic centers throughout the NCVP including: Level Mountain (Hamilton 1981), Mount Edziza (Souther 1992), the Tuya Buttes area (Watson and Mathews 1948), Alligator Lake (Eiche et al. 1987), Fort Selkirk (Francis and Ludden 1990), and Prindle volcano (Foster et al. 1966).

To test the efficacy of the AFC process, we have computed model liquid lines of descent

for six different AFC scenarios (Fig. 4A) using the thermodynamics-based program MELTS (Ghiorso and Sack 1995). The model paths are calculated using a parental magma of alkali olivine basaltic composition, consistent with basalt from the Iskut volcanic field (Lava Fork, SH-44), and include: closed system fractional crystallization (F), and fractional crystallization coupled to assimilation of granite (G), quartz (Q), albite (An_{10}), K-feldspar (Kfs) and Mg-rich biotite (B) (after Edwards and Russell 1996).

Fractional crystallization of alkali olivine basalt can produce liquids approximately equivalent to HMV rocks but the system is over 80 % crystallized by the time it produces trachyte. However, AFC processes involving the incorporation of granitic material allow the system to reach the field of HMV samples while a substantial melt fraction remains (Fig. 4A). Thus, compared to AFC processes, closed system fractional crystallization is a much less efficient means to produce magmas similar to the HMV samples. Scenarios involving assimilation of individual constituent minerals of the granite are used (Fig. 4A) to show the relative effects of increasing the modal amounts of these phases in the average crustal contaminant. Increasing quartz content, for example, drives the system away from the compositional field for HMV rocks. Conversely, increasing the K-feldspar or Mg-biotite content of the average crustal material facilitates reaching the HMV field; assimilation of K-feldspar-rich granitoids produces compositions equivalent to HMV phonolites at a point where 50 % of the original mass is still melt. These simulations clearly demonstrate that the volumes of derived phonolitic magma available are maximized by AFC processes, relative to the volumes derived from fractional crystallization alone.

These model differentiation paths are also unique in their physical properties: melt density and percent melt remaining in the system (F) (Fig. 4B). Lower density values enhance buoyancy-driven, extraction of magma from the crustal lithosphere. Reaching phonolite compositions at higher values of F produces a greater volume of phonolite for a given mass of parental basalt. Again, closed system fractionation is a less viable explanation for HMV because, not only are values of F low when the phonolite field is reached, but the density of the residual melts are equivalent or higher than the densities of the crustal rocks underlying HMV (e.g., Hammer and Clowes 2000). The melts produced by this path are unlikely to erupt under lithostatic pressures until the system achieves 60% crystallization. Conversely, melts resulting from AFC processes involving granite or K-rich granitic material become buoyant relative to

Stikine crust at 30% crystallization (ie., melt fractions as high as 70 %; Fig. 4B). The densities of melts along these paths compare favorably to the densities of melts computed for HVM phonolites and their fractionation products (Fig. 4B; heavy line). In summary, the AFC processes that involve alkali olivine basaltic magma from the Iskut volcanic field and incorporation of granitic rocks are clearly capable of producing extractable, voluminous magmas similar in composition to the HVM phonolites.

Physical Storage of Magmas

Several lines of evidence are consistent with the AFC processes taking place at mid- to upper crustal depths. Edwards and Russell (2000) and Trupia and Nicholls (1996) have shown that for many NCV basaltic magmas crystallization of phenocryst assemblages occurred at pressures between 0.4 and 0.8 GPa or at depths between 18 and 29 km. Phenocryst assemblages in lavas from the Iskut volcanic field suggest crystallization at pressures up to 0.6 GPa. Secondly, we have calculated an equilibrium crystallization map for HVM phonolite using MELTS (Fig. 4C). These calculations, used in conjunction with the observed petrography, limit the ponding of the magma and crystallization of the phenocryst assemblage ($Af_{sp} \gg Mt > Cpx$) to confining pressures > 0.15 GPa. Lastly, recent data from Lithoprobe have produced a geophysical picture of the part of Stikinia that underlies HVM (Hammer et al. 2000; Hammer and Clowes 2000). The seismic data show the depth to the MOHO (~ 37 km) and the nature of density stratification in the crust. The crust of Stikinia features a substantially denser lower crust ($2850\text{--}2900\text{ kg m}^{-3}$) between 37 and 19 km depth (Hammer and Clowes 2000). The mid-crust (19–13 km depth) is defined by a 2–3 % reduction in density (2800 kg m^{-3}) and the transition to upper crust (13 km) coincides with an additional 2–7% decrease in average density ($2600\text{--}2750\text{ kg m}^{-3}$) (Fig. 4D). Hyndman and Lewis (1999) suggest that, due to the high ambient heat flow in the Cordillera, the brittle–ductile transition may be situated at a depth of 10–12 km.

On the basis of these observations the transition between upper and middle crust is probably marked by a substantial change in composition and rheology. We suggest that this pronounced break in properties of crustal rocks serves as a filter to some asthenosphere–derived magmas. Magmas derived from relatively deeper levels of the asthenosphere are compensated for by lithostatic forces that allow them to traverse the entire crustal lithosphere and erupt; many

of these magmas are nearly aphyric and entrain mantle–derived xenoliths attesting to their relatively fast and uninterrupted path through the lithosphere (Edwards and Russell 2000). A second class of magma is represented by basalt erupted in the Iskut volcanic field and many other places within the NCVF (Edwards and Russell 2000). These lavas are olivine +/- plagioclase porphyritic, never contain mantle–derived xenoliths, but commonly entrain crustal xenoliths and xenocrysts. These magmas have probably ponded within the crustal column and undergone crystallization prior to erupting (Fig. 4C,D). The phonolitic rocks at HMV are an expression of these stalled magmas continuing to interact with the crust and creating new melt compositions.

Conclusions

Hoodoo Mountain volcano formed over a protracted period of time and reflects a complex combination of physical and chemical processes. The lack of significant chemical and mineralogic variability over this same time period argues for the sustained production of evolved magma of nearly uniform composition. Field observations, petrological data sets, and thermodynamic modeling support derivation of large volumes of low density phonolite magmas from alkali olivine basalt by a combination of fractional crystallization and crustal assimilation at mid–crustal pressures. The attendant changes in magma density permitted continued buoyant ascent and eruption of phonolitic magmas that ultimately derived from parental basalt magmas that had stalled in the Stikine crust.

Acknowledgements

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New U-Pb and geochemical data from the Barkerville Subterrane

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The BC Geological Survey Branch recently undertook a regional mapping program in the Cariboo Lake area of the Barkerville Subterrane. This subterrane is represented by the Late Proterozoic to Paleozoic Snowshoe Group, a deformed package of siliciclastics, volcanics and carbonates. The main objective of this project was to determine the stratigraphic and structural setting of the newly discovered Frank Creek massive sulphide showing. In addition, several mafic volcanic sequences within the Snowshoe Group were examined and sampled for whole rock geochemistry.

One of the main outcomes of this mapping program is the suggestion that, in the Cariboo Lake area, several units presently defined within the Snowshoe Group may be structural repeats. Specifically, rocks of the Downey and Ramos successions may be equivalent to the Keithley succession. In addition some or all of the Hardscrabble Mountain succession may equate with the Harveys Ridge succession. Furthermore, it is also suggested that the Bralco limestone is Early Cambrian in age.

The deformed nature of the Snowshoe Group, together with the lack of suitable material for geochronological dating, has resulted in a paucity of detailed age controls for this package. Diorite to quartz diorite intruding Snowshoe stratigraphy was sampled at three different localities for U-Pb geochronology in hopes of better constraining the age of this package. Two samples were taken from a large body of diorite intruding rocks presently assigned to the Downey succession. This body is sub-alkaline, displays arc-like geochemistry and is along strike from volcanics in the Mt. Barker area that contains sills and dikes of similar composition. The third dated rock was sampled from a sill intruding Harveys Ridge clastics south of Keithley Creek. This sill and other intrusions in the area, have trace element geochemistry suggesting MORB affinities. Diorite in this area also displays within plate signatures.

The ages of all three samples are similar. The two samples from the Mt. Barker area yielded ages of 277.3 ± 4.8 Ma and 281.0 ± 5.2 Ma, and the sample from the Keithley Creek region gives an age of 281 ± 12 Ma.

Originally, it was thought that the subalkaline diorite along Mt. Barker was co-magmatic with Mt. Barker volcanics. Chemically the two are quite distinct, with the latter being mafic alkaline volcanics. Regional correlations, together with trace fossil evidence, indicates that Keithley and Harveys Ridge rocks are Late Proterozoic to Early Cambrian, whereas the Downey succession may as young as the Ordovician (although the senior author believes the latter to be equivalent to Keithley rocks). Therefore, diorite sills and dikes sampled here for U-Pb geochronology are considerably younger than the hosting Snowshoe stratigraphy and only give a broad upper age limit.

The age and chemistry of diorite intruding the Harveys Ridge succession near Keithley Creek is similar to oceanic rocks of the Slide Mountain Terrane, represented in

this area by the Crooked amphibolite and Slide Mountain Group. If they are related, then this diorite intruding Snowshoe stratigraphy may be the result of extension during the formation of this oceanic basin to the west.

Diorite to quartz diorite of calc-alkaline affinities along Mt. Barker is more problematic. These arc-related igneous rocks have no correlatives of similar age in the immediate vicinity. Mafic arc volcanism of the same age has been shown to form basement to Quesnellia and comprises rocks of the Harper Ranch Subterrane south of the map area. It is conceivable that these are also present immediately west of here, but are buried beneath the thick cover of Mesozoic arc volcanics and sediments of Quesnellia. If this linkage is correct, then the coeval nature of diorite of both MORB and calc-alkaline nature is consistent with a model in which oceanic rocks of the Slide Mountain ocean represent a marginal basin produced by extension behind a west facing arc. More significantly, the presence of these calc-alkaline intrusions within westernmost pericratonic rocks of the northern Kootenay Terrane indicates that, if this model is correct, this Late Paleozoic arc was positioned immediately west of the edge of Ancestral North America and that the width of the Slide Mountain ocean was quite narrow.

KINEMATICS AND RELATIVE AGES OF VEINS, FRACTURES, FAULTS AND FOLDS IN THRUST-RELATED STRUCTURES, SOUTHERN LIVINGSTONE RANGE, ALBERTA

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INTRODUCTION

Geologic fieldwork conducted this past summer in the southern Livingstone Range (Figure 1) continued with the main goal of this project to conduct geological mapping at a scale of 1:10000. Mapping in the summer of 2000 outlined the Green Creek fold pair, and focussed on mesoscale fabrics of the Green Creek anticline, which is a fault-propagation fold (Cooley et. al, 2000) exposed in Green Creek and Morin Creek (Figure 2). This past summer of 2001, additional mapping to the north of Morin Creek outlined similar structures exposed in the Caudron Creek valley (Figure 3). Fieldwork also involved documenting faults, minor folds, fractures and veins associated with the larger structures, and included extensive sampling of veins throughout the study area.

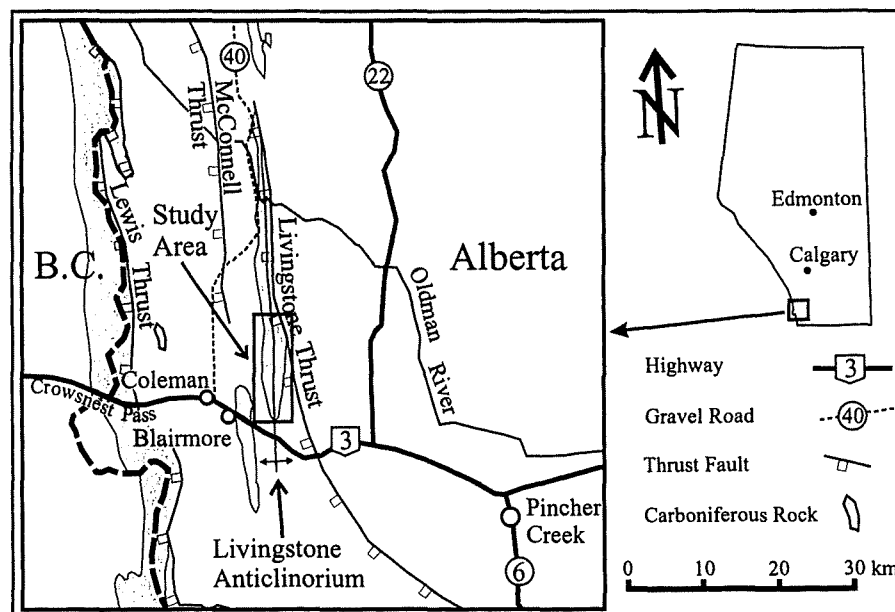


Figure 1. Location map of the study area at the southern end of the Livingstone Range, southern Alberta foothills. The Livingstone Range is an anticlinorium comprising Carboniferous and younger rocks in the hanging wall of the Livingstone Thrust.

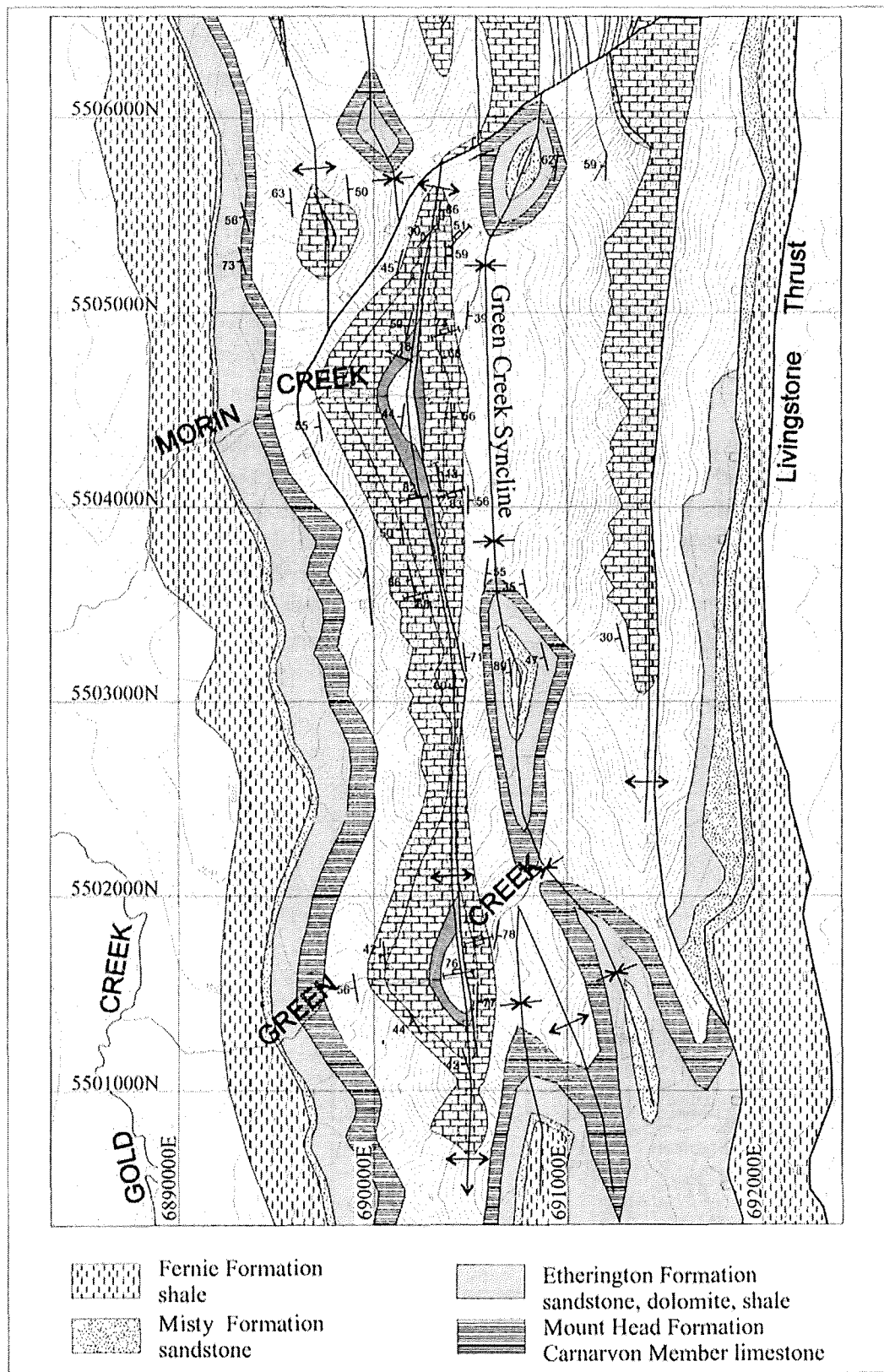


Figure 2. Southern half of the study area. The legend is shared with Figure 3.

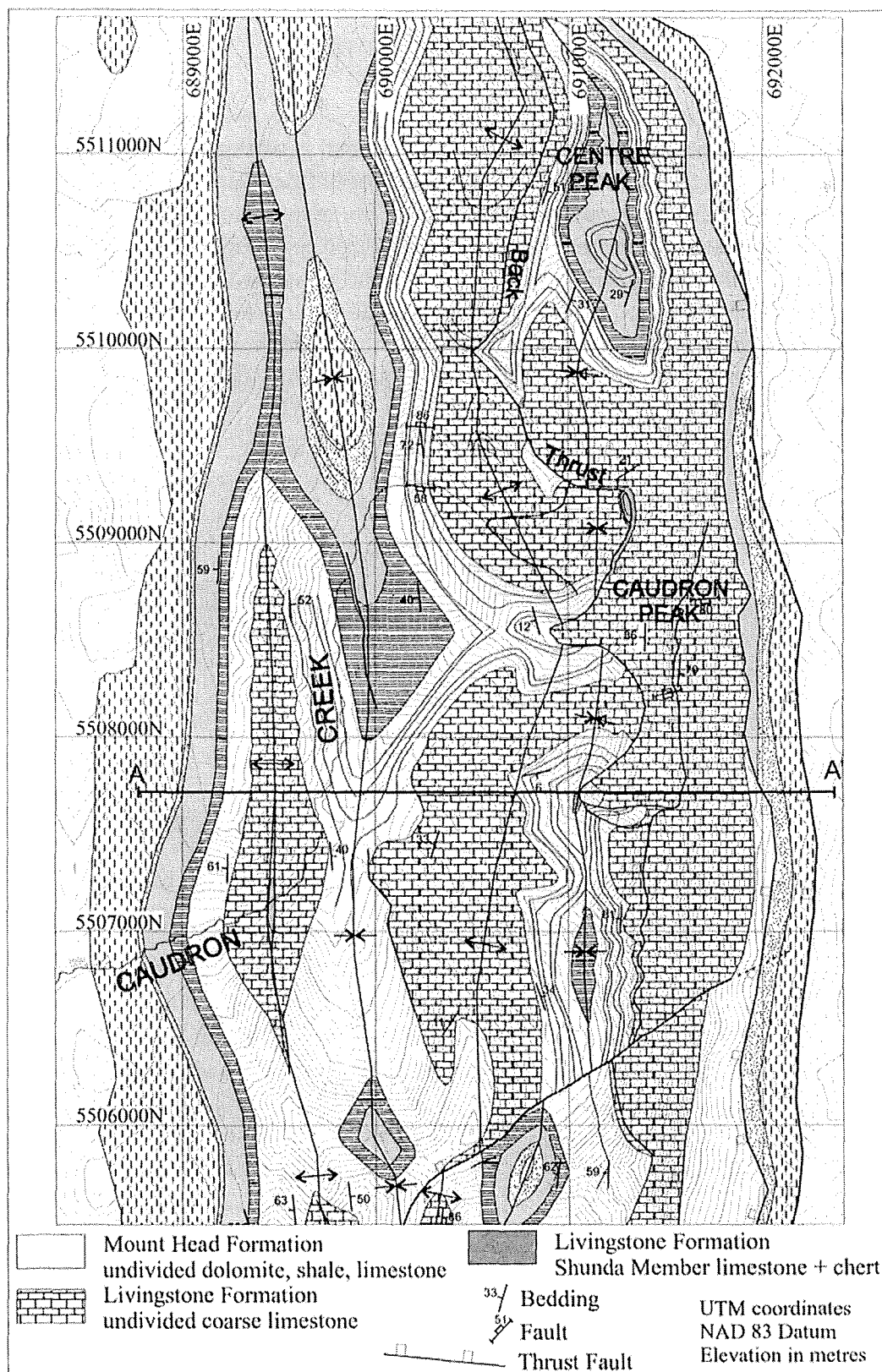


Figure 3. Northern half of the study area. The legend is shared with Figure 2.

CAUDRON CREEK STRUCTURES

Centre Peak Back Thrust

Detailed mapping has outlined the back thrust that is exposed south and west of Centre Peak as shown by Norris (1993). The surface trace of this "Centre Peak back thrust" continues to the south along the western slopes of the range where it links up with an underlying fore-thrust approximately 1 km south of Caudron peak (Figure 3). The two faults join where the fore-thrust crosses over the main ridge and is exposed on the east side of the range. Mount Head Formation strata in the hanging wall of the fore-thrust and the foot wall of the back-thrust form an eastward-tapering wedge of material which has indented to the east relative to the overlying and underlying rock (Figure 4). An initial interpretation of this back thrust implied that the back-thrust cut the fore-thrust, however, from further investigations it is assumed that the two thrusts were active at the same time.

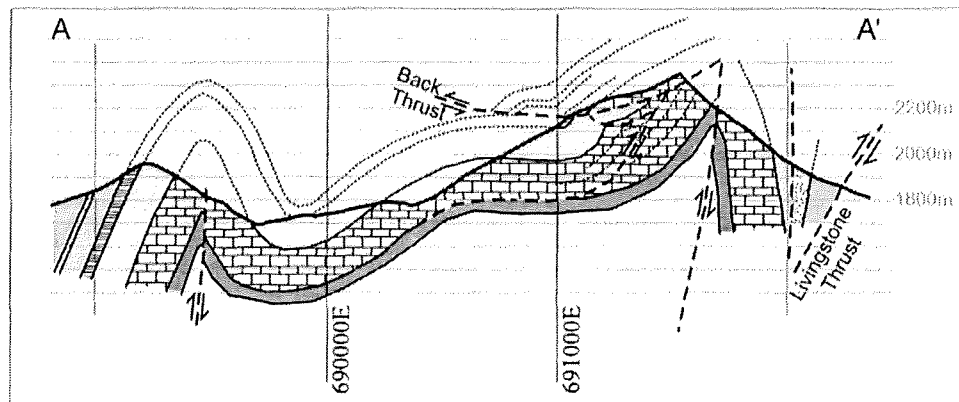


Figure 4. Cross section through the northern half of the study area. See Figure 3 for line of section. The legend is on Figure 2 and Figure 3.

Thrust Fault/Oblique Ramp/Tear Fault

A previously unmapped tear fault is present at the central part of the map area at 692000E, 5506200N (Figure 2). This east-northeast-trending structure has an apparent dextral displacement of approximately 250 metres as estimated from the average offsets of three major folds cut by this fault. To the east the surface trace of the fault is obscured on the east flank of the range. To the west the fault becomes an oblique ramp, with an anticline/syncline fold pair beginning in the hanging wall. South of the lateral ramp anticline the fault trace is assumed to link up with a bedding-parallel thrust fault with no apparent stratigraphic thickening.

Fault-Propagation Fold

As mentioned above, an anticline-syncline fold pair begins north of an oblique ramp fault just north of Morin Creek. This anticline is another fault-propagation fold that is similar to the Green Creek anticline. Where this fold is dissected by Caudron Creek, the core of the fold can be observed to consist of two steeply-dipping limbs with a thrust fault occupying the core of the fold (Figure 3 and Figure 4).

VEINS OBSERVED THROUGHOUT THE STUDY AREA

Samples of vein material have been collected from various positions along each type of structure present in the study area. In many cases, cross-cutting relationships have been observed between vein sets and between veins and stylolites and faults. The following observations are preliminary, and a more thorough analysis of vein relationships using thin sections and polished slabs is under way at present. Preliminary observations suggest that the occurrence of veins can be divided into the following general categories.

Major Anticline Cores

Veins within the core of the anticlines commonly consist of at least two sets of calcite veins; an early set which strikes east-west and is steeply dipping, and a second set which is not always present which consists of subhorizontal veins which usually cut the early vertical veins. A third set of veins is rarely observed which cuts the first two and lies parallel to the first set. This sequence of vein formation is not present everywhere along the core of the fold, in fact there are relatively few veins observed in the anticlines. The cross-cutting veins could represent a sequence of deformation involving early thrust-sense compression, followed by faulting and ramping of stratigraphy and later followed by tightening of the fold and compression in the hinge region.

Major Syncline Cores

In the major synclines there are early calcite veins that generally strike east-west and are steeply dipping. In two locales these veins are cut by late quartz+calcite veins and stylolites, which record a phase of east-west compression, north-south extension and vertical faulting. In total, veins containing quartz were observed in only 4 locations, 3 of which are associated with the Green Creek syncline, and one which is associated with the syncline that starts north of the oblique ramp discussed earlier. The source for the quartz is likely the silica-rich sediments of the Mount Head Formation and younger stratigraphy that are only present in outcrop in the synclines. One occurrence of calcite + fluorite veins was observed at 690315E, 5509560N. These veins are steeply-dipping, east-west-trending and appear to be associated with fracturing related to local faulting

Thrust Faults

Veins associated with thrust faults occur in the cores of drag folds and along the fault contacts themselves. Only calcite veins were found to be associated with these structures, commonly in association with stylolites.

Cross Faults

Cross faults generally trend east-west and are steeply-dipping. They tend to be the only structures that are commonly associated with breccia, and contain evidence for open space filling in the form of drusy calcite. These faults are interpreted to be late because they appear to cut veins and stylolites associated with thrust faults and folds, and because the calcite that fill the veins is usually undeformed and euhedral. Most veins associated with other structures contain calcite that has been deformed or recrystallized.

FLUID INCLUSION STUDIES AND ISOTOPE GEOCHEMISTRY

As discussed above, there are many large structures in the study area such as fault-propagation folds, thrust faults, and tear faults that have an apparent sequence of formation. Locally these large structures are associated with smaller structures in the form of veins that also show a sequence of formation and exhibit cross-cutting relationships. Additional information about the large structures may be obtained from fluid inclusion studies of veins and from the analysis of the carbon and oxygen isotopes of the veins and host rocks.

Fluid Inclusion Studies

Fluid inclusion studies are an indirect method of determining the density and composition of fluids trapped in inclusions in minerals by freezing and heating the mineral sample and observing the temperature at which phase changes occur. The type of fluid present in the inclusion will be a rough indicator of the permeability of the structures as they were forming. Saline formation fluids indicate structures were tight and migration of fluids was limited to locally derived formation water that travelled short distances. The presence of hydrocarbons will be an indication that oil or gas was migrating through the structures and was trapped in the veins. The presence of low salinity meteoric water indicates that the structures were breached and infiltrated by surface waters, a condition that was likely coincident with rapid cooling of the rocks that would have shut down the oil maturation process. Hydrocarbon traps may have also been breached and the oil or gas lost along these late structures that must have connected to the surface.

Perhaps the most important information gained from heating of fluid inclusions is the temperature of homogenization which indicates the temperature the rocks were at when the veins formed.

Isotope Geochemistry

The purpose of conducting isotope analyses of veins and host rocks of the study area is to determine the source of the fluids from which the veins formed. The possible sources of the fluids are;

- locally derived (isotopic compositions of vein and host rocks are similar)
- transported from elsewhere (isotopic compositions of vein and host rocks are different)
- meteoric (isotopic composition of vein resembles meteoric water, ie; isotopically light)
- magmatic (isotopic composition is enriched in ^{18}O , isotopically heavy)

The materials that this project will be analysing are CaCO_3 (limestone, calcite veins), CaMgCO_3 (dolomite rocks and veins) and SiO_2 (sandstone, quartz veins). The isotopes that will be analysed are oxygen (^{18}O , ^{16}O) and carbon (^{13}C , ^{12}C). Once the isotopic analyses are completed, the source of the vein-forming fluids can be assumed by comparing the isotopic composition of the veins and host rocks.

Isotopes can also be used to determine the temperature of vein formation if the vein contains at least two minerals that precipitated at the same time. If the two minerals are assumed to have been in equilibrium when they precipitated, a fractionation factor for O can be used to determine the temperature of formation. This can provide a check for the temperature information obtained from fluid inclusion analysis.

The geochemical results will be used to characterise the fluids that were migrating through the structure at each phase of deformation, and to determine how the fluid system evolved during deformation. This should provide insight into the hydrodynamics of formation fluids, hydrocarbon maturation, migration and trapping.

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DEEP REFRIGERATION OF AN EVOLVING THRUST AND FOLD BELT BY ENHANCED PENETRATION OF METEORIC WATER?: THE LEWIS THRUST SHEET, SOUTHERN CANADIAN ROCKY MOUNTAINS

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Refrigeration because of advective heat transport by deeply penetrating meteoric water is an integral part of the evolution of a foreland thrust and fold belt. Topographic relief generated during thrusting and folding drives large-scale deep penetration of cold meteoric water; and dilation due to the brittle deformation that accompanies thrusting and folding provides enhanced pervasive permeability. These considerations may resolve an apparent paradox in the thermal evolution of the Lewis thrust sheet and the rocks beneath it. There is an intuitive appeal to the notion that tectonic burial beneath a thick thrust sheet must lead to substantial heating of the rocks that are overridden; but this is contradicted in the case of the Lewis thrust sheet in a region where it was > 8 km thick. Coalification levels in Upper Cretaceous rocks beneath the Lewis thrust sheet are low (0.74 % VR); and the thermal history of rocks both within and below the thrust sheet, as deduced from apatite and zircon fission track analysis, requires rapid cooling at the time of thrusting. This paradox may be ascribed to unwarranted assumptions about the relative roles of conductive and advective heat transport during and after thrusting. Meso-scale tectonic fabrics record brittle dilation and associated solution transport and precipitation during thrusting and thrust-related folding. "Penetrative" deformation was accomplished mainly by fracturing and by slip on joints, bedding, and small faults. Reconnaissance regional stable-isotope geochemistry of vein-filling minerals indicates that deep penetration of meteoric water was widespread in the southeastern Canadian Cordillera (Nesbitt and Muehlenbachs, GSAB, v. 107, 1033-1050, 1995; CJES, v. 32, 1699-1719, 1995). Diachronous, foreland-migrating, Upper Cretaceous chemical remagnetization of Devonian and Mississippian carbonate rocks in the Front Ranges of the southern Canadian Rocky Mountains provides evidence of hot pore fluids flowing through these rocks prior, probably just prior, to their incorporation into the evolving foreland thrust and fold belt (Enkin et al., GSAB, v. 112, 929-942, 2000). This "hot flush" evidently was driven ahead of the deeply penetrating cold meteoric water that refrigerated the developing thrust sheets and made them even more brittle.

GEOMETRY, TIMING AND THE EVOLUTION OF THE BASE OF A CRYSTALLINE THRUST SHEET IN THE HINTERLAND OF THE SOUTHERN CANADIAN ROCKIES

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Deep crustal levels have been exhumed in extensional Cordilleran core complexes from beneath the magmatic arc that forms the hinterland of the southern Canadian Rocky Mountain thrust belt. This has exposed mid-crustal structures and permits us to propose links between the foreland thrust tectonics and deformation of the hinterland at depth. A number of models have been proposed for the nature of this geodynamic linkage (channel flow, transient shear zones, distributed shear, etc.) and, in particular, for the interpretation of thick, gently dipping shear zones in ductile domains 5 - 10 km thick. The southern (China Creek) culmination of the Valhalla core complex in southeast British Columbia in the vicinity of Castlegar and Trail between latitudes 49°10' N and 49°20' N exposes such a shear zone, the Gwillim Creek shear zone (GCSZ), first identified in the northern Valhalla complex. It lies near the base of a stack of layered and sheet like plutons of Middle Jurassic, Early, middle and Late Cretaceous, as well as of Paleocene age. Strain over a 2 to 5 km wide zone increases downward in the plutons to the gently domed top of the GCSZ from a high level where the plutons are unstrained and igneous textures and contact structures are well preserved to the upper level of the GCSZ proper where all igneous textures are destroyed and the plutonic rocks have been completely transformed into strongly foliated and lineated mylonitic gneiss. The downward increase in strain progresses irregularly but one can recognize three zones; an upper zone where strongly foliated and lineated layers are irregularly distributed through weakly foliated and lineated plutonic rock (zone 1), a middle zone, where the plutonic rocks are strongly lineated and foliated but where lenses on a metric and dekametric scale are present in which the original coarse plutonic textures and plutonic structures are preserved, albeit in a strained state, (zone 2), and zone 3, which we take to be the top of the Gwillim Creek shear zone, where plutonic remnants are absent or rare and where inclusions, dikes and veins are completely transposed.

On the southwest side of the China Creek culmination, the lower part of zone 2 incorporates the basal sheet of Mid-Jurassic tonalite whereas on the east side of the culmination, the basal sheet of the Mid-Jurassic tonalite is incorporated into zone 3 even though it is at a higher level in the plutonic sequence than on the west side. Similar relationships can be documented with a mid-Cretaceous sheeted pluton; intense, GCSZ-related (lower zone 2) shear is at a higher level in the plutonic layers on the east side of the culmination than on the west side.

The GCSZ rises at least 10 km from the culminations in northern Valhalla complex to the China Creek culmination, southward along the Eocene N-S axis of the complex. This rise cannot be explained by one simple lateral ramp

because the N-S axis of Valhalla complex is an Eocene structure overprinted on the concave-to-the-west arcuate structure of the Jurassic and Cretaceous southern Kootenay Arc. A portion of the southward rise of the GCSZ must actually represent a rise toward more easterly, external tectonic elements. China Creek culmination therefore exposes a structurally higher portion of the GCSZ, that is more "easterly", than the portion seen in the culminations of northern Valhalla. The eastward rise of the GCSZ is therefore seen at a local scale in the plutonic edifice south of Castlegar and at a more regional scale across Valhalla complex. In northern Valhalla a mid-Cretaceous granodiorite sheet is duplicated within the shear zone and, throughout Valhalla complex there are kinematic indicators, sheared veins, σ and δ structures, S-C fabrics, etc. that suggest top-to-the-east motion. These observations taken together with its geometry, indicate that the GCSZ is a ductile thrust zone that rises in its transport direction. It is several kilometres thick and, in southern Valhalla carries a pluton dominated crystalline sheet. The low strain and well established stratigraphy at high levels show that the gentle dips are original and not merely to be explained by transposition. In northern Valhalla, and further north, a lesser abundance of plutons above the GCSZ led to a different structural style relative to that of the southern Valhalla complex. Geochronology, crosscutting relations and fabric of metamorphic and plutonic rocks show an evolution of the shear zone that extends from ca. 80 Ma to ca. 60 Ma, and suggest a linkage with movement on a system of thrust faults in the Rocky Mountains related to the Lewis thrust. A general downward younging of ages of metamorphism and ductile strain has been proposed in other complexes, but in Valhalla the age pattern of intrusive sheets and ductile shear as well as of metamorphism is more complex. Eocene granite sheets and ductile extensional shear are superimposed on the compressional structures in the upper levels of the Valhalla complex.

The tectonic significance of 'thrusts' and transposition in high grade metamorphic rocks.

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Current tectonic models for high grade metamorphic rocks in regions of nappe development generally ignore, or fail to explain, the observed penetrative deformational structures. A commonly accepted model is simply a ductile version of the Rocky Mountain brittle thrusting, but this ignores most of the structures, and in fact is incompatible with some of the salient observations. For such a model to produce the observed folds it is essential that the sense of shear in the thrusts reverses repeatedly, at the scale of folding (typically microscopic to regional). This is contrary to the evidence which indicates a constant sense of shear across whole fold packages. If the thrust model and a constant sense of shear is assumed, the resultant fold geometry bears no resemblance to what is generally observed. The only model that is capable of explaining all of the observations involves crustal scale non coaxial flow, and modification of earlier folds by that flow. The early folds may be pre-flow, may be dragfolds produced by the flow, or most likely, both. In this model the structures that are generally interpreted as thrusts, are not thrusts, but are rotated and stretched pre-flow discontinuities, C- or Y-shears, and/or a combination of both.

If the non coaxial flow model is applied to the Monashee Complex, all observations can be explained, including: (a) intense transposition involving more than one generation of recumbent folds, (b) parallelism of the unconformity and all foliations above and below the unconformity, (c) numerous discontinuities at various scales, (d) a regionally recognisable stratigraphy which is generally right way up, (e) contacts that at outcrop scale are gradational, and (f) in rocks of the Selkirk allochthon, upward gradation into lower temperature rocks with upright folds. In the Monashee Complex and Selkirk allochthon there is regional scale reversal in sense of shear and this is readily explained by relating the non coaxial flow to extrusion (cf. channel flow, Beaumont et al., 2001).

The non coaxial flow model can result in palinspastic reconstructions that are very different to those produced by the ductile thrusting model. For example, the amount of crustal shortening predicted by the two models can be very different and is generally less for the non coaxial flow model.

Since high grade rocks with the same structural signature are very common, it is important to consider what tectonic environment they are indicative of. The necessary conditions of large magnitude, non coaxial flow are to be expected in Benioff zones, so that high temperature, high pressure rocks might be expected to have formed there. Large magnitude, non coaxial flow is also to be expected below mountain chains in zones of ductile detachment, extrusive flow, or channel flow. Finally, if convection occurs in the mantle, there has to be local detachment somewhere between slower moving crust and the mantle and this could conceivably give rise to horizontal shear zones within thermally- or fluid-weakened crust.

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When did deformation and peak of metamorphism take place in the Monashee Complex?

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Cordilleran deformation and related peak of metamorphism in rocks of the Monashee Complex and overlying rocks, up to the Selkirk and Cariboo Mountains, have been interpreted as younging towards deeper structural levels (Parrish, 1995, and references therein). Monazite, zircon, titanite and xenotime U-Pb dates within the Monashee Complex, that have been interpreted as metamorphic ages, range from ~78 Ma in the highest structural level to 49 Ma in the core of the complex (Gibson et al., 1999; Crowley and Parrish, 1999; Carr, 1992). Older metamorphic ages are reported away from the complex. They are as old as ~170 Ma in the Selkirk and Cariboo Mountains, and are supported by stratigraphic and fossil evidence (Parrish, 1995, and references therein).

Kuiper et al. (2001) reinterpret dates younger than ~75-70 Ma in the Monashee Complex, that were previously interpreted as representing upper amphibolite facies peak of metamorphism and crustal shortening, as being a result of crustal extension-related fluid activity. If this is correct, metamorphism must have occurred earlier, perhaps in the middle to Late Cretaceous (~100-90 until ~73 Ma) or Middle Jurassic (~175-160 Ma), as has been proposed for areas surrounding the Monashee Complex. Granitoids of these ages and of Early Mississippian (~360 Ma) age, are present in rocks that now surround the Monashee Complex, and that were at higher structural levels than the Monashee Complex before extension. (The exact geographic position of these rocks with respect to the Monashee Complex at this time is unknown; cf., Williams and Jiang, in review). This may indicate that deformation, metamorphism and melting occurred in the Monashee Complex at the times indicated by the ages of the granitoids. While deformation was ongoing, melts and fluids migrated up from the Monashee Complex, and crystallised at higher structural levels. This explains the absence of Palaeozoic and Mesozoic granitoids at the deepest structural levels (Monashee Complex). No Palaeozoic and Mesozoic U-Pb dates have been obtained in the Monashee Complex, perhaps because minerals used for U-Pb dating, that grew at an earlier time in the Monashee Complex (e.g., in the Proterozoic) partly or completely dissolved, and migrated up with these melts. Dissolution of these minerals in the Monashee Complex is supported by dissolution rims around Proterozoic zircon cores. These cores are overgrown by Eocene rims (in this study interpreted as being a result of extension-related fluids). $^{40}\text{Ar}/^{39}\text{Ar}$ hornblende samples with Palaeozoic and Mesozoic inverse isochron ages, have strongly disturbed spectra, which explains why no concordant Palaeozoic and Mesozoic $^{40}\text{Ar}/^{39}\text{Ar}$ ages (plateau ages) exist in the Monashee Complex.

Some suggestions of Palaeozoic and Mesozoic ages in the Monashee Complex may be given by U-Pb discordia intercepts and $^{207}\text{Pb}/^{206}\text{Pb}$ ages, and by $^{40}\text{Ar}/^{39}\text{Ar}$ hornblende inverse isochron ages. Zircon and monazite U-Pb discordia chords have Palaeozoic and Mesozoic upper and lower intercepts (with errors that are too large to derive ages with certainty) (Crowley, 1997, 1999; Crowley and Parrish, 1999). Monazite $^{207}\text{Pb}/^{206}\text{Pb}$ ages are as old as ~300 Ma (Kuiper, unpublished data). The lack of concordant U-Pb ages is not surprising if melts originated at these times from the Monashee Complex and moved up, carrying the elements for the minerals used for U-Pb dating, as explained above. Inverse isochron ages of hornblende with disturbed $^{40}\text{Ar}/^{39}\text{Ar}$ spectra also point towards Palaeozoic and Mesozoic ages (Spark, 2001; Vanderhaeghe, 1997; Sanborn, 1996). However these hornblende crystals did not remain closed for argon, due to heating, reheating, or interaction with fluids. $^{40}\text{Ar}/^{39}\text{Ar}$ and U-Pb ages are obscured by Late Tertiary-Early Eocene meteoric-hydrothermal activity.

Thus, although evidence is obscured by melting, heating and late fluid/rock interaction, U-Pb and $^{40}\text{Ar}/^{39}\text{Ar}$ dates in the Monashee Complex do seem to hint at deformation and metamorphism sometime in the Palaeozoic and/or Mesozoic. A model in which this interpretation is shown to be consistent with the regional geology is presented in Kuiper et al. (2002).

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**Evidence for orogen-oblique non-coaxial flow in the Thor-Odin culmination,
Monashee complex, BC.**

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Detailed mapping and field observations in the northeastern Thor-Odin culmination are used to interpret the strain history of the Monashee complex. It is suggested that an episode of NNE directed, non-coaxial strain may have occurred during Cordilleran orogenesis.

A non-coaxial strain path is interpreted on the basis of transposition, modification of upright folds into asymmetric north verging folds, drag folding and re-orientation of fold axes. Garnet inclusion trails are also consistent with top to the NNE non-coaxial flow.

Rocks in the Monashee complex underwent syn-tectonic amphibolite facies metamorphism. These relatively high temperature conditions promoted ductile deformation processes.

Although finite strain is path independent, it is demonstrated that the observed fold overprinting relationship would require a contrived explanation if the strain path was coaxial. The lack of well developed sheath folds suggests that the true strain history was one of triclinic general non-coaxial flow.

Geophysical evidence and the NW-SE trend of the foreland belt suggests that Cordilleran orogenesis was driven by northeast directed convergence. In the northeastern Thor-Odin culmination folds plunge between W and S, oblique to the trend of the foreland belt. The orientation of these structures can be explained by NNE directed non-coaxial flow (drag folding model). This strain path requires that deformation at depth (Monashee complex) not be fully partitioned relative to the bulk strain imposed by convergence. Furthermore it implies that there is no direct link between foreland folding and thrusting and ductile deformation in the Monashee complex.

New Constraints on the Age and Distribution of the Chase Quartzite, Vernon (82L) and Lardeau (82K) map areas: Regional Implications

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At the 2001 SNORCLE meeting, Thompson and others documented the existence of a distinctive, calcareous quartzite marker succession. The belt of outcrops is semi continuous from the town of Vernon east to Upper Arrow Lake, interrupted by a series of north-south trending, steep dipping, normal faults.

The age of the quartzite marker succession was inferred to be Upper Paleozoic based on a correlation with similar calcareous quartzite mapped on the eastern border of the Kuskanax batholith at Tenderfoot Lake (Read 1973).

Additional field work has confirmed that the Chase Quartzite, a relatively thick succession of calcareous quartzite that occurs 60 km northwest of Vernon at Little Shuswap Lake is laterally continuous with the calcareous quartzite marker unit. In other words, a relatively thin succession of calcareous quartzite, indicative of shallow water platform deposition, spanned the entire Vernon map area and at least the western extremities of the Lardeau 1:250,000 scale map area.

The age of the Chase Quartzite is Devonian. Three detrital zircon analyses gave concordant ²⁰⁷Pb/²⁰⁶Pb ages of 405±13, 424±13 and 1008±2 Ma (L. Heaman, unpublished data) reflecting the crystallization ages of the grains and constraining the maximum depositional age of the quartzite to latest Silurian. At the town of Chase, the quartzite, together with

pelitic schist (Silver Creek Formation) overlying it, are cut by latest Devonian granodiorite (L. Heaman, unpublished data). On the basis of these relationships, we suggest the Chase Quartzite and Silver Creek Formations are of Devonian age.

This age assignment helps constrain the age of the Tsalkom, Sicamous and Eagle Bay formations which overlie the Chase Quartzite near the town of Sicamous and occur up-section of it between Little Shuswap Lake and Adams Lake. The base of the Tsalkom Formation, a thick lens of mafic volcanic rocks, is interlayered with black, carbonaceous meta-argillite which is cut by Devonian granodiorite, and intruded by a felsic sill which has yielded an earliest Carboniferous U-Pb age (R. Freidman, unpublished data). Therefore, the Tsalkom Formation is no younger than earliest Carboniferous. The overlying Sicamous Formation overlies the Tsalkom Formation stratigraphically and is presumably Devonian or earliest Carboniferous as well. Age of the Eagle Bay Formation (in Vernon map area) is less constrained; like the successions discussed above, it may be as old as Devonian; however a Carboniferous or Permian age cannot be ruled out.

Rocks that underlie the Chase Quartzite appear to be Mesoproterozoic age, a relationship that has important regional implications. South of the town Sicamous on either side of the northern Okanagan Valley, Chase Quartzite overlies peraluminous

paragneiss and schist (Shuswap assemblage of Johnson, 1990) that continue 50 km farther east to Three Valley Gap where Parkinson (1992) obtained a 1.57 Ga upper intercept age from zircons in amphibolite. Age of the enclosing paragneiss may be older if the amphibolite is a metamorphosed sill rather than a volcanic flow. The paragneiss-schist assemblage contains a distinctive marble-calcsilicate-amphibolitic schist succession which hosts stratiform sulfide mineralization, notably at the Kingfisher and Big Ledge properties. The age of the sulfide-rich layers is interpreted to be Paleoproterozoic (1.7 to 1.85 Ga; Hoy, 2002). The marble-calcsilicate-amphibolite extends to the western side of the north Okanagan Valley at Mara Lake where it forms prominent cliffs. We conclude that the Devonian Chase Quartzite overlies Meso- or Paleo-proterozoic paragneiss and schist.

Farther south, along the Okanagan Valley, between the towns of Armstrong and Vernon, a thick succession of Chase Quartzite and underlying Shuswap assemblage schist and paragneiss has been mapped from Fortune Creek along the eastern slopes of the Okanagan Valley to the Kelowna Airport. The marble-calcsilicate-amphibolite marker present at Mara Lake reappears along the eastern shore of Kalamalka Lake. The relationship of Devonian Chase Quartzite on Proterozoic paragneiss and schist is consistent.

Our mapping has shown that the same rock successions exist at several places on both sides of the northern Okanagan Valley, making it unlikely that the valley contains a crustal-scale detachment at least from Kalamalka Lake to Sicamous.

Similarly, the presence of a marker unit on either side of the Upper Arrow Lake segment of the Columbia River Fault suggests little offset there. Chase Quartzite occurs both along the eastern side of the lake and to the west, near Arrow Park Lake. If the Fault is interpreted as a crustal-scale detachment having 10s of kilometers of down to the east

displacement (the amount required to offset the Monashee Décollement to depths consistent with westward projection of the basal décollement to the Rocky Mountain Fold and Thrust Belt), then the distribution of the same unit, at the same level, in both the hanging wall and the footwall, is difficult to reconcile with the large displacement interpretation.

Farther north, at Revelstoke, we reexamined the succession of schists, quartzites and marbles underlying the Standfast Creek Fault. The footwall succession is typical of the Silver Creek Formation mapped farther west in Vernon map area, leading us to suggest that the Standfast Creek Fault is an east-over-west thrust that places Lower Paleozoic Hamill, Badshot and Lardeau strata onto Devonian Silver Creek Formation.

It appears there was a significant westward change in basin configuration during the Proterozoic and Lower Paleozoic along an axis that now coincides with the trace of Upper Arrow Lake. East of the lake, is a thick Proterozoic (Belt-Purcell, Windermere) and Lower Paleozoic (Hamill, Badshot, Lardeau) succession; west of the Lake, essentially all of this succession is not present. One interpretation is that a western “high” consisting of Paleoproterozoic “basement” (e.g. the 1.8-2.2 Ga orthogneisses in the core of the Monashee Mountains), extended outboard of the Cordilleran continental terrace wedge.

Crustal cross section interpretations have linked the large offsets interpreted within inclined reflective zones in Southern Cordillera Lithoprobe lines 7 through 11, with deformation in cover rocks. However, our experience mapping the Chase Quartzite from Little Shuswap Lake to Upper Arrow Lake, a distance of nearly 120 Km across the strike of the Cordillera, suggests there are no large overlap structures at the surface. There appears to be a complete discordance in style and magnitude of structures between subsurface and surface (0 – 2 seconds depth on reflection data). One explanation is that the large

structures inferred from subsurface reflections predate evolution of the Cordilleran continental terrace wedge (Belt-Purcell and younger successions). The presence of a relatively thick and competent crustal block overlain by a thin veneer of sedimentary rocks, west of the thick basinal deposits characteristic of the continental terrace wedge, would create boundary conditions that could explain the abrupt change in structural style between the intensely folded and faulted Kootenay Arc, and the relatively little deformed and shortened supracrustal rocks exposed across the Vernon map area and farther west: A rigid crust, carrying supracrustal cover, appears to have plowed into the sedimentary basin east of it, creating a "bow wave" of folds and thrusts. Our observations support the wedge-shaped geometry of seismic reflections recorded at the western margin of the Selkirk Mountains near Revelstoke (Cook, 1986), and the crustal wedging model proposed by Price (1986).

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Evidence of Mesoproterozoic crust in the southern "Quesnel terrane": Extending the ancient Pacific Margin outboard

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Introduction

The Nicola horst is a Cenozoic uplift of amphibolite-grade metamorphic rocks intruded by a range of plutonic bodies in the hinterland of the Cordilleran orogen in southern British Columbia. The horst is surrounded by low-grade volcanic and volcanoclastic strata of the late Triassic Nicola Group (Fig. 1) which have been included in the "Quesnel terrane" and interpreted as an intra-oceanic arc built on a Devonian to Permian primitive basement (e.g., Monger et al. 1991). However, recent geological mapping to the west of the Okanagan Valley near Spa Creek, where the Quesnel terrane had been inferred from Lithoprobe crustal seismic data to be up to 20 km thick (e.g., Clowes et al. 1998), has led to the suggestion that a shallow Proterozoic cratonic basement locally underlies the Nicola Group (Erdmer et al. 2001).

In the Nicola horst, metamorphic rocks appeared from reconnaissance reports to be lithologically similar to those near Spa Creek, which offered an opportunity to test the hypothesis that evolved continental crust extends farther outboard, and the related hypothesis that Quesnel terrane strata form a thin stratigraphic succession on the ancient margin rather than a thick allochthonous crustal element. As part of the Ancient Pacific Margin NATMAP (National Geoscience Mapping Program) project of the Geological Survey of Canada, we studied the distribution, age and history of metamorphic rocks in the Nicola horst to elucidate their relations with the surrounding Nicola Group and test proposed regional correlations. We discuss the results of geological mapping, isotopic analysis, geochronology and metamorphic studies near Bob Lake, approximately 40 km south of Kamloops, British Columbia.

Results and discussion

A pericratonic succession in part of late Paleozoic age, the Bob Lake assemblage, includes felsic to intermediate metavolcanoclastic rocks, metaconglomerate, porphyroblastic pelitic schist, phyllite and other rock types. From U-Pb zircon analysis, a felsic metaporphry cobble in metaconglomerate is 1.04 Ga old. The oldest detrital zircons in metaconglomerate and schist are also near 1 Ga. The depositional age of the Bob Lake assemblage is unclear. The assemblage was intruded by 230 Ma tonalite, 219 Ma diorite and 64 Ma leucogranite, and includes 161

Ma felsic porphyry and 157 Ma rhyodacite (Fig. 2). Amphibolite-grade metamorphism and ductile flattening and stretching affected all of these rocks, but not crosscutting Paleocene granite and granodiorite. Ductile stretching lineation plunges moderately west, with overall equivocal shear sense.

The age of the Bob Lake assemblage is unclear. There are at least three samples (PE99-54, PE99-55 and PE01-2B) that give indications of Jurassic detrital zircons, and igneous rocks within the Bob Lake assemblage have yielded Jurassic U-Pb zircon dates: quartz-feldspar porphyry (sample PE98-27A) crystallized at 161 Ma, fragmental rhyodacite porphyry (sample JM761A) crystallized at 157 Ma, and a date of 158.3 ± 0.6 Ma was obtained for meta-dacite porphyry (Moore et al. 2000). However, because contact relations are ambiguous owing to poor exposure and penetrative strain, it is unclear whether the igneous rocks are intrusive, or extrusive and syn-depositional, or extrusive and post-depositional with respect to the rest of the Bob Lake assemblage.

The southernmost exposed part of the assemblage is intruded by the 230 Ma Clark Creek pluton, further indicating the need for caution when interpreting the Jurassic dates. An explanation of the black coating on Jurassic zircons is that it could result from passage through or incorporation of the graphitic pelite in the Bob Lake assemblage. Middle Jurassic zircons can be accommodated in the Bob Lake assemblage, if 1) the igneous clast hosting black-coated zircons (sample PE99-54) is part of a dismembered dyke rather than a detrital clast, 2) detrital black-coated zircons in metaconglomerate (sample PE99-55) are derived from such dykes, and 3) the dated Jurassic igneous units (samples PE98-27A and JM761A, and the 158 Ma metaporphry of Moore et al. 2000) are intrusive or later extrusive bodies. If these interpretations are correct, the Jurassic zircons do not constrain the depositional age, and an original stratigraphic contact between the Bob Lake assemblage and the Nicola Group is possible, with the Late Triassic Nicola Group overlying older rocks of the Bob Lake assemblage. Alternatively, if the Jurassic zircons place a limit on the maximum age of the Bob Lake assemblage and the 230 Ma Clark Creek 95

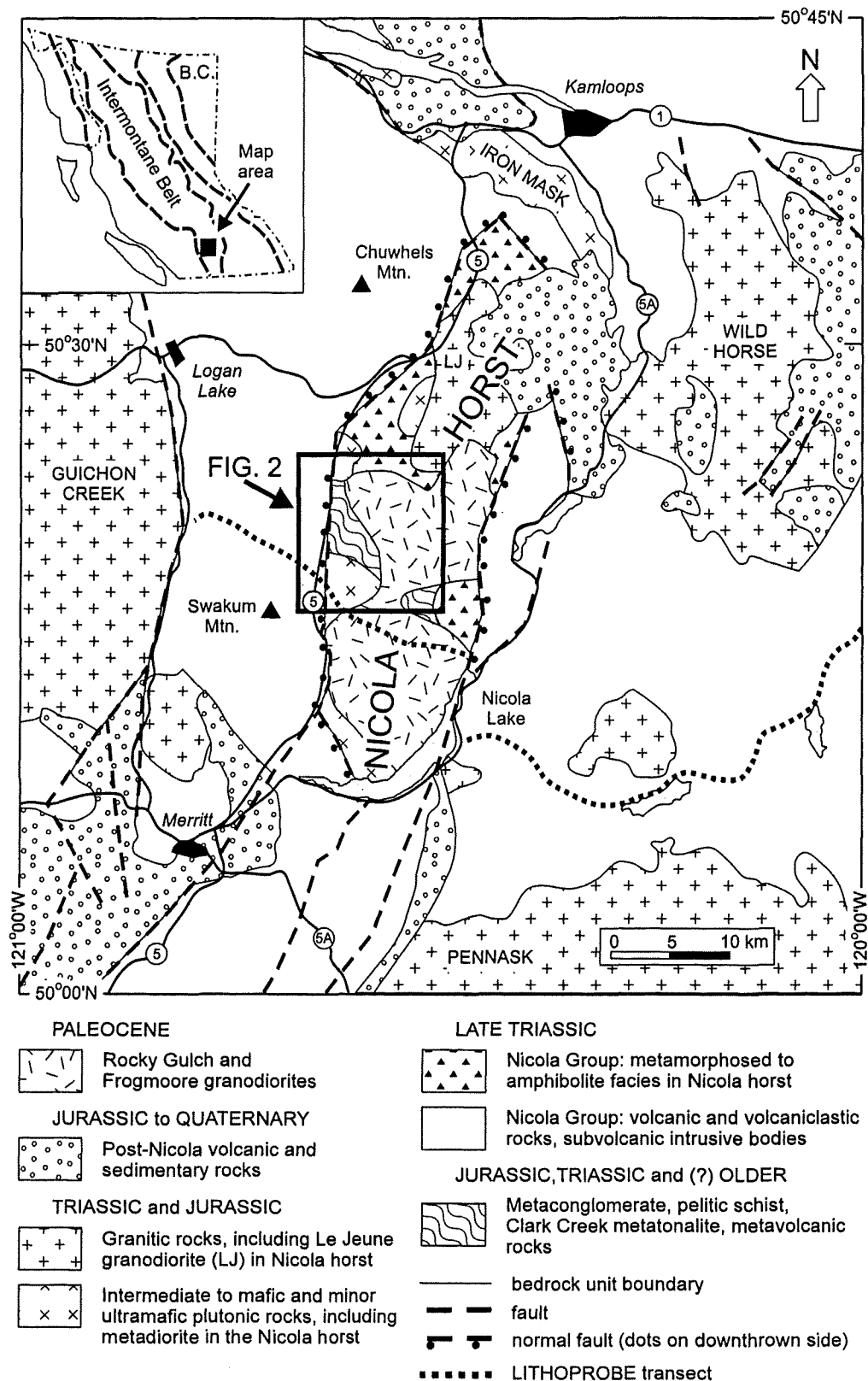
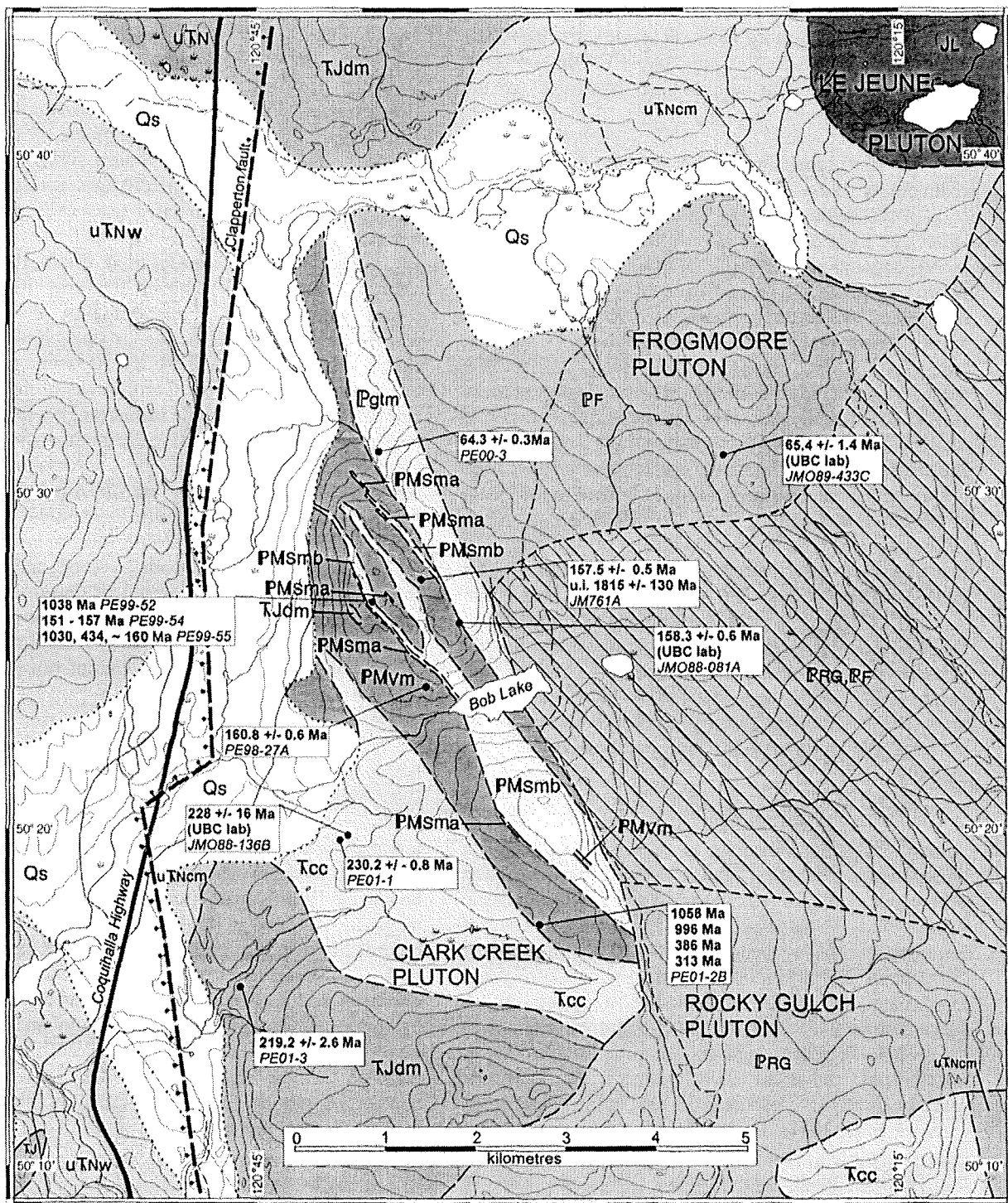


Figure 1. Location of the study area within the Cordillera of southern British Columbia, with topographic features and general geology of the Nicola horst and surrounding region. Major plutons are labeled and highway numbers shown.



QUATERNARY PLEISTOCENE AND RECENT

Qs Glacial, fluvioglacial, fluvial and lacustrine deposits; colluvium, landslide deposits

STRATIFIED ROCKS

LATE TRIASSIC

NICOLA GROUP

uTNw Western volcanic facies: mafic to felsic, plagioclase- and augite-phyric flows, pyroclastic and epiclastic breccias, tuff, wacke, limestone, limestone conglomerate

uTN Mafic and intermediate volcanic and volcanoclastic rocks, undivided

METAMORPHOSED NICOLA GROUP (?)

uTNcm Upper greenschist - mid amphibolite facies metavolcanic and metasedimentary rocks, interpreted to be metamorphosed Nicola Group; includes amphibolite, and biotite, biotite - hornblende and hornblende schist.

PALEOZOIC(?) or MESOZOIC

BOB LAKE ASSEMBLAGE

PMvm Intermediate and felsic fragmental biotite - plagioclase bearing metavolcanic (including ignimbritic?) rocks

PMSm Metasedimentary rocks, a: dark gray porphyroblastic pelite schist, phyllite, metasilstone, b: chert- and quartzite clast metaconglomerate; includes minor quartzite and quartz-rich schist

INTRUSIVE ROCKS

PALEOCENE

PRG ROCKY GULCH granite: hornblende - biotite and biotite tonalite, granodiorite and granite with K - feldspar megacrysts

PF FROGMOORE granodiorite: foliated hornblende - biotite and biotite quartz diorite, tonalite and granodiorite

Pgtm Foliated and lineated fine-grained leucogranite

JURASSIC

JL LE JEUNE metagranodiorite: foliated hornblende - biotite and biotite meta - quartz diorite, granodiorite and granite with K - feldspar megacrysts

TRIASSIC AND JURASSIC

TRIASSIC

TJ Hornblende diorite

TRIASSIC

Tjdm Hornblende metadiorite

TRIASSIC

Tcc CLARK CREEK metatonalite: biotite leucometatonalite

Lithologic contact: defined,

approximate, assumed

Fault: approximate, assumed;

solid circles on downthrown side

Topographic contour (100-ft interval)

Figure 2. Geologic map of the Bob Lake area, showing location of analyzed and dated samples. Samples JMO89-433C and JMO88-081A, labeled "UBC lab", are dates reported by Moore et al. (2000). The result for sample JMO88-136B labeled "UBC lab" is from R. M. Friedman (personal communication 2000).

intrudes an unrecognized older succession, the Bob Lake assemblage is younger than the Nicola Group and could be, for example, a metamorphosed equivalent of the Ashcroft Formation.

The unmetamorphosed and dominantly clastic sedimentary rocks of the Ashcroft Formation, which contain fossils as young as Callovian (McMillan 1974; upper age limit of ~ 157 Ma on time scale of Okulitch 1999), depositionally overlie Nicola Group rocks to the north, west and south of the Nicola horst (see Monger and McMillan 1989, Moore and Pettipas 1990a, b) in scattered small basins that define a middle Jurassic regional topographic surface close to the present land surface. This datum precludes any significant differential uplift of rocks surrounding the Nicola horst since the Middle Jurassic. If the Bob Lake assemblage is a metamorphosed part of the Ashcroft Formation, the area of crust that could have been deeply buried after Callovian time is limited to the extent of the horst. In addition, Proterozoic basement near Bob Lake would have been exposed during the deposition of the Ashcroft Formation; however, although no geochronological data are available for clasts from Ashcroft conglomerate, clasts readily match rock types of the Nicola Group, of crosscutting plutons such as the late Triassic Guichon Creek batholith, and of the Cache Creek Group farther to the west (see Travers 1978) and there is no record of other significant sources. For these reasons, the correlation with the Ashcroft Formation is considered unlikely.

Part of the Bob Lake assemblage could instead correlate with the Devonian to Permian Harper Ranch Group, which includes argillite, siltstone, sandstone, chert pebble conglomerate and felsic to mafic volcanoclastic rocks underlying the Nicola Group to the north and east of Kamloops (see Monger and McMillan 1989). The regional character of the Harper Ranch Group is poorly known and more data are needed to test such a correlation. A succession with the same relative stratigraphic position as the Bob Lake assemblage and sharing its lithological, metamorphic and strain characteristics is the Spa Creek assemblage exposed 70 km to the east of the Nicola horst (Erdmer et al. 2001). It includes quartzite- and granite-clast metaconglomerate, mica schist, and metamorphosed intermediate fragmental volcanic rocks as well as crystalline limestone and quartzite, and reflects a comparable pericratonic setting. Another potential correlative, exposed approximately 200 km to the east of the Bob Lake area, is the Mississippian Milford Group, a part of the ancient North American margin succession which has a component of siliciclastic rocks including chert- or quartz-pebble metaconglomerate and slate or mica phyllite, and tholeiitic greenstone and green tuffaceous metavolcanic and metavolcanoclastic rocks (Klepacki and Wheeler 1985).

The Bob Lake metaconglomerate is largely siliciclastic and has little matrix feldspar, indicating both

a continental source and a sedimentologically mature depositional setting. A continental source is also indicated by the ϵNd values of ~-1 to ~-4 for metaconglomerate and metapelite. From the middle Proterozoic T_{DM} ages of four of the Bob Lake assemblage samples, the continental source is interpreted to have had a significant crustal residence history. The Nd model ages overlap those of Devonian to Permian siliciclastic strata of the Cordilleran miogeocline in British Columbia, which have an average T_{DM} age of 1.54 Ga (see Garzione et al., 1997); thus, although Nd data can never uniquely identify the continental source of the sediments, the data are permissive of a North American miogeoclinal origin. In contrast, volcanic rocks of the Nicola Group outside the horst and the meta-Nicola Group rocks in the horst are characterized by Nd isotopic values ranging from ~+3 to ~+7, consistent with primitive arc magmatism. The Bob Lake, Spa Creek and Milford successions are all consistent with a magmatic arc of Late Paleozoic age built on a partly exposed cratonic block within a back-arc basin, where sediment composition and chemistry would reflect the proximity of both evolved and primitive sources. From its cobble size, the 1.04 Ga igneous clast in metaconglomerate (sample PE99-52) is inferred to have been deposited only a short distance from its source, and we propose that its Proterozoic crustal source was and remains nearby.

The 434 Ma, 386 Ma and 313 Ma zircon detrital grains in metaconglomerate and mica schist in the Bob Lake area (samples PE99-55 and PE01-2B) provide evidence of middle Paleozoic magmatism which is unknown in the miogeocline to the east. They may record activity in an outer margin arc or in exotic crust, or may be recycled grains from such settings. Middle Paleozoic magmatic activity at this latitude in the Cordillera is recorded by Devonian plutons west of the miogeocline, with the closest ones exposed about 75 km to the northeast and east of the Bob Lake area (Mt. Fowler batholith, 360 Ma, Okulitch 1985; unnamed 360 Ma pluton near Spa Creek, Erdmer et al. 2001). Thus, it is possible that the source of both the 1.0 Ga igneous component and the quartz-rich clastic component is not local and that the Bob Lake assemblage rocks are allochthonous.

The nearest present region of North America where continentally-derived quartz-rich clastic strata are interbedded with continental arc volcanic rocks, the locus of an inferred Jurassic arc graben, is more than 1,700 km to the south, in southeast California and southwest Arizona (e.g., the Cowhole Mountains, Wadsworth et al. 1995). That region is relatively close to exposures of Grenville-aged rocks in western Texas, New Mexico, southern Arizona and southeastern California, potential sources of 1.0 Ga detritus, although the youngest known magmatic rocks in those areas are ~ 1.12 Ga (see, e.g., Bickford et al. 2000). A potential source of material of more closely matching age is a suite of mid-Proterozoic

anorogenic plutons in Colorado and Wyoming (e.g., the 1.06-1.09 Ga Pikes Peak batholith, see Unruh et al. 1995).

But evidence also exists that links the Bob Lake rocks with nearby parts of ancient North America. The upper intercept date of 1815 ± 130 Ma from one zircon fraction in sample JM761A (Jurassic rhyodacite porphyry) is compatible with ages of both the buried and exposed parts of the North American crystalline basement to the east. A remarkably close match of the 1.04 Ga clast age in the Bob Lake assemblage is with a distinctive peak (at ~ 1037 Ma) of detrital zircon ages from the Ordovician Mt. Wilson Formation, a miogeoclinal quartzite exposed in southern British Columbia and southwestern Alberta. The peak has been attributed to two igneous pulses, at 1030 Ma and 1053 Ma, that would have generated unidentified plutons along the Cordilleran margin (see Gehrels and Ross 1998, and Gehrels 2000). The outer high of Proterozoic continental crust, i.e., the locally emergent block that yielded 1.0 Ga coarse detritus, was likely of significant extent. That inference is not substantially different from the "rifted continental block" proposals of Roback et al. (1994), Høy and Dunne (1997) and Dostal et al. (2001), in which the Quesnel volcanic arc was built on a fragment of continental crust that rifted away from the margin late in the Paleozoic. Our evidence neither supports nor invalidates a scenario of Late Paleozoic margin rifting but it does support the existence of exposed cratonic crust.

Because 64 Ma leucogranite cutting the Bob Lake assemblage is ductilely strained and the Rocky Gulch and Frogmoore plutons are not strained, the peak of synkinematic metamorphism in the horst is early Paleocene. The Nicola Group surrounding the horst is not ductilely strained and is therefore structurally decoupled from rocks within the horst. Meta-Nicola Group rocks and high-level or extrusive Jurassic porphyries inside the horst were buried to 12 to 15 km depth after 157 Ma and uplifted by the Paleocene; unpublished $^{40}\text{Ar}/^{39}\text{Ar}$ data indicate that unroofing of the Rocky Gulch and Frogmoore intrusions was substantially complete by 50 Ma (S. Ghosh, personal communication, 2001). Although regional metamorphism of Paleocene age was unknown until now in this part of the Cordillera, the time of metamorphism near Bob Lake is the same as the brief and intense Paleocene metamorphic peak that characterizes the Monashee complex, an uplifted area of North American cratonic basement approximately 150 km to the east (see, e.g., Parrish 1995).

As ductile mylonitic fabrics, recording either east- or west-directed simple shear, were imposed on 64 Ma leucogranite, the uplift of metamorphic rocks in the Nicola horst was rapid. To the west of Kamloops (see Fig. 1), there is evidence of easterly-directed thrusting and overturned structures in the Ashcroft Formation. Although the thrusting has been suggested to have taken place soon after Ashcroft deposition, i.e., in the mid-Jurassic (Travers 1978), its age is unknown. If there was a Paleocene thrust

belt, the Nicola "horst" could be a breached horse or duplex of deeper crustal slices within such a belt. However, there are no thrust faults in the Nicola Group with a strike length of 50 km or more, the present length of the Nicola horst.

Instead, we interpret the horst as a "window" through the brittle-ductile transition in a region where the transition was anomalously close to surface in Paleocene time. The Nicola Group outside the horst would represent brittle upper crust immediately above rocks in which ductile fabrics developed at shallow depth during regional extensional flow of the middle crust. In the southeastern part of the Canadian Cordillera, regional metamorphism spanned from the Late Cretaceous to the Eocene, with a thermal peak in the Paleocene, and was accompanied by significant crustal stretching (see, e.g., Parrish 1995). A hallmark of the resulting linear extension structures is their gentle plunge and general E-W trend, characteristics also seen near Bob Lake. If part of the pre-Triassic section in the Bob Lake area flowed ductilely out from beneath the Nicola Group culminating in the latest Cretaceous or Paleocene, it could have allowed 12-15 km-deep crust to rise to shallower depth. Such flow could have taken place in a low-viscosity crustal channel (see Beaumont et al. 2000a, b) and could account for the steep paleothermal gradient, young regional metamorphism, and extensional simple shear along moderately dipping surfaces that characterize the Bob Lake area. Final exhumation is interpreted to have taken place by late uplift, possibly with block tilting, along the brittle steep faults with only small throws that define the horst.

The Nicola Group has been included with other Mesozoic arc-related rocks in the "Quesnel terrane" (Monger et al. 1991), thought to have accreted in mid-Jurassic time with more than 200 km of structural overlap with respect to underlying rocks of the ancient margin of North America (e.g., Ghosh 1995). Proterozoic basement was known to be exposed beneath the eastern part of the Quesnel terrane (the Grand Forks gneiss and Vaseaux gneiss, see Armstrong et al. 1991), and zircon with a Paleoproterozoic inheritance age (~ 2.4 Ga) has been recovered from a 200 Ma granodiorite of the Quesnel terrane near Greenwood (see Dostal et al. 2001). The evidence from the Bob Lake area points to additional cratonic Proterozoic crust near the surface in the western part of the Quesnel terrane. A crustal thickness of ~ 20 km of the terrane, which would consist of oceanic arc rocks of mid-Paleozoic to Mesozoic age and include the Nicola Group, is increasingly incompatible with the evidence of cratonic rocks at or near the surface.

We suggest that Proterozoic cratonic rocks constitute the upper crust beneath Quesnel terrane rocks at this latitude and that the Nicola Group in many places is thin - locally at most a few kilometres, as inferred near Bob Lake. This does not preclude possibly thick arc rock successions in other parts, but the local thinness of the

Quesnel "terrane" imposes a mechanical constraint that precludes its emplacement as a nappe of crustal proportions. This supports the hypothesis that the Nicola arc succession was originally deposited on the ancient continental margin (see Unterschutz et al. in press), a margin which appears to have been wider than thought until now.

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Igneous and metamorphic history of the Metchosin Igneous Complex, southwestern Vancouver Island, British Columbia

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The Metchosin Igneous Complex (MIC) is a partial ophiolite which crops out on the southwestern tip of Vancouver Island and is considered part of the Crescent Terrane. The complex was formed circa 53 Ma and accreted to Vancouver Island shortly thereafter. It is composed of gabbroic bodies overlain respectively by sheeted dikes, pillow basalt and basalt flows.

In this study, detailed sampling of the volcanic sequence was conducted along four dip-parallel transects. Samples were examined for their petrography, metamorphic mineral chemistry, and bulk rock geochemistry. The purposes of the study is to: (1) Determine the tectonic environment for the formation of the MIC. (2) Determine the nature and extent of metamorphism in the MIC (3) Distinguish seafloor hydrothermal alteration of the basalt from later emplacement-related regional metamorphism. (4) Use these data to improve our understanding of the Cenozoic history of Vancouver Island.

The immobility of most elements despite lower greenschist facies metamorphism was demonstrated using the phenocryst assemblage in conjunction with Pearce element ratios. Only K and Rb exhibited substantial mobility. Trace and rare earth element patterns reveal a weak mantle plume signature, suggesting that the complex formed in a tectonic environment transitional between mid-ocean ridge and hot spot. Trace element patterns lie along a mixing line between N-MORB and OIB end-members and correlate the MIC with other portions of the Crescent Terrane. Petrographic and mineral chemical data reveal that the complex has been uniformly metamorphosed to lower greenschist facies with carbonate and zeolite present solely as late phases. The extent of groundmass alteration increases with proximity to intrusive bodies and major faults, especially the terrane-bounding Leech River Fault to the north. The peak metamorphic conditions and cooling ages of the MIC and the Pacific Rim Terrane are similar. Therefore the accretion of the MIC beneath the Pacific Rim Terrane is proposed as the cause of regional metamorphism in the complex. In light of these results, the MIC is interpreted to be a seamount or island formed on or near the Kula-Farallon ridge as a result of the proximity of the Yellowstone hotspot and subsequently metamorphosed during its accretion to North America.

The environment for the eruption of the upper sequence of sheet flows remains undetermined, although circumstantial evidence suggests that they were erupted subaerially. Further field work in the eastern portion of the complex may resolve this question more satisfactorily. Laser ablation ICP data and stable isotope data for metamorphic minerals may help to distinguish between seafloor hydrothermal alteration and emplacement-related metamorphism.

PROGRESS IN THE UNDERSTANDING OF THE INTERNAL AND EXTERNAL PARTS OF CONTRACTIONAL OROGENS FROM NUMERICAL MODEL EXPERIMENTS

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We report on progress in both mechanical and thermo-mechanical plane-strain, cross-sectional, arbitrary Lagrangian-Eulerian finite element model experiments. The experiments are designed to improve our understanding of large hot orogens and their foreland fold-and-thrust belts in general and with potential application to the Western Canadian Cordillera, in particular.

CRUSTAL SCALE EXPERIMENTS

At the crustal scale, we model a cordilleran-type orogen using a similar approach to that described in our 2001 SNORCLE report (Beaumont et al., Lithoprobe Report 79, p.112-170, see also Beaumont et al., *Nature*, v.414, p.738-742). In this instance, however, the basal velocity boundary conditions assume eastward dipping subduction of an oceanic plate beneath a continent with the subduction zone fixed in space. The continent moves westward and deforms as it is compressed against the subduction zone. This is believed to be the tectonic style of the Andean orogen, with South America moving westward and contracting against the anchored Nazca subduction zone (e.g. Pope and Willett, *Geology*, v.26, p.511-514, 1998) and is adopted here as a prototype for the Southern Canadian Cordillera. Material (sediment and allochthonous terranes) is detached from the oceanic plate and accreted/obducted to the western side of the orogen, overthrusting the continent. However, the main crustal thickening is a consequence of the continental contraction which creates an orogen from autochthonous crust, in this case including a large prism of continental margin sediments. When the radioactive heat production of the sedimentary prism is $2 \mu\text{W}/\text{m}^3$ and the continental velocity is 1 cm/yr the continental mid-crust reaches 700°C in approximately 30 Myr [after the start of the model] and a gravitationally driven mid-crustal channel flow is well established by 45 Myr. Although such models are much simpler than the evolution of the Southern Canadian Cordillera, they provide a perspective on the potential role of channel flows in creating paired distributed 'normal' and 'thrust' sense shears within the crust corresponding to the upper and lower limbs of the channel. Unlike the Himalaya, the channel, if it existed beneath the Southern Cordillera, does not appear to have been exhumed to the surface during convergence. However, subsequent crustal scale extension may have exposed the relict and overprinted channel in the interior of the orogen.

EXTERNAL THIN-SKINNED FOLD-AND-THRUST BELT EXPERIMENTS

At the scale of the sedimentary cover, we consider external fold-and-thrust belts, such as the Foothills Belt of Western Canada, to be tectonic wedges with the same underlying mechanics as those described by Chapple (*Geol.Soc.Am.Bull.*, v.89, p.1189-1198, 1978),

Stockmal (JGR, v.88, p.8271-8287, 1983), Dahlen (JGR, v.89, p.10125-10133, 1984) and others. The complex deformation pattern stems from the internal deformation of a wedge that is created from a layered stratigraphy in which material properties vary among layers. The wedge is also subject to syn-tectonic surface denudation and sedimentation. We illustrate some basic styles of deformation and their dependence on properties of the model layers. We also investigate the response to slope-dependent surface denudation and show the structural response of the wedge as it attempts to regain the critical state. Preliminary results suggest that denudation and the formation of foreland-vergent thrust duplexes are connected.

EXTERNAL THIN-SKINNED FOLD-AND-THRUST BELTS CONTAINING SALT

The basic structural style of tectonic wedges with weak (salt, shale) basal décollements and the tendency to form symmetric box folds, pop-ups and other structures is well understood in terms of the internal failure planes (e.g. Davis and Engelder, *Tectonophysics*, v.119, p.67-88, 1985). Here we present some preliminary results for the problem of fold-and-thrust belts forming above a finite thickness buoyant salt layer (e.g. the Zagros fold-thrust belt and Potwar Plateau/Salt Range Pakistan). Specifically, the experiments consider contractional inversion of a salt basin with Coulomb frictional overburden. They demonstrate the formation of a critical Coulomb wedge as the overburden is expelled from the basin, the intrusion of the salt into the wedge décollement, the adjustment of the tectonic wedge to the new weak, viscous, basal boundary condition above the salt, and the final failure of the wedge with synchronous piercing by one or more diapiric salt walls.

Examples of these results can be viewed on the Dalhousie Geodynamics website
<http://adder.ocean.dal.ca>

(See: Nature paper, ACPI Proposal-Appendix 2, and Thin-Skinned Fold and Thrust Belts 2000 and 2002.)

***ABSTRACTS of
POSTER
PRESENTATIONS***

Geology of the Johnson Lake Area, Southern British Columbia.

Sean L. Bailey¹, Suzanne Paradis², Stephen T. Johnston¹

Abstract

1: 20,000 scale geological mapping of the Eagle Bay Assemblage was conducted in the Johnson Lake area to document lithologies that host economically important base and precious metal deposits. The Eagle Bay Assemblage consists of polydeformed, greenschist metamorphic grade rocks derived from volcanic, intrusive and sedimentary protoliths, and is assigned to the pericratonic Kootenay Terrane of the Canadian Cordillera.

Devonian-Mississippian age felsic, intermediate, mafic volcanic/intrusive rocks, and siliciclastic rocks, are structurally overlain by Lower Paleozoic (and older?) mafic volcanic/volcaniclastic, limestone, and siliciclastic rocks. Age constraints include a 36064.6 Ma U-Pb zircon age on felsic volcanic rocks of unit 9, and correlation of a thick limestone with the archaeocyathid bearing, Early Cambrian Tschinakin Limestone. The limestone is intercalated with mafic volcanic/volcaniclastics rocks and siliciclastic rocks, thus likely constraining them to the Lower Paleozoic.

Geological units and structures trend northwest-southeast. The geometries are controlled predominantly by west verging folding and thrusting of Jurassic age. Folds are tight to isoclinal with overturned limbs, have synmetamorphic axial planar cleavage and developed synchronously with thrust faulting. Fault geometries can be explained by treating the thrusts as a series of ramps and flats. Thrust Flats trend northwest and dip moderately east concordant with stratigraphic layers. Ramps trend northsouth, dip steeply east and cut across stratigraphic layers.

These rocks host VMS (Rea, K7, Homestake) and quartz-carbonate vein silver (Samatosum) deposits that demonstrate the exploration potential of the area. The Homestake massive barite/ sulphide deposit is hosted within altered felsic volcanic rocks of unit 6 that continue along strike beyond the map area. The Rea and Samatosum deposits occur within a fault repeated altered argillite and quartz-wacke that is truncated to the south. However this horizon continues northwest of the map area and may host undiscovered massive sulphide deposits.

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Stratigraphy and structure of the Mount Copeland area, Monashee complex, BC: Preliminary results.

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Within the southern Omineca belt, high-grade metamorphic basement and cover rocks of the Monashee complex are exposed through a tectonic window in the Selkirk allochthon. Clastic, carbonate, and volcanic cover rocks of a passive rift margin sequence lie unconformably above Proterozoic migmatitic ortho- and paragneiss of the North American basement. The stratigraphy and structure of the southern end of Frenchman Cap dome, located in the northern part of the Monashee complex, has not been clearly established by previous workers. New stratigraphic and structural mapping indicates a need for reinterpretation of the stratigraphy and suggests a previously unrecognized three-dimensional structural geometry.

Two summers of detailed field mapping at 1:20,000 scale have been completed. The contact between Proterozoic basement gneisses and the sequence of unconformably overlying metasedimentary cover rocks is well exposed within the field area. The regional lithology (transposed stratigraphy) in ascending order is orthogneiss and migmatitic paragneiss of the crystalline basement, basal quartzite, micaceous schist and calc-silicate gneiss with minor quartzite, white marble, and calc-silicate gneiss. This sequence was observed at several localities, and repetitions are attributed to isoclinal folding (F1/F2).

Four phases of folding on the mesoscopic scale have been recognised. Type II and III interference patterns have also been observed on the mesoscopic scale and are inferred to be present on the macroscopic scale from map patterns recorded during this study, and mapping by previous workers. The dominant transposition foliation (S2) developed before and during the growth of high-grade metamorphic minerals. Isoclinal folds F1 and F2 are found within the plane of the transposed stratigraphy. Many of the structures have almost been obliterated by progressive deformation, and no primary structures have been observed within the area. Predominant structures are northwest southeast trending, which include kilometre-scale, close to tight northeasterly verging F3 folds. These F3 structures control map-scale geometry throughout the area. Recently identified gentle folds with a more east-west orientation are interpreted as fourth phase structures (F4).

Mapping by previous workers has suggested the existence of a major fault along the basement – cover boundary. Based on our new data this structure, previously called the “Bews Creek Fault” is interpreted to be an early ductile feature with only minor displacement. Local zones of high strain and truncations of lithologic units adjacent to the basement–cover boundary are most likely a result of high ductility contrast.

COMPARISON OF LITHOSPHERIC STRUCTURES ACROSS THE ALASKAN AND CANADIAN CORDILLERA: A TALE OF SUBDUCTION ACCRETION AND COLLISION

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During the past decade, results from seismic experiments across the Cordillera of Alaska and western Canada have provided unprecedented information on the subsurface structure of the crust and upper mantle. In particular, three transects – central Alaska, northern British Columbia and southern British Columbia – have been the focus of both seismic reflection and seismic refraction/wide-angle reflection profiles (Figure 1). Results from these studies enable a comparison of the similarities and differences in lithospheric structure along the northern North American Cordillera.

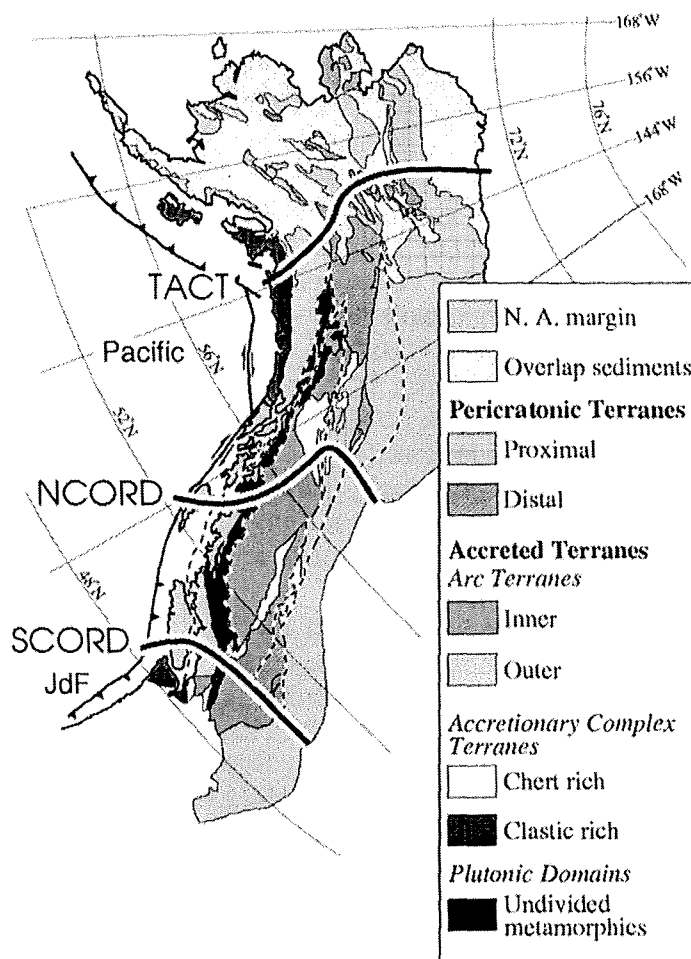


Figure 1. Generalized geological and terrane map of northwestern North America with the locations of the three transects along which interpretations of lithospheric structure from seismic reflection and refraction experiments have been derived. Abbreviations: TACT - Trans-Alaska Crustal Transect; NCORD - northern Canadian Cordillera transect (LITHOPROBE, ACCRETE, and others); SCORD - southern Canadian Cordillera transect (LITHOPROBE). JdF - Juan de Fuca plate.

In Alaska, the Trans-Alaska Crustal Transect (TACT) extends from the Aleutian trench across the Cenozoic–Cretaceous accretionary margin and the Late Jurassic–Cretaceous Brooks Range fold–thrust belt to the Early–Late Cretaceous North Slope passive margin. The two transects across the Canadian Cordillera profile lithospheric structure from the ocean basins to the North American cratonic core. In northern British Columbia, studies from the U.S. ACCRETE project and the Canadian LITHOPROBE SNORCLE Transect extend from the Pacific basin, across the Wrangellia and Alexander accreted terranes of the Cretaceous Insular Belt, the Coast Belt, Stikinia and Cache Creek accreted terranes of the Jurassic Intermontane Belt, the parautochthonous Omineca belt and further eastward to the Jurassic–Cretaceous fold–thrust belt built on the Proterozoic passive margin of the North American craton. In southern British Columbia, seismic results from the LITHOPROBE Southern Cordillera Transect provide extensive coverage from the Juan de Fuca plate across the Cascadia subduction zone, the Wrangellia accreted terrane of the Cretaceous Insular Belt, the Coast Belt, Quesnellia and other terranes of the Jurassic Intermontane Belt, the complex tectonic boundary of the Omineca Belt and further eastward to the Jurassic–Cretaceous fold–thrust belt.

Velocity–depth structure sections are compiled for each of the three transects. These are displayed with common distance and color scales for ease of comparison. Interpretations based on the original papers, including those from reflection images where available, are presented with the velocity–depth sections and combined to show the varying lithospheric structure along the Cordillera (Figure 2). The transects exhibit many similarities, but a number of substantial differences emphasize along–strike variations in terranes, orogen geometry and post–orogenic tectonics.

The accretionary margins of Alaska (TACT) and southern British Columbia (SCORD) have a number of similar tectonic elements with differing aspects; e.g., an active subducting plate, Cenozoic accretionary prism, Mesozoic accretionary prism (tectonic wedge in TACT, underplated in SCORD), and backstops to these prisms. In contrast, the northern Cordillera (NCORD) region no longer experiences subduction. The western boundary of North America is the Queen Charlotte dextral transform fault. While all transects include a part of the Wrangellia/Alexander accreted terranes, the extensive Coast Plutonic Complex (CPC) does not extend as far north as TACT. This Cretaceous–Paleogene suture is a prominent structure in the Canadian transects, exhibiting significant along–strike variation. Along NCORD, the CPC includes the associated Coast Shear Zone along its western edge; along SCORD, the structure equivalent to the Coast Shear Zone lies within the eastern CPC.

Crustal thicknesses are approximately 30–35 km beneath the accreted terranes in all three transects. Along SCORD, crustal thickness decreases from about 35 km to 30–32 km across the CPC and into the Insular Belt to the west (*Clowes et al., 1995*). In contrast, the NCORD transect shows significant thinning in the Insular Belt; crustal thickness decreases to 20–25 km west of the CPC (*Morozov et al., 1998*). Below the CPC, crustal thickness is ~31 km. To the west below Stikinia, this thickness increases to ~36 km (*Hammer et al., 2000*). In the interior of the Alaskan Cordillera, crustal thickness below the accreted terranes is about 30 km (*Beaudoin et al., 1994*). However, further to the north below the southern Brooks Range, much of the crust is detached by a crustal–scale, south–dipping tectonic wedge of North Slope lithosphere (the North Slope plate), above which uplift and deformation has occurred and below which mantle is depressed forming a 50 km thick crust (*Fuis et al., 1997*). The geometry of indentation of the North Slope plate is similar to that in other tectonic environments that are associated with subduction and continental collision. Specifically, there is considerable similarity with the collision of the Fort Simpson and Hottah terranes as revealed by the SNORCLE Line 1 reflection data (*Cook et al., 1999*). No similar structure is recognized in the Canadian Cordillera.

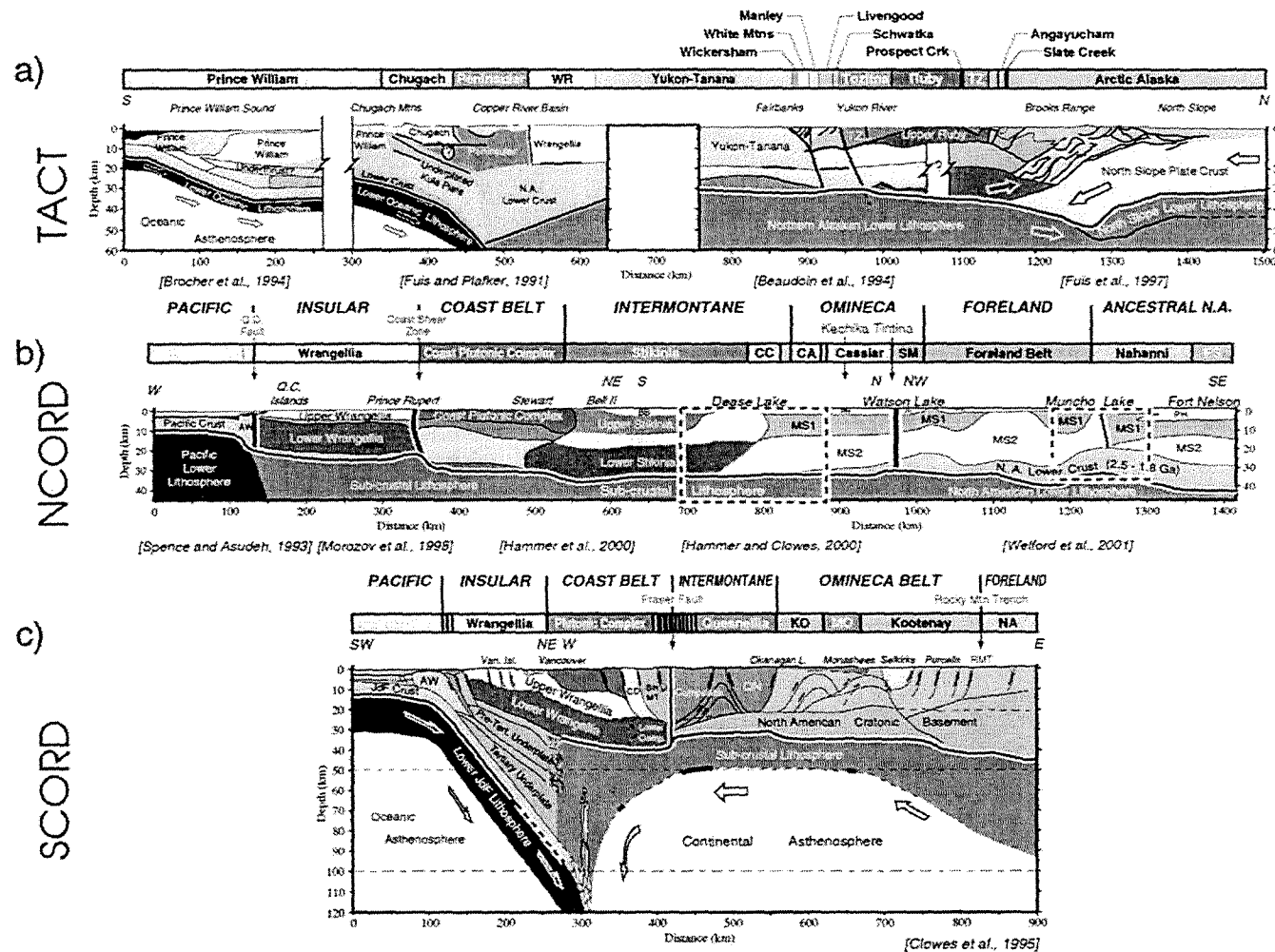


Figure 2. Schematic cross-sections (2.7:1 vertical exaggeration) based on results of trans-Cordillera transects. a) TACT. b) Northern Cordillera (Queen Charlotte traverse, ACCRETE, and SNORCLE transects). Dashed boxes described in Figure 3. c) Southern Cordillera transect. Short, solid bars at base of "Sub-crustal Lithosphere" identify locations of wide-angle reflections from the upper mantle. Abbreviations: AW - Accreted wedge, AX - Alexander, BB - Bowser Basin, CA - Cassiar, CC - Cache Creek, CD - Cadwallader, DY - Dorsey, FS - Fort Simpson, KO - Kootenay, m - undivided metamorphics, MO - Monashee, MS1 - Metasediments (1.2 to 0.8 Ga), MS2 - Metasediments (1.8 to 1.2 Ga), MT - Methow, OC - Olympic core (wedge rocks), PH - Phanerozoic sediments, QN - Quesnellia, SH - Shuksan, SM - Slide Mountain, ST - Sukinia, TZ - Tozitna, WR - Wrangellia.

In the interior of the NCORD transect, east of the CPC, interpretations indicate that the crust of the Stikinia accreted terrane (Intermontane Belt) extends to the Moho but the collage of accreted terranes east of Stikinia have only a limited depth extent (< 5 km) and are underlain by a layered wedge of Proterozoic metasedimentary rocks (Figure 3; *Snyder et al., 2002*). Crustal thickness remains relatively consistent at about 35 km, becoming more variable further east below the Foreland fold–thrust belt (*Welford et al., 2001*). In the interior of the SCORD transect, the crustal rocks of the accreted terranes also are detached from their subducted lithosphere and attached as flakes to the underlying Proterozoic cratonic rocks, but here they extend to as deep as 20 km. Crustal thickness also is relatively consistent at about 33 km, again becoming more variable further east below the fold–thrust belt (*Clowes et al., 1995*).

In all three transects, the unconformity or decoupling decollement that separates the interior accreted terranes from the underlying rocks dips westward (or southward for TACT and the indenting North Slope plate). The decollement extends from the front of the thrust–fold belt (Brooks Range for TACT, Foreland region for NCORD and SCORD) to the Moho along TACT and NCORD and to the lower crust along SCORD. Above the decollement, crustal–scale imbrication, antiforms and other deformation were generated by collisional tectonics. Along SCORD some of this deformation has been unroofed and exposed by Eocene extension but the surface rocks along NCORD and TACT mask much of the extensive crustal deformation inferred from the seismic results.

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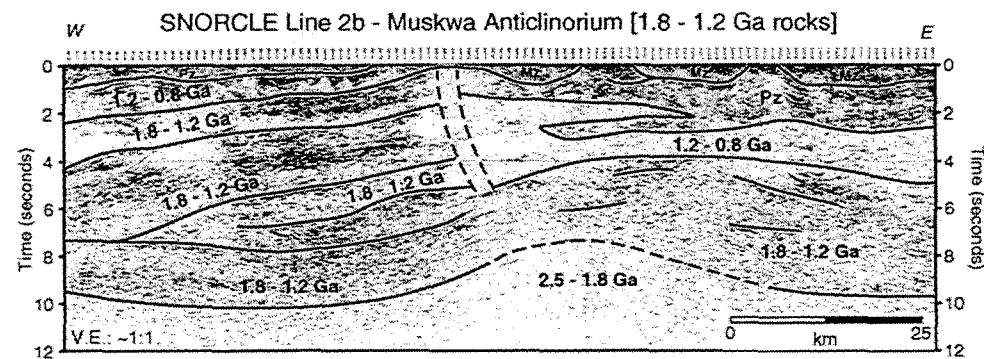


Figure 3a) SNORCLE Line 2b migrated reflection data showing correlation of reflections in Muncho Lake region with the Proterozoic Muskwa complex (1.8 - 1.2 Ga). Continuity of reflective sequence to the west provides basis for interpretation of Proterozoic metasedimentary rocks extending as a wedge to vicinity of Dease Lake (see Figure 2 and 3b). Abbreviations: Mz - Mesozoic, Pz - Paleozoic.

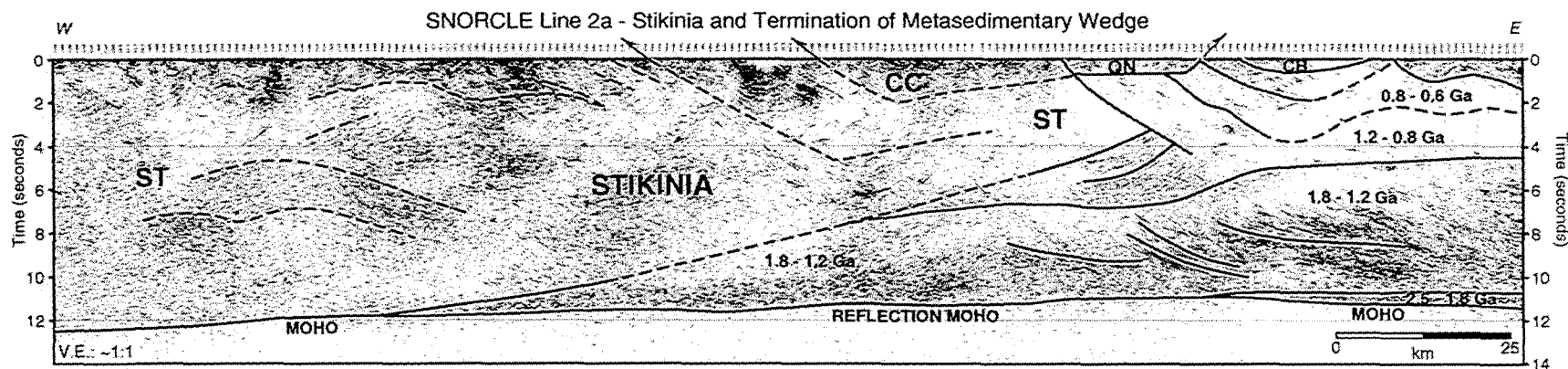


Figure 3b) SNORCLE Line 2a migrated reflection data showing termination of Proterozoic metasedimentary wedge overlain by the accreted terrane, Stikinia (ST), which has a full crustal section to the west. In contrast, other accreted terranes [Cache Creek (CC) and Quesnellia (QN)], are imaged as thin flakes of about 5 km thickness. Abbreviations: CB - Cassiar batholith, Pz - Paleozoic, - known faults.

Cross-section of the lithosphere from the Slave Province to Kula Plate constructed from seismic reflection data: The supracrustal crust

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Introduction

One of the primary objectives of the SNORCLE transect is to construct a cross section of the lithosphere across the North American plate from the Archean rocks of the Slave Province to the young rocks of the western Cordillera. Here, portions of each of the three regional reflection profiles and their initial interpretations are compiled into such a cross section by correlation from one to another along strike, assuming a total of ~350-400 km of dextral motion along the Tintina - Northern Rocky Mountain Trench fault zone. The resulting section provides a unique view of northwestern Canada following the contractional phase of the Cordillera in Late Cretaceous-Early Tertiary and prior to subsequent strike-slip motion on the Tintina fault.

Construction of the section: Strike projections

The SNORCLE seismic reflection program included acquisition of four major reflection profiles (1, 2a, 2b, and 3) between 1996 and 2000. Initial interpretations of these data for separate parts of the data set have been presented elsewhere (Cook et al., 1999; Cook et al., 2001; Snyder et al., submitted; Cook et al., in preparation). In this poster, we make some observations about the regional framework derived from combining all data into a coherent cross section across northwestern North America.

To accomplish this, and to keep the cross section as orthogonal to strike as possible, it is necessary to minimize the lengths of line segments that are parallel to strike of deep structures, and to project along strike to connect one line or line segment to another. While this may seem to be only a simple matter of following the strike of surface structures, two problems arise. First, in some regions (e.g., the Western Canada Sedimentary Basin; WCSB)

the surface rock layers are subhorizontal so that the regional strike is undefined. In this case we inferred the strike of regional crustal structures from potential field data (gravity and magnetics).

The second problem is that even where the surface strike is apparent (as with structures in the Cordillera), the strike of deep structures may be different, particularly if they are older. This more difficult problem may be partially addressed by analysis of potential field data (e.g., Cook, 2000; Geiger and Cook, 2001).

Construction of the section thus rests on the following assumptions and correlations:

1. Nearly all of Line 1 is used in the section to provide a continuous profile from the Slave Province to the Proterozoic Fort Simpson basin. Only the part of Line 1 (~stations 8000-8500) that crosses the Great Bear arc along strike (NW-SE based on potential field anomalies; Cook et al., 1999) has been removed from the section (~30 km).
2. Projection from Line 1 to Line 2b is more challenging, in part because the deeply buried rocks of the Fort Simpson basin are covered by the WCSB, but also because the interpretations of potential field patterns are somewhat divergent. A large (>500 nT) north-south oriented magnetic anomaly exhibits a transition from high values to low values along a N-S trajectory from near Fort Simpson, NWT, southward to east of Fort Nelson, BC. By following the magnetic anomaly pattern, it would be logical to correlate from about station 2000 on Line 1 to the east end of Line 2b. In contrast, however, isostatic gravity anomalies, and particularly intermediate wavelength bandpass filtered anomalies, exhibit a more arcuate (convex eastward) shape such that the (filtered) Fort Simpson gravity high projects from near station 2000 on Line 1 toward Stone Mountain Provincial Park (Muskwa anticlinorium) along Line 2b. Because the gravity anomalies appear to correlate positively with deep structural relief throughout much of the eastern Mackenzie Mountains and the Fort Simpson basin (Cook, 2000), we have chosen to use the filtered isostatic gravity for projection. This is convenient (perhaps fortuitously) because the seismic reflections from Proterozoic strata near Stone Mountain Provincial Park are of high quality and match well those on the west side of Line 1.

3. West of the Muskwa anticlinorium, Line 2b follows the Alaska Highway northward from about stations 19000 to 17600. This portion of line is clearly along the strike of the surface structures, and appears to be oblique to the strike of the deep structure (as interpreted from projection of filtered gravity). On that basis, this segment has been removed for the cross section.

4. West of about station 17500 on Line 2b, the profile crosses Cordilleran structures oblique to strike, but it is nearly orthogonal to filtered gravity trends and is therefore included in the section.

5. To illustrate the structure of the Cordillera prior to the assumed amount of dextral movement along the Tintina-Northern Rocky Mountain Trench fault, it is necessary to restore the fault displacement. Estimates of displacement range from about 450 km to >900 km for various geologic times and it is likely that displacement varies along the length of the fault zone (e.g., Gabrielse, 1985). For the purpose of the present cross section, we have chosen to restore 350-400 km, or about the distance from Line 2b to Line 3. If a larger offset were to be chosen, 500 km for example, Line 3 west of the Tintina fault would reconstruct to a position southwest of the Muskwa Anticlinorium and the resultant cross section would be similar, but slightly shorter, to the one presented here. The choice of a smaller offset (e.g. 150-200 km) has equally little effect on our overall interpretation because the deep layers that are interpreted as Proterozoic strata are present on both profiles.

Description of the Section

The most obvious feature of the cross section is that the crust is considerably more reflective than the mantle, irrespective of the age of the region. A corollary is that the transition from crust to mantle, delineated as the reflection Moho on these interpretations, is remarkably flat over the ~1200 km length. The travel time for the reflection Moho varies between about 10.0-11.0 s everywhere except beneath the Fort Simpson Basin where it locally rises to ~9.5 s. The reasons why travel time (depth) to the reflection Moho varies little over such a large part of the continent, beneath regions of crust of such different ages

(Archean to Tertiary), are uncertain, but this observation may shed new light on the nature of the Moho (e.g., Cook, 2002).

Reflections from the upper mantle are rare, although likely tectonically significant. Two important examples are located on Line 1 beneath the Wopmay orogen (Cook et al., 1999) and on Line 3 beneath the Coast Belt (Cook et al., 2001). Both have been interpreted as relict subduction zones, the Line 1 version being Proterozoic and the Line 3 version being Eocene. Although 1.8 billion years separates the generation of these two zones, their geometry (e.g., dip and position relative to the coeval magmatic arc) is similar.

The amount of supracrustal layering (sedimentary and metasedimentary) that comprise the crust along the section is large. In the western Slave Province, layering in the lower crust has recently been interpreted as accreted supracrustal rocks (van der Velden and Cook, submitted). From the Slave Province to the Fort Simpson Basin, much (perhaps most) of the crust displays layering that can be interpreted as deformed supracrustal strata of the 2.1-1.9 Ga Coronation margin (or alternatively, the Hottah terrane: Cook et al., 1999). These layers do not appear to be disrupted by magmatic activity associated with the younger Great Bear magmatic arc. Similarly, from the Fort Simpson axis to the western margin of the Cordillera, much of the crust appears to consist of layers interpreted as 0.6-1.8 Ga Proterozoic supracrustal strata. All of the observed reflection geometry can be explained as basinal/clastic apron deposits on the edge of the progressively growing North American continent without significant contribution of oceanic, arc or other accreted crust.

Concluding Statements

One unresolved issue is the thickness of the strata. Snyder et al (2002) proposed that as much as 30 km thickness of layered wedge strata may be present beneath the Cordillera. Cook et al (1999) have suggested similar thicknesses for supracrustal strata beneath the Wopmay orogen. However, these interpretations remain to be tested. A related fundamental question is: What is "the basement"? From the section displayed here, much of the continental "basement" of northwestern Canada may consist of supracrustal strata of various ages that are contiguous and recognizable as such, even if buried, metamorphosed, intruded and subsequently deformed.

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Preliminary geological interpretation of parts of SNORCLE Line 2

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SNORCLE Line 2 (Cook et al., 2001) crosses the northern Canadian Cordillera from Fort Nelson to Watson Lake, and Watson Lake to Stewart. This poster focuses on the region westward from Muncho Lake where the migrated seismic reflection profile (Cook et al., 2001) displays strong reflectivity for most of the crust. The poster attempts to integrate previous geoscientific studies of the northern Canadian Cordillera with the seismic profile and is intended to present thoughts, questions, biases, and constraints on interpretations of the profile.

Some of the main concerns are: the composition and structure of the crust east of the Northern Rocky Mountain Trench; the expression of mappable transcurrent faults west of the Northern Rocky Mountain Trench; the seismic expression of the Cassiar batholith; the seismic expression of the Slide Mountain Terrane; the relationship between the Cache Creek Terrane and Stikinia; and the character and composition of Stikinia. The Northern Rocky Mountains, from surface studies, appear to have a much more complicated stratigraphic and structural history than is apparent in the interpreted seismic section. Although the NRMT fault appears to cut the entire crust, other transcurrent faults which are in part earlier (possibly as old as Jurassic) do not seem to penetrate the entire crust. Geochemical evidence suggesting that the Cassiar batholith has been derived largely from crustal sources is not easily reconciled with the seismic profile. Slide Mountain and Cache Creek terranes, as interpreted from geological evidence as relatively thin sheets, are clearly expressed in the seismic profile. The collision of Stikinia with the western margin of North America during the early Middle Jurassic closed the Cache Creek ocean and resulted in obduction of Cache Creek Terrane southwest on to Stikinia. Interpretation of the seismic profile suggests that at some time, both terranes moved northeast onto Ancestral North America. An intriguing problem is the great thickness of Stikinia, and its character at depth. Extrapolation from Stikinia surface geology is difficult because 1) its pre-Devonian history is entirely unknown, 2) its Paleozoic history is fragmentary, 3) its pre-Mesozoic structural trends are poorly known, and 4) interference of phases of deformation, including the Mesozoic Skeena Fold Belt complicate seismic interpretation. Some structures of the Skeena Fold Belt which affect the Bowser Basin, and are therefore Mesozoic, also affect pre-Mesozoic strata of Stikinia.

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Modeling the phase relationships and geophysical properties of eclogitic mantle lithosphere with reference to the Slave craton

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Introduction

Cratonic mantle lithosphere is dominated by peridotite but also hosts subordinate amounts of eclogite. Thermobarometric studies of mantle-derived xenoliths of eclogite suggest that this rock type is distributed across most of the mantle lithosphere (e.g., Kopylova et. al 1999a,b). The presence and the ubiquity of eclogite in cratonic mantle lithosphere bears directly on processes of craton stabilization, the thermal state of mantle lithosphere, and the mantle processes operating since the time of stabilization. In this regard, the origin of cratonic mantle eclogite is an important issue; there are several possibilities. The two main explanations involve metamorphism of subducted basaltic crust or crystallization of primary basaltic melts. It is also conceivable that eclogite results from both end-member processes, or that other mantle processes (e.g., metasomatism) are responsible, in part.

Our premise is that the mineralogy, mineral compositions and mineral abundances of eclogite contain important information on their mode of origin and physical-chemical conditions of formation. Furthermore, these data have been underutilized in studies of cratonic mantle lithosphere and, specifically, have not been employed to decipher or constrain the origins of eclogite. Our approach is to use state-of-the-art thermodynamic databases (e.g., Holland and Powell 1998) to compute pressure (P)-temperature (T) pseudosection phase diagrams (e.g., Holland and Powell 1998). These diagrams provide a means to illustrate eclogite phase relationships as a function of P and T for a specific bulk composition. Thus, the effects of protolith on the mineralogy and physical properties of eclogite can be shown explicitly by calculating pseudosection phase diagrams for a variety of pertinent bulk compositions.

We use these quantitative models of phase relationships in eclogite at mantle conditions in three different ways:

- (1) to predict mineral composition and abundance, and rock density and heat capacity along a calculated geotherm for a variety of basaltic protoliths (Table 1),
- (2) to search for diagnostic minerals whose presence, even as xenocrysts, places significant constraints on the origins or pressure-temperature conditions of eclogite formation,
- (3) to compute P and S wave velocities of isotropic eclogite from the predicted mineral assemblages as a function of depth along the calculated geotherm.

Table 1. Summary of compositions and normative mineralogies of basalt protoliths used to model eclogite properties in lithospheric mantle as a function of P-T conditions.

No.	1	2	3	4	5	6
Rock Type ¹	AOB*	Ankaramite	AOB	Tholeiite	MORB	Hi-Al basalt
SiO ₂	45.39	43.28	47.38	48.35	49.20	51.95
TiO ₂	2.52	4.12	2.81	1.57	2.03	1.15
Al ₂ O ₃	14.69	14.43	13.50	15.49	16.09	18.03
Fe ₂ O ₃	1.87	0.70	3.18	3.26	2.72	4.97
FeO	12.42	10.92	9.74	8.05	7.77	5.20
MnO	0.18	0.13	0.19	0.17	0.18	0.12
MgO	10.37	11.68	9.24	7.03	6.44	5.52
CaO	9.14	11.22	9.92	9.92	10.46	8.70
Na ₂ O	2.62	2.49	2.58	2.76	3.01	3.35
K ₂ O	0.78	0.83	0.54	0.51	0.14	0.98
P ₂ O ₅	0.02	0.31	0.40	0.24	0.23	0.24
H ₂ O	0.00	0.08	0.21	1.52	1.65	0.00
Sum	100	100.19	99.69	98.87	99.92	100.21
Normative Mineralogy						
q		-	-	-	0.08	3.05
or		4.90	3.20	3.01	0.84	5.97
pl		33.32	44.59	51.72	55.45	59.59
lc		-	-	-	-	-
ne		7.32	-	-	-	-
di		22.52	18.49	15.64	16.61	8.21
hy		-	12.00	16.04	16.97	13.60
ol		22.49	9.50	2.68	-	-
mt		1.02	4.62	4.72	3.94	7.20
il		7.83	5.35	2.97	3.85	2.19
ap		0.74	0.95	0.57	0.54	0.57

¹ Sources include (1) AOB* is composition used by Green & Ringwood (1967); Analyses 2-6 are taken from Raymond (1995).

Thermodynamic Models for Phase Assemblages

Mantle eclogite generally contains the assemblage clinopyroxene–garnet–rutile with subordinate kyanite, orthopyroxene, and other phases. SiO_2 polymorphs are generally lacking. The phase assemblage is a reflection of bulk composition and T and P conditions in the cratonic mantle lithosphere. Pressure-temperature pseudosections were calculated from PERPLEX (Connolly 1990), using the database and solution models of Holland and Powell (1998). Our model chemical system comprises Na_2O – CaO – MgO – FeO – Al_2O_3 – SiO_2 – TiO_2 and we considered the following phase components: quartz, coesite, anorthite, high albite, kyanite, diopside, hedenbergite, Ca-tschermak pyroxene, Mg-tschermak pyroxene, Fe-tschermak pyroxene, jadeite, enstatite, ferrosilite, forsterite, fayalite, pyrope, almandine, and grossular. The idealized protolith compositions (Table 1) were chosen to span the full range of basalt compositions that could be expected as primary magmas or as subducted seafloor crust. For these calculations all ferric iron was converted to ferrous iron.

Calibration of P-T pseudosections

We have used the results of high-pressure crystallization experiments from Green and Ringwood (1967) to validate the model. Their experimental data provide approximate modes and mineral compositions produced in samples of alkali olivine basalt (protolith 1 in Table 1) as a function of T and P. The experimental database covers a pressure range of 1 bar–30 kbar at 1100 °C. We have used PERPLEX to simulate these experimental results. Model predictions completely reproduce the sequences of mineral occurrence and disappearance in P-T space. Specifically, our model correctly predicts with increasing pressure, the appearance of garnet, the disappearance of olivine, the disappearance of plagioclase and the appearance of rutile. The model and experiments differ in that our model predicts olivine disappearance and plagioclase disappearance at lower pressures than observed. Our model also predicts that ilmenite and rutile will coexist in stable equilibrium at high pressure whereas Green and Ringwood noted only rutile in the experimental charges. We have also compared our pseudosection calculations to those of Holland and Powell (1998) for aluminous andesitic basalt and although they used different thermodynamic solution models the results are in reasonable agreement.

Phase relationships in basaltic protoliths

The thermodynamic modelling uses as input the bulk compositions of protoliths to compute eclogite mineralogy. The protolith compositions that we have used are basaltic and are representative of isochemical subduction of unaltered oceanic crust. They include Mid-Ocean Ridge Basalt, Alkali Olivine Basalt, High Al Basalt, Tholeiite Basalt and Ankaramite (Table 1). We have computed P-T phase diagrams for each protolith using a temperature and pressure range of 750 to 1500 K and 10 to 60 kbar, respectively. The results are shown in five pseudo-section diagrams (Fig. 1); each phase diagram summarizes the mineral assemblages expected as a function of mantle conditions (i.e., P and T). The differences shown by the five phase diagrams result directly from the differences in protolith composition.

Figure 1 makes two important points. First, the phase relationships of cratonic mantle eclogite are highly dependent on protolith composition; there are substantial differences in the general appearance of these 5 P-T phase diagrams. Second, the overall mineralogy of mantle eclogite is relatively simple and somewhat insensitive to the ranges of composition found in these basaltic protoliths. Over the full range of P-T conditions explored, the minerals Grt, Cpx, Pl, Opx, Ol, Ilm can be produced from all five protolith compositions (mineral abbreviations after Kretz, 1983). In detail, however, there are substantial mineralogical differences between protoliths based on the specific minerals (i.e., Ky, Qtz) or the P-T conditions that support specific minerals (i.e., Ilm, Ol).

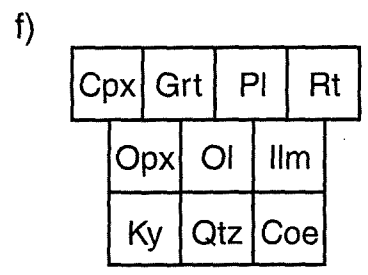
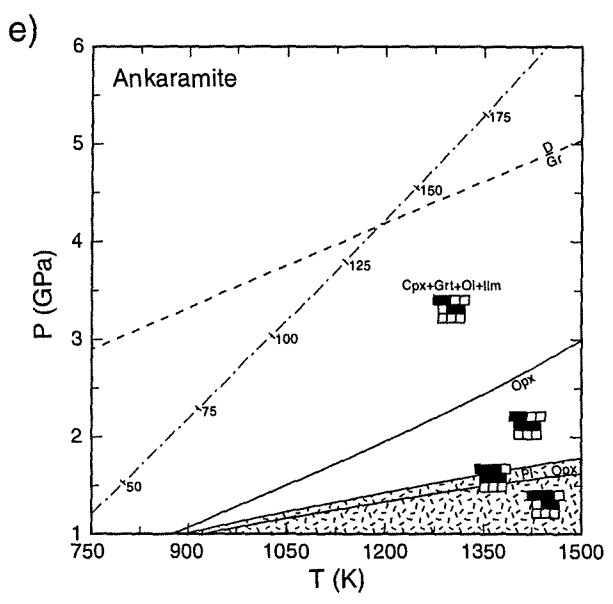
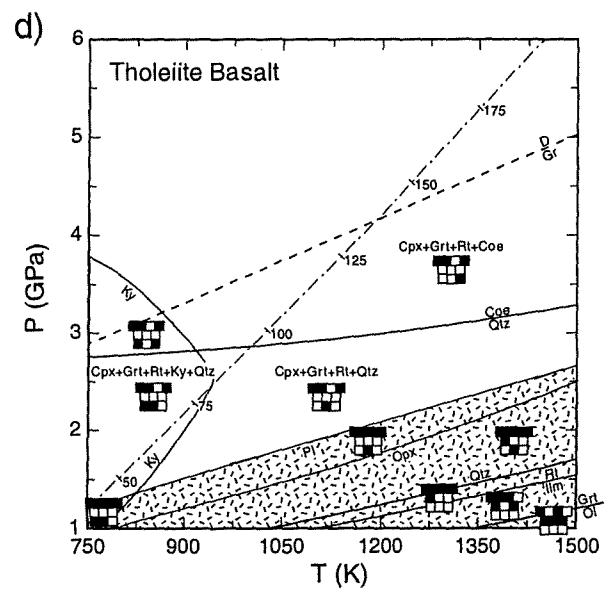
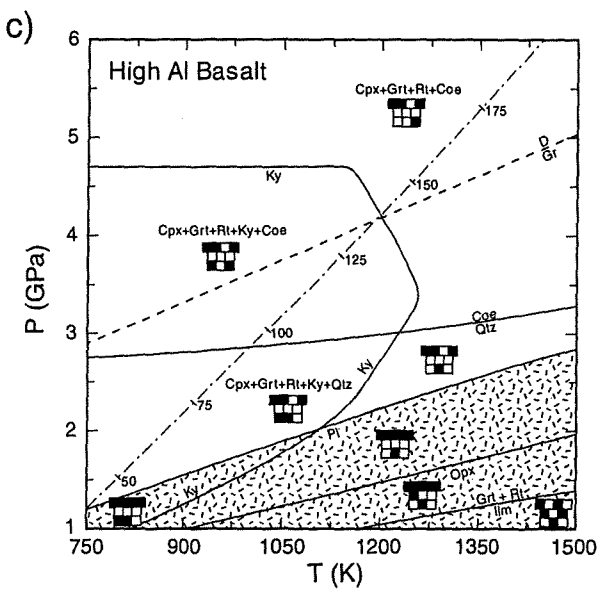
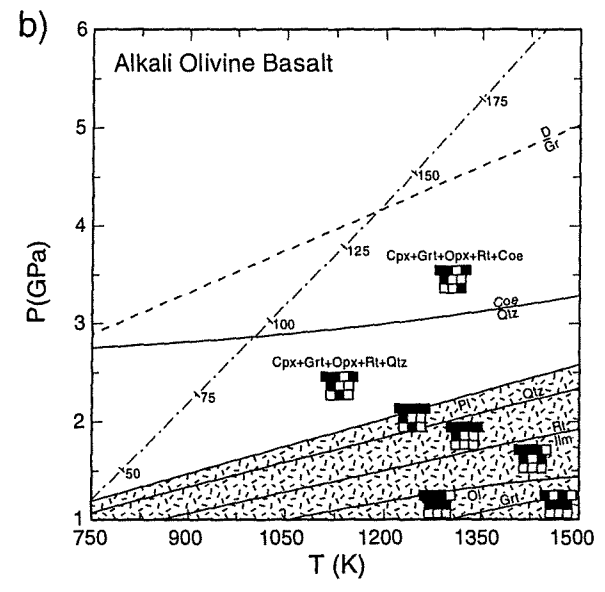
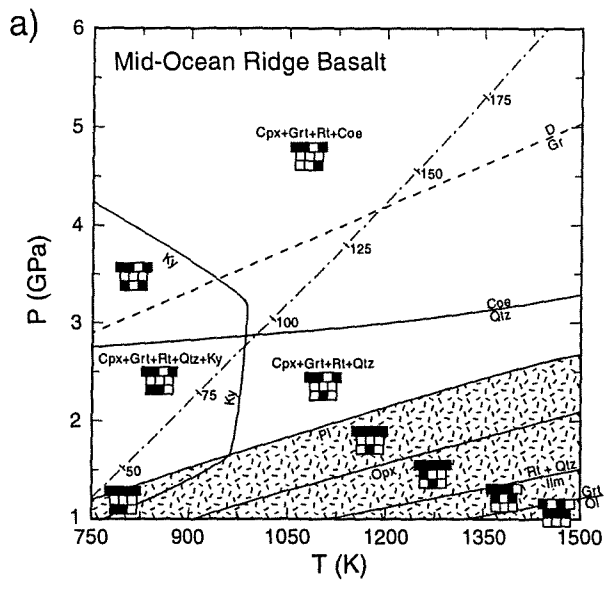
All protoliths support the widespread (i.e. P, T) occurrence of the minerals Cpx and Grt. Pl is also common to all protoliths (coarse pattern) at low pressure; its stability field is severely reduced in ankaramite (Ne-normative). Opx and Rt are stable over a wide range of P-T conditions for all protoliths except Ankaramite. Ol and Ilm are practically absent in all protoliths except at high-T, low-P conditions. Olivine is replaced by Opx and Ilm by Rt. Ankaramite is an exception; Ol is stable and replaces Opx at higher pressures and Ilm replaces Rt over all P-T conditions (Fig. 1). The occurrence and stability of SiO₂ polymorphs is highly dependent on silica activity. Neither Qtz nor Coe are present in Ankaramite; in alkali olivine basalt Qtz production occurs near the Pl out line. High-Al basalt supports silica polymorphs over all P-T conditions and Qtz occurs in MORB and tholeiite except at very high-T and low-P conditions.

Kyanite forms in several protoliths (Fig. 1). The largest stability field occurs in High-Al basalt followed by MORB and Tholeiite basalt; silica undersaturated compositions lack Ky. The P-T range over which Ky is stable in any of these protoliths is restricted. In normal metaluminous compositions Ky is stable at temperatures below 975 K and pressures below 4.2 GPa (only at T < 800 K); High-Al basalt increases the stability field to temperatures below 1200 K and pressures below 4.7 GPa. The presence of Ky-bearing eclogite (e.g., Kopylova et al., 1999b), therefore, can potentially constrain both the protolith composition, as well as, the formation conditions. Unless the protolith is exceptionally high in Al₂O₃, eclogite must derive from the shallow mantle (e.g., < 3 GPa).

Eclogite on the Slave geotherm

In Figure 1 we showed the P-T phase relationships for model eclogite over the complete range of temperatures and pressures found in the cratonic mantle lithosphere. However, the P-T conditions found in the mantle are a small subset of this space and can

Figure 1 (next page). Calculated P-T pseudo-section phase diagrams for eclogite in cratonic lithospheric mantle shown as a function of protolith composition. Basaltic protolith compositions are reported in Table 1 and include: (A) MORB, (B) Hawaiian alkali olivine basalt, (C) Hi-alumina tholeiite, (D) tholeiite basalt from Columbia River plateau, and (E) ankaramite. Shaded fields denote the P-T stability fields for plagioclase (coarse) and for kyanite (fine). Dashed line indicated the relative stability fields of graphite (Gr) and diamond (D). Dashed-dotted line shows the model geotherm for the Slave Craton established from thermobarometric studies of peridotite xenoliths (Kopylova et al., 1999a; Russell & Kopylova 1999; Russell et al; 2001); labels are depths (km). Solid lines pertain to eclogite phase relationships. All mineral abbreviations are taken from Kretz (1983) except D (diamond) and Coe (coesite).



be approximated by the steady-state conductive geotherm derived from thermobarometric analysis of peridotite xenoliths. Consequently, our analysis focuses on variations in eclogite mineral content and physical properties as a function of protolith composition along the geotherm for the Slave mantle lithosphere (Kopylova et al., 1999a; Russell & Kopylova, 1999; Russell et al. 2000). A comparison of the model geotherm to the calculated phase relationships for each protolith (Fig. 1) shows that there is a very much more restricted range in mineralogy permitted for individual protoliths. For example, kyanite-bearing eclogite is limited to the graphite stability field even in high Al basalt. Within the diamond stability field each pseudosection had a single mineral assemblage along the Slave geotherm, although the mineral assemblage does vary between protoliths. MOR, High Al basalt and tholeiite all contain Cpx+Grt+Rt+Coe, alkali olivine basalt contains this assemblage plus Opx, and ankaramite consists of Cpx+Grt+Ol+Ilm. The implication is that for a relatively homogeneous protolith such as MORB, diamondiferous eclogite will have one characteristic mineral assemblage.

A property of pseudosection phase diagrams that is not evident in Fig. 1 is that the abundance and composition of minerals may vary within each divariant field. This information is computed by the Gibbs Free Energy minimization routine in PERPLEX and we have used it to map variations in mineral abundance and composition with depth along the Slave geotherm (Fig. 2). The mineral abundances at specific P-T conditions have also been used to compute the expected density and heat capacity of the eclogite with depth in the mantle; these variations reflect the effects of mineralogy, mineral composition and P-T conditions. We are currently using these data to compute seismic velocities of eclogite as a function of protolith composition and depth along the geotherm.

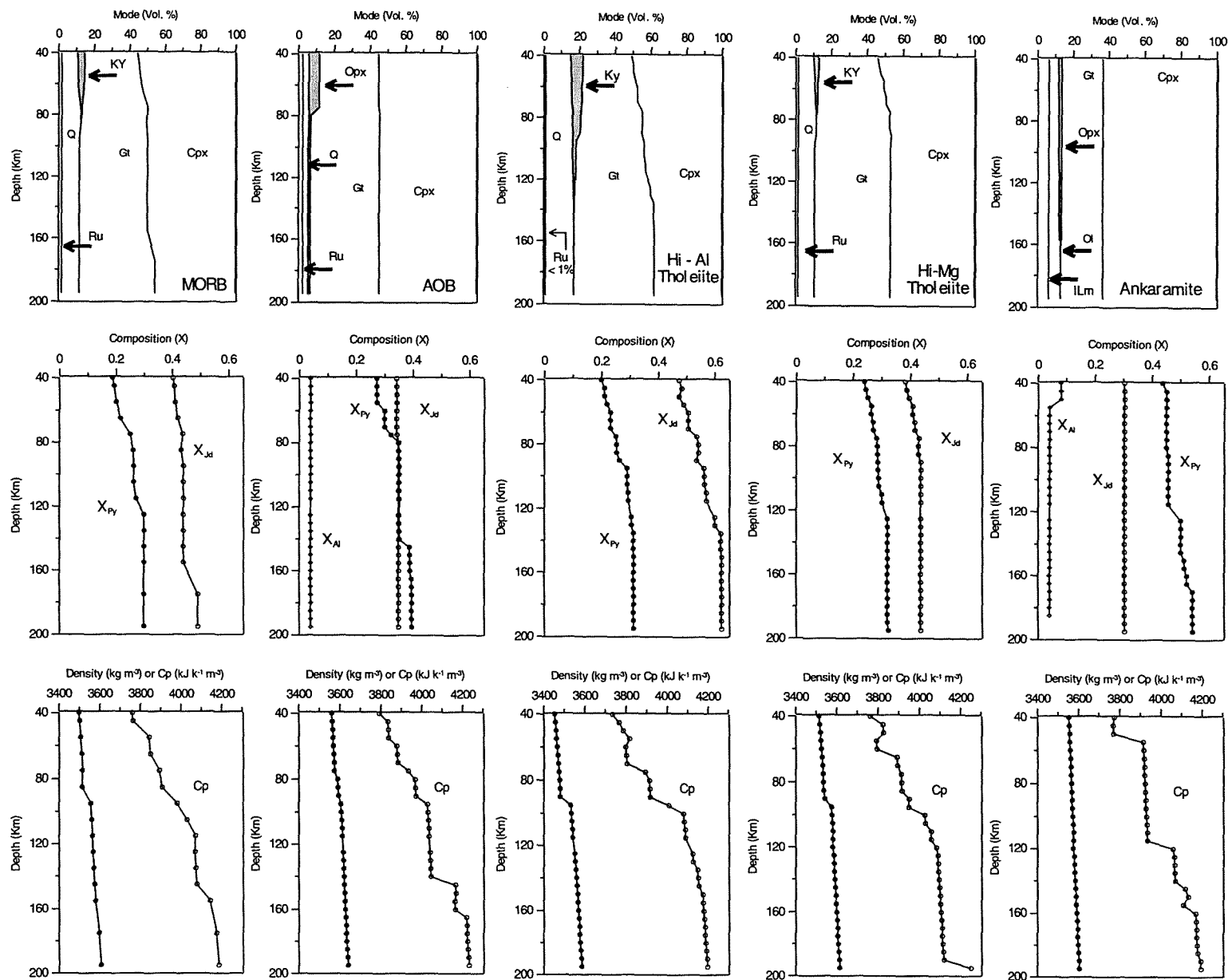
Future Investigations

These preliminary results represent an alternative means of learning about the cratonic mantle lithosphere. Specifically, this thermodynamic-based modelling provides constraints on the origins of eclogite and insights into the organization of the mantle lithosphere. There are several avenues of investigation that logically derive from our results to date.

Other Origins

A comparison of the bulk compositions for mantle-derived eclogite (Type A of Coleman et al, 1965) with Type C eclogite associated with blueschist (Coleman et al, 1965), Jericho eclogite, and basaltic protoliths indicate significant compositional differences. This could be the result of metasomatism of original basaltic protoliths although this metasomatism is not apparently related to alteration of subducted oceanic crust. Another possibility is fractional crystallization of basaltic magma at mantle depths

Figure 2 (next page). Phase relationships calculated for eclogite at P-T conditions defined by the steady-state geotherm shown in Fig. 1 are shown as a function of depth (km) and include: mineral abundance (mode), mineral composition (jadeite component in Cpx and pyrope component in Grt), and bulk density and heat capacity of the eclogite.



and subsequent metamorphism.

We are, therefore, exploring the phase relationships of protoliths resulting from these two other processes. For example, we have used average compositions of seafloor altered basalt (Humphris and Thompson, 1978), to model eclogite formation from altered seafloor crust. Alternatively, our models will use as protoliths the compositions of igneous residues produced by the crystallization of basaltic melts at mantle pressures. Using Pmelts (Ghiorso and Sack, 1995) we have modeled the consequences of equilibrium crystallization at upper mantle pressures for each of the basalt protoliths in Table 1.

Geophysical Implications

The main attribute of using a thermodynamic model to describe the equilibrium state of eclogite at mantle conditions is that it provides a complete mineralogical characterization at specific P-T conditions. We are using the mineralogy, modal abundances, and mineral compositions to compute *in situ* P and S wave velocities as a function of protolith composition and depth along the geotherm from the isentropic bulk modulus and shear modulus of the relevant isotropic polycrystalline aggregates.

Jericho Eclogite

We have compared the range in properties (mode, mineral composition) with the measured properties of eclogite xenoliths from the Jericho kimberlite pipe. On the basis of this comparison, we are able to constrain the nature of the protoliths to these mantle-derived eclogite. In addition, we calculate an ideal pseudosection phase assemblage for the Jericho eclogite xenoliths based on their bulk compositions and their apparent mantle P-T conditions. Differences between the ideal phase assemblages, modes and mineral compositions and the observed properties may indicate metasomatic effects or sampling problems for small samples of xenoliths.

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Tectonic Evolution of the Selkirk Fan: A Composite Middle Jurassic – Cretaceous Structure, Northern Selkirk Mountains, Southern British Columbia

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A zone of structural divergence marks the transition from the hinterland (Omineca belt), characterized by penetrative ductile deformation, high-grade metamorphism and plutonism, to the “thin-skinned” foreland. An en echelon series of regional-scale northwest-trending structural fans occupy this zone, and extend from northeastern Washington to east-central Alaska. Understanding their development is fundamental to elucidating the transition from hinterland to foreland tectonics in the Cordillera.

In the northern Selkirk Mountains of the southern Omineca belt there is excellent exposure of one of these structures, termed the Selkirk fan structure. The fan is underlain by medium- to high-grade metamorphic rocks, and comprises at least three generations of superposed deformation. Southwest verging, second generation folds (F_2) with shallow dipping axial surfaces (S_2) dominate the west flank of the fan, and become progressively steeper towards its axis. To the east of the fan axis are northeast verging, second (F_2) and third generation folds (F_3) with axial surfaces that become progressively overturned eastward toward the Rocky Mountain Trench, merging synthetically with structures in the Foreland belt. Assignment of fold generation is based on overprinting and geometric observations; no regional timing correlation is implied, especially across the fan. During the past few decades, the kinematic development of the Selkirk fan structure has been the focus of considerable debate. New U-Th-Pb geochronologic data obtained by small-fraction Isotope Dilution-Thermal Ionization Mass Spectrometry (ID-TIMS) and Sensitive High Resolution Ion Microprobe (SHRIMP) analyses point to a complex and protracted origin for the Selkirk fan structure, requiring significant revision of previous models.

Dated zircon, monazite, and xenotime from variably deformed leucocratic dykes, monzonitic plutons, and pelitic schists indicate that the thermo-structural development and exhumation of the west flank of the fan occurred principally in the Middle Jurassic (ca. 173-167 Ma). In contrast, data from east of the fan axis strongly suggest that there has been significant deformation (associated with F_3 ?), and metamorphism superimposed on an early transposition fabric, and that exhumation occurred in the mid-Cretaceous to Early Tertiary (ca. 100-62 Ma). We propose a new tectonic model that attempts to reconcile these data with previously documented geochronologic, structural, and metamorphic data.

During the Early to Middle Jurassic accretion of the Intermontane superterrane, a proto- $F_{1,2}$ fan developed above a singularity where oceanic or marginal basin lithosphere was subducting eastward beneath continental lithosphere. Southwest verging structures developed within the prowedge (i.e., F_2 in west flank), immediately followed by development of northeast vergent structures in the retrowedge (i.e., F_2 in east flank). Subsequent decoupling of the fan above a basal decollement allowed it to be carried northeastward toward the foreland. Rocks to the east were progressively incorporated into a critically tapered orogenic wedge as it propagated northeastward. In general, this resulted in diachronous, eastward younging of medium to high-grade metamorphism and ductile deformation (i.e., southwest dipping F_2 in east flank of fan younger than northeast dipping F_2 in west flank).

The mid-Cretaceous (ca. 100 Ma) accretion of the Insular superterrane to the western margin of North American may have resulted in the rejuvenation of compressional forces acting upon the orogenic wedge. If so, out-of-sequence thrusting (e.g. Purcell thrust) and folding (F_{2-3}) may have served to thicken the tectonic pile, thus re-establishing critical taper and continued deformation in the foreland. This could explain the development of mid- to Late Cretaceous (ca. 100-80 Ma), northeast vergent F_3 folds found primarily in the east flank of the fan. At this time, the east flank of the fan was uplifted and exhumed relative to the western flank, bringing up deeper-seated rocks that were significantly tectonized in the mid-Cretaceous. The juxtaposition of higher structural levels with an older deformation history in the west flank relative to lower levels with a younger deformation history in the east may have also been accommodated in part by a number of southwest dipping normal faults, possibly as northeast directed contraction was waning. Thus, the Selkirk fan may be thought of as a composite structure of juxtaposed Middle Jurassic and Cretaceous structures, rather than a singular fan that developed during one progressive event.

Database of Rock Seismic Properties (DRSP)

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Whereas the upper crust is accessible to geological sampling and mapping, the deeper portions of the crust and the upper mantle are relatively inaccessible. Much of our understanding about the composition, structure and evolution of the deep crust and upper mantle has been derived from various seismic refraction and reflection measurements. Interpretation of these seismic data are mainly based on our knowledge of the elastic properties of rocks and rock-forming minerals, the latter obtained from laboratory measurements of ultrasonic velocities and density. In addition, physical properties of rocks have significant practical applications in the mining industry and geotechnical engineering (e.g., construction of buildings, dams, tunnels, bridges and nuclear waste storage). Therefore, it is of great importance to assemble and collate as well as statistically analyze all the data on seismic properties of rocks from around the world.

The past 4 decades have been a virtual explosion in the amount of data available on the seismic velocities and anisotropy of rocks from around the world. Currently, there is no centralized location for information on seismic properties of hundreds different types of rocks from continents and oceans. Yet, there is little infrastructure available to help provide robust and rapid response to queries over the internet. Thanks to the support from the Canadian national LITHOPROBE project, we established an internet database of rock seismic properties (DRSP, http://tectono.civil.polymtl.ca:8080/jsp_suli/homePage.html), which comprises almost all data available in the literature published in English and French during last 4 decades and is more complete than any previous compilations. Obviously, such a computerized database, compared with a traditional compilation in the form of tables, can perform much more complex tasks with much higher speed and efficiency, and permit enormous volumes of data to be studied in a far more comprehensive fashion than has hitherto been attempted.

The DRSP allows its user to retrieve the compressional and shear wave velocities (V_p and V_s) and their anisotropy (A) of rocks according to geographic location (country, province, region, longitude and latitude), tectonic province, rock category (igneous, metamorphic and sedimentary rocks and ores), lithology (granite, basalt, eclogite, mylonite, sandstone, and many others), propagation and vibration directions (X - parallel to the stretching lineation, Y - perpendicular to the lineation and parallel to the foliation, and Z - normal to the foliation), strain (deformed or undeformed), humidity (dry or wet) and source literature.

The DRSP also provides its user a statistical display on:

- Relationship between rock density and seismic velocity (average, or for each given propagation direction) at given pressure and temperature;
- Histogram distribution of seismic velocities and anisotropy for each lithology or rock category as well as for all the rocks from a given region;
- Relationship between sedimentary rock porosity and seismic velocities;
- Effects of pore-fluid on seismic properties and anisotropy;
- Quantitative correlation between seismic properties and mineralogy and chemical composition;
- Correlation between seismic properties and finite strain axes, providing a valuable tool to constrain the interpretation of seismic and teleseismic data from various tectonic provinces.
- Average V_p and V_s data for calculating the V_p/V_s ratio and Poisson's ratio, with a special attention paid to the possibility of using these ratios to determine lithological composition.

The DRSP is also an educational resource at both the undergraduate and graduate levels. With the help of the database, students can better understand the P-and S-wave velocities and their anisotropy for rocks which constitutes the crust and upper mantle at different depths. Such knowledge is needed in order to interpret the seismic data from refraction and reflection experiments. As a bridge between petrophysics and engineering the database provides infrastructure to the increasingly interdisciplinary nature of modern research.

Thus the DRSP can be used for various purposes. It will serve as a convenient, compact, yet comprehensive source of basic information and will keep the world earth

[illegible]

Possibility of channel flow in the southern Canadian Cordillera; a new approach to explain existing data.

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Existing structural, metamorphic and geochronologic data in and close to the Shuswap Metamorphic Complex in the southern Canadian Cordillera are shown to be consistent with a channel flow model. Previously, ductile deformation was explained in terms of large scale thrusts or detachments, that were linked with brittle thrusts in the Rocky Mountains. These detachments required extensive displacement, and explain little of the ductile deformation. Ductile deformation involves large non-coaxial strain which can explain the transposition, and can transform discontinuities, such as unconformities and faults, into thrust-like structures. These structures have thrust geometry but lack thrust-related displacement (Williams, 1999). The non coaxial flow converts pre-existing folds, and dragfolds that form during the non-coaxial flow, into isoclinal dismembered folds, as seen throughout the southern Omineca Belt. Up to three generations of such recumbent folds can be recognised in the southern Canadian Cordillera (e.g., in the Monashee Complex). These recumbent folds are overprinted by one or more generations of late, upright, or steeply inclined, generally post-metamorphic folds. All structures are modified by (Late Cretaceous-?) Early Tertiary crustal extension.

In the southern Canadian Cordillera, three general structural levels can be recognised, based on the recumbent folds. (1) At the deepest structural level (e.g., within the Monashee Complex, and immediately overlying rocks within the Shuswap Metamorphic Complex) folds of all generations are NE vergent. (2) At a higher level within the Shuswap Metamorphic Complex, early folds are NE vergent and late folds are SW vergent. (3) At the highest structural level (e.g. the Selkirk Fan, Illecillewaet Synclinorium, Clachnacudainn Complex and Cariboo Mountains), all folds are SW vergent. The rocks of the highest structural level cooled in the Middle Jurassic, whereas rocks in the lower levels cooled in the Late Cretaceous (Kuiper et al. 2001), or in the Early Eocene.

The three levels may represent levels within a channel flow zone (cf., Beaumont et al., 2001): The NE vergent folds of the lowest level may have formed during top-to-the-NE detachment flow and/or in the lower part of a channel flow zone where the sense of shear would be the same. When detachment flow changed to channel flow, the sense of shear would have changed in the upper part of the zone. NE vergent folds would be overprinted by SW vergent folds, as is observed in the intermediate level. If temperature was increasing, it is conceivable that the channel flow zone got wider and this would explain why all of the structures in the highest level verge to the SW.

The channel flow model as presented here explains all ductile structures, accommodates existing metamorphic and geochronologic data, and requires less displacement than earlier models. Although this model may not be a unique solution, it demonstrates that there are different, and perhaps more reasonable ways of explaining existing data. It is important that current interpretations are constantly reevaluated, and that new models are developed.

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Influences on Structural Geometry in the Fort Liard region: New data from the Central Foreland NATMAP Project

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After two years of concentrated fieldwork in the Fort Liard (95B) and La Biche (95C) map areas, the Geological Survey of Canada and its partners have published 5 Open File maps at 1:50,000 scale, and several more are in production. This first reassessment of the regional mapping since the early 1960's incorporates much recent stratigraphic information into the new maps. Although the overall regional geometry identified by previous regional maps remains valid, our more detailed mapping has resulted in revisions to key aspects of the structural geometry, with important implications for the resource industry.

The characteristic sigmoid geometry of structures in the area is achieved partly by curvature of axial surfaces, and partly by *en echelon* arrangement of structures. This geometry appears to reflect the interference between two structural trends. The north-northwest trending structures define the regional Cordilleran deformation trend. The interfering northeastward trend has been postulated previously to reflect the influence of a series of basement structures that make up the Liard Line. Together these trends create local culminations along individual structures. Thus individual large anticlines may contain several distinct and unconnected gas accumulations. Ongoing study of seismic transects and potential field data are being used to improve our understanding of the three-dimensional geometry of these complex interference structures.

The structures observed at the surface are dominated by large-scale folds with wavelengths of 10 to 25 km. The wavelength of these structures is mainly controlled by a thick, strong structural beam, comprising the Carboniferous Flett and Mattson Formations, bounded above and below by weaker shale. Folding style in the area varies from symmetrical to asymmetrical. On the Kotaneelee, La Biche, and Beavercrow structures a strong western vergence is developed, and minor west-directed thrust faults locally cut the west limbs of these structures. Thrust faults have subsidiary importance at surface and appear to be best developed as accommodation features in areas of localized high strain. Regionally, the structures are detached in the subsurface. On the basis of previous experience in adjacent northeastern British Columbia, the principal décollement horizon in the Liard-La Biche region is interpreted as the Besa River Formation, a thick shale succession of Late Devonian to Early Carboniferous age. Because the Manetoe facies dolomite, occurring in carbonate strata below the Besa River Formation, is an important local reservoir in structural traps, additional décollements must occur beneath the lower Paleozoic carbonate succession and locally ramp upward through the carbonates.

Lineations of the Monashee Complex: Implications for tectonic interpretation.

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We describe the morphology and mineralogy of lineations found within the Monashee Complex and their relationship to other structures at the microscopic and macroscopic scale. Using these basic descriptions and the spacial distribution of lineations, we emphasize two important problems which are relevant to the tectonic evolution of the southern Omineca Belt.

Lineation is commonly used in conjunction with other structures to interpret the kinematic history of rocks. Recent studies show that the relationship between lineation and kinematics is complex and subject to interpretation. A thorough understanding of the genetic relationship of lineation to other relevant structures is essential. Examples of these complexities are considered for the Monashee Complex, which suggest that the movement picture, even with simplifying assumptions, is not without ambiguity.

Fabrics (including lineations) in deformed rocks are routinely referred to by their genetic significance. One example is naming the lineation that is interpreted as having developed during an crustal extensional as the "extension lineation." Though this is acceptable within the context of a tectonic model, these terms commonly become entrenched. The prominent lineation developed on the eastern edge of the Monashee Complex, is shown to be a composite lineation that was not necessarily produced by crustal extension. Thus, description of lineation based on morphology is preferred, as tectonic interpretations are subject to revision.

Finlayson Lake Targeted Geoscience Initiative: New Bedrock Geological Map of the Northern Part of Finlayson Lake Area (105G), Southeastern Yukon

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The joint Geological Survey of Canada – Yukon Geology Program Finlayson Lake Targeted Geoscience Initiative comprised programs of new bedrock geological mapping and till geochemistry in the northern two thirds of the Finlayson Lake map area. The goals of this project were to extend to the north and west recent mapping of the Yukon-Tanana Terrane host rocks of the Finlayson Lake massive sulphide district and to use till geochemistry to 'remote sense' new exploration targets in this poorly exposed area. The till geochemical data and a new bedrock geological compilation of northern Finlayson Lake area are now available from Yukon Geology Program.

Highlights of the bedrock mapping program include delineation of the stratigraphy of Yukon-Tanana Terrane north and west of the area of known VMS deposits and identification of key structural boundaries within the terrane and between Yukon-Tanana Terrane and the North American miogeocline. The western part of the area, from the Tintina Fault to just north of the Campbell Highway, is underlain by rocks of the highly prospective Grass Lakes succession, including chloritic phyllite of the Fire Lake mafic metavolcanic unit, host of the Fyre Lake deposit, and carbonaceous phyllite and lesser muscovite-quartz phyllite of the Kudzu Kayah felsic metavolcanic unit, the rock unit that hosts the Kudzu Kayah and GP4F deposits. The lower part of the Grass Lakes succession is intruded by a large body of Devon-Mississippian granitic meta-plutonic rock, the Mink Creek orthogneiss. A previously undocumented roof pendant in the orthogneiss is made up mainly of locally pyritic, chloritic phyllite of the Fire Lake unit. The Grass Lakes succession and orthogneiss are sandwiched structurally between underlying Triassic phyllite and undated mafic and ultramafic metaplutonic rocks, and overlying felsic to intermediate volcanic rocks and Carboniferous limestone of the Money Creek thrust sheet. The rocks of the Money Creek thrust sheet are in turn thrust to the north over carbonaceous phyllite, greywacke, chert-pebble conglomerate, chert and lesser felsic metavolcanic rocks of the Finlayson succession along the Jules Creek thrust. Basalt and chert of the Early Permian Campbell Range succession and a syn- to post-tectonic polymictic conglomerate of mid-Permian to possibly Triassic age locally overlie the Finlayson succession and the Jules Creek thrust. The Finlayson and Campbell Range successions and the Permian-Triassic conglomerate are thrust to the north over Silurian to Triassic rocks of the North American miogeocline along the Inconnu thrust.

REVISION OF DORSEY TERRANE ROCK UNITS DEPICTED IN A NEW MAP IN SOUTHERN YUKON (105B/4)

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Exposed bedrock in the Dorsey Terrane was previously traversed during the Ancient Pacific Margin NATMAP project with assistance of members of the British Columbia Geological Survey. Dorsey Lake map area (105B/4) of southern Yukon encompasses most of the major units of this former terrane (made obsolete by better understanding of stratigraphy and regional correlations along the structural grain). From southwest to northeast across the map area the major mid- to Late Paleozoic units are: Big Salmon complex; a contact-metamorphosed succession of Swift River clastics, Klinkit limestone and volcanic epiclastic rocks, and the Dorsey assemblage. These record the construction of three successive arc complexes, interleaved with shelf-edge and basin sediments. The map area is representative of both the stratigraphy and regional structure, and presents challenges in interpreting the original nature of contacts through the poly-phase deformational overprint.

Big Salmon complex, previously studied in adjacent BC, underlies Hazel Ridge in the southeast third of the map area. The distribution of the metabasalt, thin meta-chert and marble succession in the complex is controlled by recumbent third-phase folds; these units are capped by an upright meta-greywacke unit on a probable disconformity. Foliated Late Devonian meta-granite intrudes the metabasaltic unit.

Dorsey assemblage quartzose metasediments in the extreme northeast form part of a 200 km trend. They indicate a continental margin origin but are structurally interleaved with mafic and Late-Devonian felsic volcanic rocks. All were deformed prior to intrusion of the mid-Permian Ram stock, and locally contain high P-T mineral assemblages unique to the region. An outcrop with similar composition and fabric several kilometers southeast of the map area suggests that Dorsey assemblage might underlie most, if not all, the eastern part of the map area.

The central part of the map area comprises 3 units that are collectively less deformed than the previous: the Swift River succession, Screw Creek limestone and Klinkit assemblage. These units form a carapace above the Cretaceous Seagull batholith; contact metamorphism generally obscures both primary structures and contact relationships. Swift River includes decimeter-bedded meta-sandstone, siltstone and black argillite, with uncommon quartzite, grit and boulder conglomerate. This succession appears to be faulted atop Dorsey assemblage but the sense of motion is inconclusive. The full age range of the Swift River succession is unknown; it is conformably overlain by Pennsylvanian Screw Creek limestone.

The Klinkit consists of three units that are to be defined for formation status. Lowest is the volcanoclastic unit of chloritic meta-arkose and fragmental metavolcanic beds. This formation yielded a magmatic zircon of mid-Permian age south of the map area; it is probably a distal arc facies. Above the volcanoclastic unit is dark meta-siltstone with minor limestone. Above these sedimentary rocks, and also unconformably above the Swift River succession in the Logjam area, lies the Teh formation, a thick black argillite unit with minor limestone containing Early to Middle Triassic conodonts. The boundary between Klinkit and Big Salmon complex is covered in the map area and cryptic in other places; it is likely faulted.

The Dorsey Lake map area (105B/4) encompasses rocks which help to show that Dorsey Terrane is not a distinct entity. Some units have close affinity to those of other terranes: Yukon-Tanana (Big Salmon Complex), others to Kootenay (Dorsey Assemblage) and Quesnellia (Klinkit Assemblage). Although exposure is mediocre and the deformation overprint challenging, this map area provides several constraints on the tectonic evolution of the pericratonic belt.

Geological map of Yukon–Tanana Terrane and younger rocks in southern Stewart River area, Yukon Territory¹

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¹Contribution to the Ancient Pacific Margin NATMAP Project

INTRODUCTION

Geological mapping in the Stewart River area is a component of the Ancient Pacific Margin NATMAP Project, initiated by the Geological Survey of Canada, Yukon Geology Program, and British Columbia Geological Survey Branch. The NATMAP Project seeks to understand the composition, relationships, and metallogeny of poorly understood pericratonic terranes lying between the ancestral North American margin and those known with more certainty to be tectonically accreted (Thompson et al., 2000; Colpron et al., 2001). The Stewart River component focuses on the Yukon–Tanana Terrane, comprising complexly deformed, mostly (?) Paleozoic meta-igneous and metasedimentary rocks. Previous work in the Stewart River area is summarized in Ryan and Gordey (2001a, b).

The objective of the Stewart River project is to investigate the stratigraphic, structural, and tectonic history and the economic framework of the Yukon–Tanana terrane, by mapping parts of the eastern two-thirds of the area over a four-year period. New data and previous work will be synthesized into a new geological map of the Stewart River area. Detailed aeromagnetic and gamma-ray surveys have been conducted in the Stewart River area under the Targeted Geoscience Initiative (Shives et al., 2001) and will provide additional information to constrain interpretation of bedrock geology.

BEDROCK GEOLOGY

The older rocks in the area (?Paleozoic) are generally schistose or gneissic, and exhibit a shallowly inclined, high-strain regional foliation (S_T), formed through transposition of bedding (S_0), unit contacts, an earlier foliation (S_1), and minor quartz veins. Later folds steepen the foliation at the local scale, and impart large open warps at the map scale. Younger intrusive rocks (mainly (?) Mesozoic) are massive to only weakly deformed. Young clastic and volcanic rocks ((?) late Cretaceous–Tertiary) unconformably overlie Paleozoic and Mesozoic rocks. Most geological boundaries on the map are portrayed as relatively straight and simplistic, due chiefly to the poor quality of bedrock exposure. It is difficult to accurately locate the trace and ascertain the dip of most contacts.

Until regional correlations are more thoroughly understood, the unit names of Bostock (1942), Tempelman-Kluit (1974), and Mortensen (1996), are discontinued in favour of a more descriptive classification (i.e. Ryan and Gordey, 2001a, b; 2002); however, we retain the name for the well established Carmacks Group. Apart from the relatively young intrusive, sedimentary, and volcanic rocks, the stratigraphic succession remains uncertain, and awaits ongoing geochronological study. The mapped geology for 2000 and 2001 is illustrated on the associated poster; more comprehensive descriptions are outlined in Ryan and Gordey (2001a, b; 2002). General rocktypes

are as follows:

Metasedimentary rocks (units 1–5)

Unit 1 - grey to white, banded quartzite

Unit 2 - metaconglomerate

Unit 3 - mica-quartz schist and paragneiss of psammitic, semipelitic, and rare pelitic composition

Unit 4 - quartz-mica (muscovite and biotite) schist,

Unit 5 - metacarbonate

Metavolcanic and volcanoclastic rocks (units 6–8)

Unit 6 - amphibolite schist and/or gneiss (hornblende-plagioclase or garnet-hornblende-plagioclase±quartz±epidote); interpreted as being derived from mafic to intermediate metavolcanic rocks

Unit 7 - quartz-sericite schist (metafelsite); interpreted as possibly derived from felsic volcanic rocks or hypabyssal intrusions

Unit 8 - mafic schist composed of biotite-hornblende ± plagioclase-quartz

Orthogneissic rocks (units 9–11)

Unit 9 - intermediate to mafic orthogneiss

Unit 10 - felsic to intermediate orthogneiss

Unit 11 - potassic feldspar augen granitic orthogneiss

Mafic and ultramafic rocks (unit 12)

Unit 12 - diverse assemblage of mafic to ultramafic rocks of uncertain origin

Younger plutonic rocks (units 13–16)

Unit 13 and 14 - monzogranite, granodiorite, quartz monzonite, and locally gabbro intrusions (Permian or Jurassic?)

Unit 15 - chlorite-altered hornblende to biotite-bearing granodiorite and monzogranite, quartz monzonite, and monzodiorite (Jurassic?)

Unit 16 - syenogranite to monzogranite plutons, and/or dykes (mid-Cretaceous, Late Cretaceous, or Tertiary?)

Cretaceous–Eocene rocks (units 17–19)

Unit 17 - conglomerate below Carmacks Group (Cretaceous?)

Unit 18 - volcanic rocks of the Carmacks Group (Late Cretaceous)

Unit 19 - quartz and/or potassic feldspar phyric rhyolite (Eocene?)

DEFORMATIONAL STRUCTURES

Regional foliation and folds

Almost all probable Paleozoic rocks in the field area exhibit the regional foliation (S_T), characterized by high-strain transposition of layering in gneiss and schist, with abundant intrafolial isoclinal folds that are commonly rootless Ryan and Gordey (2001a, 2002). The intensity of strain within the regional foliation locally grades to mylonite. Primary compositional layering (S_0) in metasedimentary rocks, unit contacts (e.g. dyke margins), and a pre-existing foliation (S_1) can be recognized around closures of the transposition folds, indicating that they are at least F_2 structures. The F_2 folds are generally recumbent to shallowly inclined, close to isoclinal, long-wavelength structures. Large F_2 folds are well outlined in quartz-rich metasedimentary rocks west of Black Hills Creek, where a strong axial-planar S_2 is present, parallel to the regional S_T . As pointed out by Ryan and Gordey

(2001a) in the Thistle Creek area, F_2 structures commonly lack an axial planar foliation. A regional extension lineation (L_2) is parallel to the F_2 axes. This relationship helps distinguish F_2 and F_3 folds, which can have very similar style (Ryan and Gordey, 2001a). F_3 folds are open, moderately inclined (with shallow to steep varieties), shallowly plunging structures, that have weak axial-planar fabric where developed in schistose layers, and have no associated extension lineation

Faults

The map area is affected by faults of various significance. Most of these could not be observed directly, but are interpreted from changes in rock type, and/or structural grain. Some of these are also well delineated by prominent physiographic and aeromagnetic lineaments. Locally, fault breccia and slickensides provide direct evidence of fault contacts (e.g. Ryan and Gordey, 2001a, 2002). Mapping across the White and Yukon rivers shows no appreciable change in bedrock composition across these valleys, ruling out any significant offset along them. Future mapping will test the significance of the pronounced lineament trending northwestward from Scroggie Creek, through the Stewart, Yukon, and Sixty Mile rivers.

METAMORPHISM

Quantitative study of metamorphic grade has not yet been done, and little additional understanding of regional metamorphism has been gained since that reported in Ryan and Gordey (2001a, c). Mafic rocks in the area typically exhibit the middle amphibolite-facies assemblage garnet+hornblende. Other than the presence of kyanite in the Grizzly dome area, the lack of aluminum silicate minerals in the aluminous metasedimentary rocks remains puzzling.

DISCUSSION

Tectonostratigraphy and paleotectonic setting

Ryan and Gordey (2001a) reported that metavolcanic (amphibolite) and associated orthogneissic rocks were much more widespread than quartz-rich metasedimentary rocks in the Thistle Creek area; however, this year's work demonstrated an increase in the proportion of quartz-rich metasedimentary rocks away from Thistle Creek. Despite the high degree of strain in these rocks, and clear local evidence for structural interleaving, numerous localities demonstrate the interstratification of the metavolcanic and metasedimentary rocks. At the map scale, the metavolcanic rocks are structurally higher than the quartz-rich metasedimentary rocks, but stratigraphic way up cannot be proven. Mafic to intermediate orthogneiss units are predominantly associated with the amphibolite, and rarely with the metasedimentary rocks. We believe that this lithological distribution is consistent with a volcano-plutonic complex, of possible island-arc affinity (Ryan and Gordey, 2001c), built upon a detrital quartz-rich (possibly continental) margin. Regional constraints indicate that this arc was probably of Devonian-Mississippian age (Mortensen, 1992); ongoing geochronological studies will resolve this question. The age of the marginal metasedimentary rocks are unknown ((?)Proterozoic, (?) early Paleozoic, (?)mid-Paleozoic).

Complexes of mafic-ultramafic rock and associated orthogneiss (unit 12) are exposed at structurally high levels, and are thought to be part of an erosionally dismembered, originally continuous tectonic sheet, that lies at the contact between the overlying metavolcanic rocks and underlying metasiliciclastic rocks. These rocks were together prior to regional transposition deformation and metamorphism. Some small occurrences of the ultramafic rocks are clearly

lozenges in high-strain zones, and the strain has removed most evidence of their original disposition. It is possible that they merely represent intrusive complexes (possibly Alaskan type) into the quartz-rich metasedimentary rocks, and were broken up and segmented during high-strain transposition of the stratigraphy during orogenesis. An alternative interpretation is that they represent dismembered portions of an originally more extensive structural sheet that was thrust on top of the metasedimentary rocks. Their internal compositional heterogeneity is reminiscent of dismembered ophiolitic fragments. Likewise, the 90° angle between pre-existing lineation in the talc-siderite schist, and the regional lineation in the surrounding schist near Shamrock Dome, is a relationship similar to that in peridotite that preserves deformation prior to ophiolite emplacement (e.g. Lewis Hills massif, western Newfoundland (Suhr and Cawood, 1993)).

Economic implications

Ryan and Gordey (2001a) reported that the Thistle Creek area is dominated by a volcano-plutonic (?) arc complex with implied potential for volcanogenic massive sulphide-type mineralization, and noted that metafelsite units were rare. No additional metafelsite units were mapped in 2001, and siliciclastic rocks were found to be more abundant than metavolcanic rocks in the western portion of the area. By this reasoning the area mapped east of the Yukon River remains the most prospective for massive sulphide mineralization.

Mid-Cretaceous (105–90 Ma) and Late Cretaceous (70–65 Ma) plutons and their country rock are still considered prospective targets for intrusion-related gold deposits in Yukon Territory and Alaska (e.g. Hart et al., 2000). We have mapped a number of young-looking granite bodies in the southern Stewart River area, and have made an attempt at dividing them into probable Jurassic and Cretaceous suites. We believe that all of these bodies have potential for occurrences of intrusion-related gold.

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Volcanology and geochemical signature of Late Paleozoic volcanic sequences: The Klinkit Formation, an example from northern British Columbia and southern Yukon

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The Canadian Cordillera is known to be a collage of oceanic and arc terranes that were accreted to the western margin of the North America craton during the Mesozoic. Upper Paleozoic magmatic sequences are an important part of these terranes. However, the understanding of the tectonomagmatic evolution of these terranes is limited, in part due to incomplete knowledge of the stratigraphy and geochemical composition of these sequences. An understanding of the geological and geochemical evolution of these sequences is essential for paleogeographic reconstruction, as well as for comparison and correlation of the various time-correlative magmatic suites in northern British Columbia and southern Yukon.

The Mississippian to Permian Klinkit Formation in northern British Columbia and southern Yukon, straddles several inboard terranes, both exotic and pericratonic, of the northern Canadian Cordillera. This well preserved volcano-sedimentary sequence provides a rare opportunity to document a pre-accretion history of this part of the Canadian Cordillera.

The Klinkit Formation, a Late Carboniferous to early Permian age volcanic-dominated sequence, conformably overlies the Screw Creek Limestone, a prominent light-coloured reef and debris-flow carbonate of Bashkirian age (Late Carboniferous-Pennsylvanian), and the Swift River succession, a dominantly sedimentary pile of rocks including dark-colored chert and argillite interspersed with quartzite-phyllite sequences, quartz wacke, clean quartzite and local pebble conglomerate. The Klinkit rocks are unconformably overlain by the Triassic Teh succession, a sedimentary unit characterized by interbedded black argillite, dark siltstone, sandstone and chert, with minor limestone and pebble conglomerate of continental affinity.

The Klinkit Formation is informally subdivided into: (1) the volcanoclastic unit, a >250 m thick dominantly volcanic pile with interbedded sediments, overlain by (2) the thin Bigfoot unit, a mixed volcanic and sedimentary sequence which contains upward increasing sandstone and dark argillite. These sequences are megaturbidites typical of a subaqueous, below-wave-base depositional setting.

The Klinkit volcanic rocks are calc-alkaline with characteristics of a mature island-arc suite including $(La/Yb)_n \sim 2.8-4.7$, negative Nb and Ti anomalies on the mantle-normalized trace element plots and positive ϵ_{Nd} values. The formation is interpreted as the distal facies of an arc system.

The Klinkit Formation is similar in composition and age to several Paleozoic arc successions of the exotic Quesnel terrane (e.g., the Lay Range Assemblage of the Harper Ranch Subterrane, northern central British Columbia; Boswell-Semenoff Hills, central Yukon), and of the pericratonic Yukon-Tanana terrane (Little Salmon succession, central Yukon).

Detailed geochemical study, associated with extensive geological mapping and physical volcanology, of the Paleozoic volcanic suites in the present assemblage of distinct pericratonic and exotic terranes in the northern Canadian Cordillera could provide the magmatic link to indicate that a large Paleozoic arc system was structurally dismembered prior to or at some stage in its emplacement on the western margin of the North American craton.

**Modeling and Interpretation of Magnetotelluric Data:
Watson Lake to Stewart (Line 2a) and Johnson Crossing to Watson Lake**

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INTRODUCTION

In the summer and fall of 1999, a regional scale magnetotelluric survey (MT) was completed within northern British Columbia and southern Yukon as part of the fifth phase of the LITHOPROBE program. Ledo et al., (2002a) describe the geoelectric structure of the Tintina Fault zone and Ledo et al., (2001b) describe MT results for Corridor 3. A total of 85 MT stations were deployed along the Alaskan Highway and the British Columbia Highway 37 on SNORCLE Corridor 2 (Fig. 1). This corridor crosses a geological region that has experienced substantial deformation and tectonism, which makes the area appealing for examining large-scale tectonic elements and their relationships. This study will focus on analysis of MT data from line 2a, which is orientated roughly north-south between Stewart and Watson Lake, and from the Connector Line, which runs between Watson Lake and Corridor 3. These lines cross a similar sequence of rocks west of the Tintina Fault zone and are defined by a total of 54 MT sites.

MAGNETOTELLURIC METHOD

The magnetotelluric method uses natural EM fields to investigate the electrical conductivity structure of the Earth's crust and upper mantle. By measuring EM field variation with time, interactions between the fields are defined. The ratio of the electrical and magnetic field strengths, as a function of frequency, yields the electrical impedance, which is used to calculate the resistivity structure with depth. Measured amplitude, phase, and directional relationships between the electric and magnetic fields depend on the distribution of the local conductivity structure (Vozoff, 1991). The depth of penetration of the fields is also dependent on the local conductivity structure, and the periodicity of the EM wave. The MT method can provide unique information on the geodynamic processes because the MT response is sensitive to different geological parameters than from other geophysical methods. MT is particularly sensitive to the presence of saline fluids, mineralization, the presence of graphite, and partial melting.

MT DATA PROCESSING

Processing of the MT data involved visual inspection of the time series, interactive editing of cross-powers, and calculation of MT parameters using robust impedance estimation methods (Larson et al., 1996; Chave and Thompson, 1989; Chave and Smith, 1994). The geoelectric strike of the regional structures was determined using Groom Bailey (1989) and McNeice and Jones (2001) theory. The results indicated a regional geoelectric strike of -45° (Wennberg et al., 2001, Wennberg 2002). This

direction agrees well with the surface geologic strike and has been adopted as the strike direction in subsequent MT modeling and inversion.

MODELING

Two-dimensional MT inversions were done using a non-linear conjugate gradient method developed by Rodi and Mackie (2001). Figures 2 to 5 show the resulting models for Line 2a and the Connector Line. The models reveal a number of significant geoelectric structures with correlation of some structures between the two lines and also with Corridor 3 (Ledo et al., 2002). The models include a number of very shallow structures at <3 km depth which are shown more clearly in Figures 2 and 4. The structure at this depth is spatially aliased by the 10 to 20 km site spacing of the MT survey but the MT results provide a sample of the geoelectric structure in most geological terranes. The structures observed at greater depth in the models (Fig. 3, 5) will be more representative of the true crustal structure. The existence and geometry of these structures has been confirmed using multiple two-dimensional inversion and forward modeling of the MT data from each line.

Line 2a

The MT model for Line 2A is characterized by two shallow upper crustal conductors and a number of larger structures in the underlying crust. Narrow (<20 km) conductive zones exist beneath the Cache Creek (labeled **X**) and Slide Mountain terranes (labeled **Y**) in the shallow upper crust (<3 km depth) (Fig. 2). The resistivity in these zones reaches values as low as 2 ohm.m.

The deeper crustal structure is dominated by a large resistive (10,000 ohm.m) wedge beneath the northern Stikinia Terrane and adjacent terranes to the north. This body is labeled **A** on figure 3. Beneath the Stikinia Terrane the body extends from the surface to ~30 km depth. Further north, the upper surface dips to the northwest, reaching a depth of ~10 km beneath the Slide Mountain Terrane. The result suggests that the lower surface of the body dips to the southeast, but the northern part of this contact is obscured by an overlying conductor (labeled **B** on Fig. 2,3).

The conductor **B** forms an elongate body beneath the Cassiar terrane. It extends approximately 60 km along the transect and from 7 to 17 km depth. The upper surface of the body is approximately flat at its southern end but dips to the northwest beneath the centre of the Cassiar Terrane. The lower surface of the body also dips to the northwest.

At the southern end of the profile there is a northwest-dipping conductor, labeled **E** in figure 3. This structure is most conductive at the end of the profile, where the resistivity reaches values less than 1 ohm.m, but a less conductive (<50 ohm.m) extension of the body dips beneath the Stikinia terrane. A similar dipping body is noted at the southwestern end of the MT model for Corridor 3 (Wolyniec, 2001, Ledo *et al.* 2002). The depth of the conductor on Line 2a (>10 km) is more shallow than the depth on Line 3 (>20 km).

Connector Line (Watson Lake to Johnson Crossing)

The MT model for the connector line also includes a shallow upper crustal conductor. There is a very conductive zone (<1 ohm.m) at depths of less than 2 km beneath the Dorsey Terrane (labeled **Z** on Fig. 4).

The deeper structure (Fig. 5) contains resistive zones extending from near the surface to 12 km depth beneath the Slide Mountain and Kootenay terranes (labeled A) and from the surface to 30 km depth beneath the southern segment of the Slide Mountain Terrane and northern Cassiar Terrane (labeled A'). Based on a strike azimuth of -45° these resistive zones are approximately along strike from the resistive zone in Line 2A.

Lying between the two resistive zones on the Connector Line, beneath sites 281, 280, and 279 is a strong conductor (labeled F on Fig. 3,5). The top surface of the conductor occurs at about 3 km depth beneath the northern Dorsey Terrane but dips to around 15 km depth to the north and south. This conductor does not correlate with any other conductors on Line 2a or Line 3 models.

The Connector Line includes a conductor at ~ 10 km depth within the Cassiar terrane (labeled B). This conductor appears to correlate with the conductor noted in the Cassiar Terrane on Line 2A, and the MT model for Corridor 3 also includes a conductor in this terrane.

INTERPRETATION

Uppermost Crust

Previous studies have suggested that the Dorsey Terrane, which is composed mostly made up of sandstone, argillite, or chert rocks, forms a thin surface terrane in the shallow upper crust (Monger et al., 1991). The MT results, which show a surface conductor that is less than 1.5 km thick, support this hypothesis. Because there are only two MT sites in the Dorsey terrane it is not possible to determine whether the whole terrane is composed of conductive rocks or whether there are localized conductors near sites 278 and 279. In the former case the enhanced conductivity could be explained by the widespread occurrence of graphitic or metalliferous rocks. In the latter case the high conductivity may be due to a localized presence of sulphides or oxides. Volcanogenic massive sulphides occur at the Mor mine in the Yukon Territory west of site 279. The associated rocks include a manganiferous metachert horizon containing stratiform magnetite.

The Slide Mountain Terrane, which is composed of mafic volcanics with related ultramafic, mafic, and granitic bodies, and predominantly fine grained clastic rocks and chert, is also interpreted to form a relatively thin surface layer (Monger et al., 1991, Monger, 1989; Evenchick et al., 2001). The metamorphic grade ranges from sub-greenschist facies to staurolite-kyanite grade (Stuick, 1986). A thin conductive zone is present in the Slide Mountain Terrane on Line 2a profile, but there is no corresponding conductor in the Connector Line model. Therefore, the likely cause of the enhanced conductivity at depths (< 2 km) is localized mineralization. The Erickson (gold), Table Mountain (gold), and Cassiar (asbestos and jade) mines are located within 15 km of the conductor at site 232. Rocks of this region have been affected by alteration that could have resulted in precipitation of graphite films on grain boundaries.

Crust and Lithosphere

The geometry of the resistive wedge beneath the Stikinia Terrane and adjacent terranes provides further support for the interpretation that the Stikinia Terrane forms a laterally extensive terrane onto which smaller, thinner accreted terranes are overlain (Evenchick et al., 2001; Cook et al., 2001). The high resistivity of the body is consistent

with the expected presence of mafic/felsic rocks, inter-layered with clastic sedimentary and carbonate rocks, and intermediate-composition plutons (Lowe et al., 1994; Monger et al., 1991). One interesting feature of the MT models is that they define a lower surface of the resistive body. If this lower surface is correlated with the base of the Stikinia terrane it suggests that the terrane may pinch out with an eastern margin near site 233.

The conductors that are present at about 10 km depth in the Cassiar Terrane on Lines 2A and the Connector Line are likely part of the same body. The different orientation of the two lines relative to the geological strike will contribute to the change in shape of the body in the two models. The conductor in the Cassiar Terrane on Line 3 occurs at a greater depth, around 20 km, than the corresponding conductors on Corridor 2. The results suggest the presence of large-scale sources of enhanced conductivity in the terrane. It is possible that the source of the enhanced conductivity is deformed metasedimentary rocks.

A northwest dipping conductor is present at the southwest ends of Corridors 2 and 3 that may be related to processes associated with the subduction of the Farallon plate beneath North America. Although the conductor occurs at different depths in the two profiles, its geometry is remarkably similar in the two models. Modeling of gravity data across the western margin of the continent using constraints from SNORE line 22 and ACCRETE seismic data shows a deepening of the Moho with distance inland from the continental margin (Hammer and Clowes, 2000). This geometry suggests that the conductor is present in the lower crust. At this time, additional inversions are needed to better constrain the depth of the conductive slab in order to determine its relationship to subduction processes.

A dipping conductor is also observed beneath the actively subducting Juan de Fuca Plate in the southern Cordillera (Kurtz et al., 1990). The source of the enhanced conductivity at this location has been interpreted to be saline pore fluids supplied by water subducting with the oceanic crust and by dehydration reactions. In the northern Cordillera, subduction ceased around 20 Ma. If the conducting slab observed on Corridors 2 and 3 is indeed associated with this subduction, it suggests that the enhanced conductivity associated with subducting slabs may be due to the presence of sediments subducted with the slabs or metamorphic reactions in the overlying crust associated with fluids released during subduction rather than the presence of the actual fluids.

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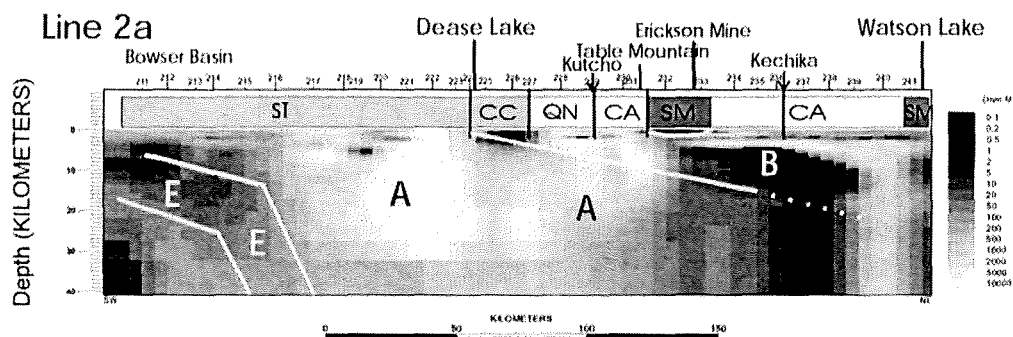


Figure 3. MT model for Line 2a showing the crust and lithosphere response.

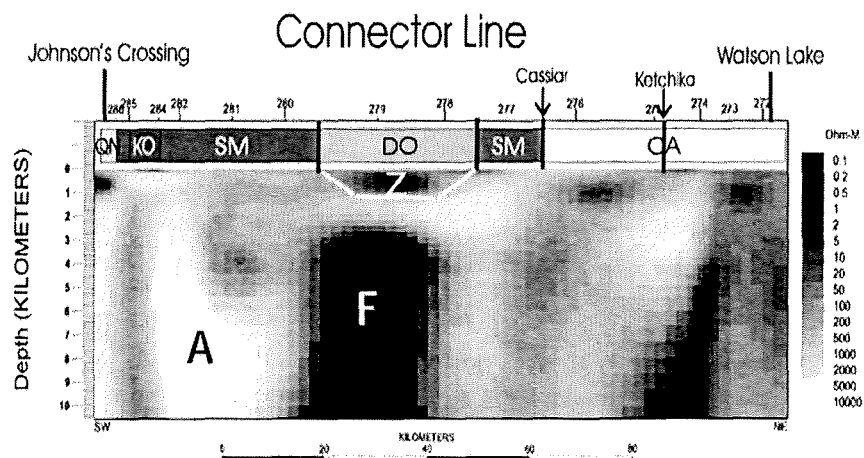


Figure 4. MT model for the Connector Line showing the upper crust response.

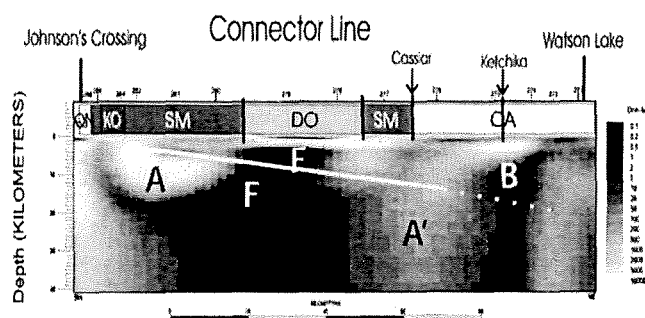


Figure 5. MT model for the Connector Line showing the crust and lithosphere response.

Geological Interpretation of electrical resistivity models along the SNORCLE Corridors 1 and 1A

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Introduction

Magnetotelluric (MT) soundings completed along LITHOPROBE SNORCLE transect Corridor 1 and 1A cross the Proterozoic Nahanni, Fort Simpson, Hottah, Great Bear Magmatic Arc, Hay River and Buffalo Head terranes, the Archean Slave Province and the Great Slave Lake shear zone (GSLsz). Phanerozoic sedimentary rocks overlie most of the terranes except the Slave Province (Fig. 1). The position of the GSLsz on geological maps (*e.g.*, Hanmer *et al.* 1992) is between sites 153 and 154, but the centre of the magnetic low with which the shear zone is correlated lies between sites 155 and 156. Examination and analysis of the MT responses (apparent resistivity and phase) has been described previously (Wu *et al.* 2000a, 2000b; Wu 2001). This paper will focus on the analysis of geoelectric dimensionality and strike and the geological interpretation of MT models.

Geoelectric Strike Angle

The determination of geoelectric dimensionality and strike involved analysis of the MT impedance tensors using Groom-Bailey decomposition and maximum phase difference methods and analysis of induction vectors. Figures 2 and 3 show summary diagrams of the geoelectric strike determined using Groom-Bailey decomposition. The results indicate the variation of geoelectric strike angle with period is remarkably similar for Corridors 1 and 1A.

The geoelectric response is approximately one-dimensional at short periods (<0.2 s), which correspond to signal penetration into the Phanerozoic sedimentary rocks. The response has a weak two-dimensionality which, considering the inherent 90° ambiguity in the determination of strike from the MT data, indicates a strike angle of either N60°E or N30°W (Fig. 2, 3). A N30°W strike angle is consistent with the gentle dip of the Phanerozoic rocks towards the southwest.

The geoelectric response indicates a two-dimensional response with a strike of ~N30°E-N35°E in the upper and mid crust, and a two-dimensional response with a strike of N60°E in the lower crust to lithospheric mantle (Fig. 2, 3). There is a suggestion that the geoelectric strike rotates closer to east-west at the longest periods (>2000 s) but this response is relatively poorly resolved by the data.

The geoelectric strike angle along Corridor 1A was re-examined using the McNeice-Jones (McNeice and Jones, 2001) multi-site, multi-frequency Groom-Bailey analysis (Fig. 3b). This analysis confirms that the strike at periods of 0.1 to 20 s averages N33°E and at 20-10,000 s averages N62°E. The

transition between these two directions initiates in the period range 10-20 s, which is approximately the period range for sensitivity of the MT signals to the base of the 15-25 km thick crust. The direction in the upper crust has been interpreted to represent the local-scale (<50 km) horizontal strike of the GSLsz and the direction in the mantle is parallel to the larger-scale strike of the GSLsz (e.g. Wu et al. 2002). However, the similarity of the strike angles on Corridors 1 and 1A suggests the strike in the crust and mantle might reflect more pervasive structures associated with the collision of the Slave and Rae Provinces and the collision of the Hottah terrane and Slave Province.

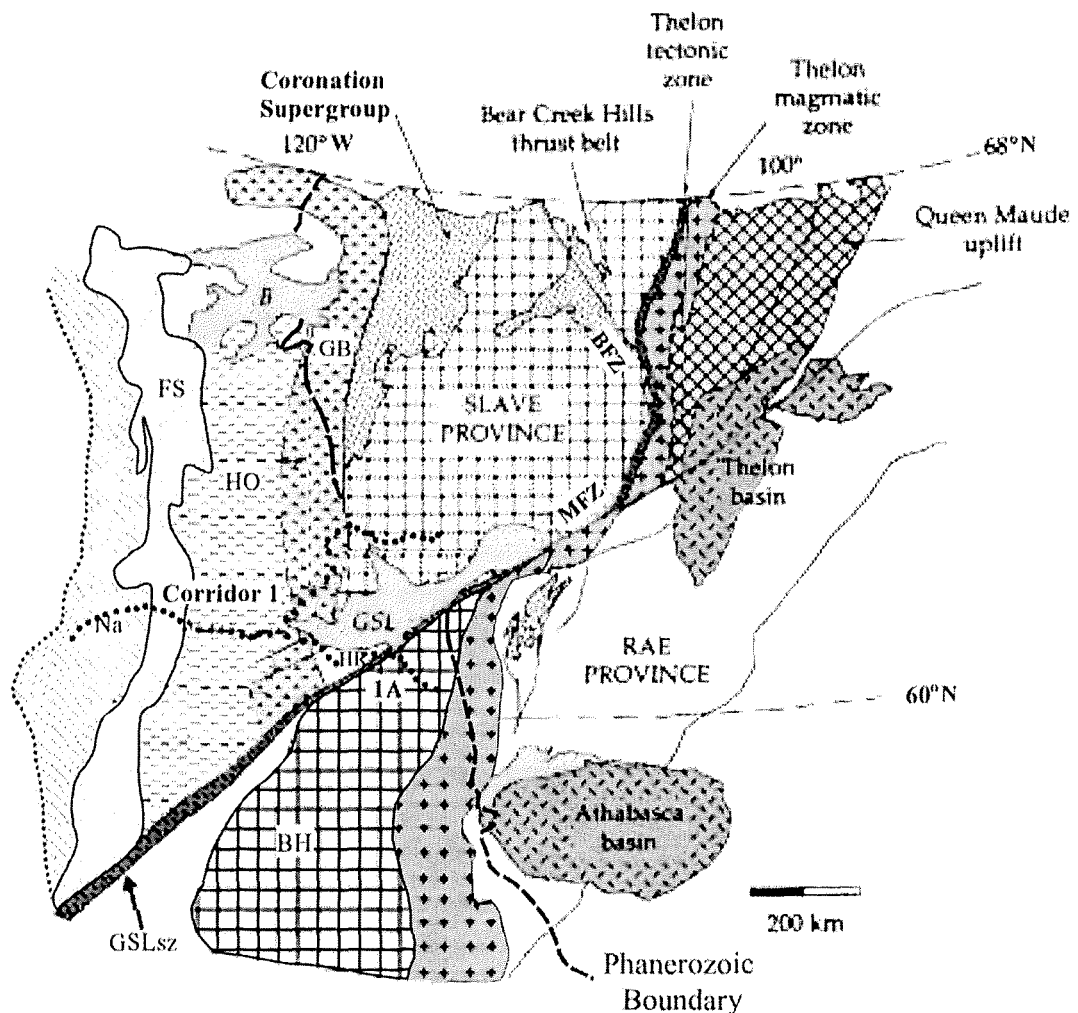


Figure 1 Selected tectonic elements and structures of the northwest Canadian Shield (after Hanmer, 1988). Na: Nahanni terrane, FS: Fort Simpson terrane, HO: Hottah terrane, GB: Great Bear magmatic arc, HR: Hay River terrane, GSLsz: Great Slave Lake shear zone, BH: Buffalo Head terrane, BFZ: Bathurst fault zone, MFZ: McDonald fault zone. The circles show the location of MT sites on Corridor 1 and 1A.

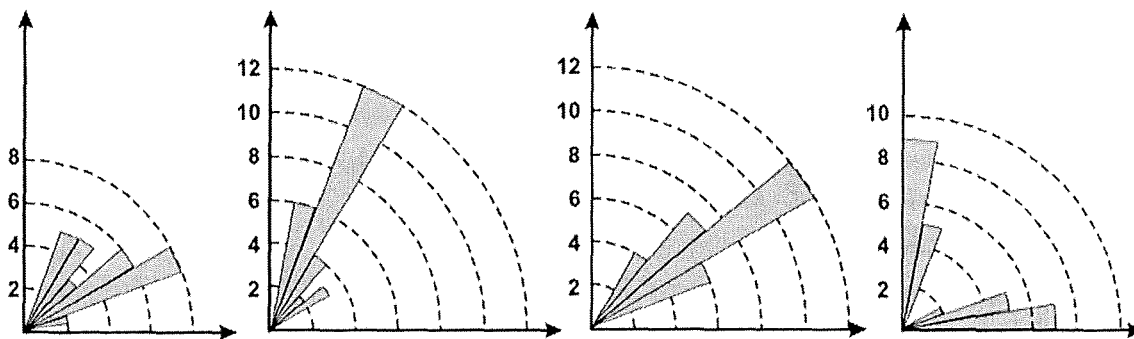


Figure 2 Rose diagram of the Groom-Bailey regional strike angle for four period ranges based on 29 sites located in Nahanni terrane to Great Bear magnetic arc along Corridor 1. The angle bin is 10° . The results for band 1 and 3 include 29 resolved strikes angles, the results for Band 2 include 27 strikes, and for Band 4 include 28 strikes.

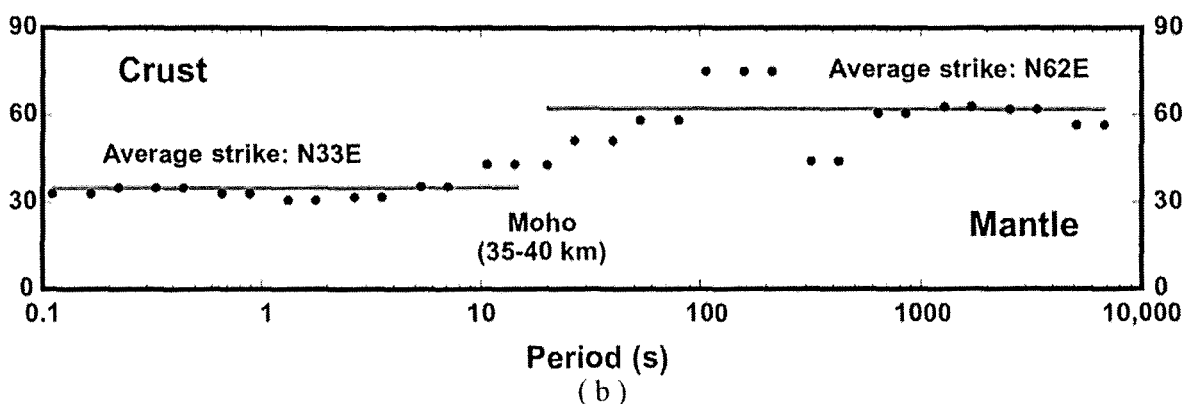
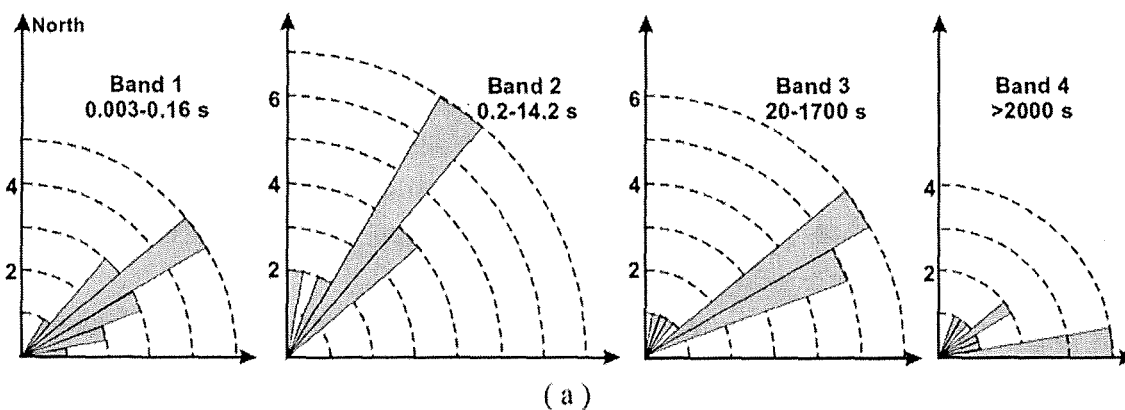


Figure 3 (a) Rose diagram of the Groom-Bailey regional strike angle for four period ranges based on 15 sites along Corridor 1A. The angle bin is 10° . The results for the first two period bands include 15 resolved strikes angles, the results for Band 3 include 14 strikes, and for Band 4 include 10 strikes. (b) Geoelectric strike angle from multi-site, multi-frequency extended GB decomposition that best fits all sites along Corridor 1A simultaneously.

Structure of Phanerozoic sedimentary rocks

The Phanerozoic sedimentary rocks form an electrically conductive layer in the MT models. The sedimentary rocks have a maximum thickness of ~1000 m above the western Fort Simpson terrane and thin in a north-northeast direction. Around site 142, the conductive surface layer is thinner than in surrounding areas. This feature is interpreted to be the Liard High, a structure formed during the Cambrian and early Ordovician (Law 1971; Merjer-Drees 1975).

In the southeastern part of Corridor 1 and along Corridor 1A the geoelectric structure consists of three layers: a relatively resistive first layer and a conductive second layer overlying resistive basement (Fig. 4b). There is a clear resistivity boundary near sites 152-153 where the profile crosses the Presqu'île Barrier. This structure hosts the major Mississippi Valley type (MVT) Pb-Zn Pine Point deposit.

To the northwest of the transition, the rocks consist of a moderately resistive upper layer (20-100 m) overlying the conductive lower layers. The resistive layer is interpreted to be associated with the overburden and/or Upper Devonian bioclastic and sandy limestone (Hills et al. 1981, Douglas and Norris 1973). The conductive lower layer can be subdivided into a more conductive ($<10 \Omega\text{m}$) lower unit subcropping between sites 148 and 153 and a more resistive ($10\text{-}25 \Omega\text{m}$) upper unit subcropping to the northwest of site 147. The lower unit corresponds to the thick dark grey and black shale sequences of the Upper Devonian Horn River Formation (Douglas & Norris 1973) and the upper unit corresponds to the Upper Devonian Twin Falls, Tathlina, and Redknife Formations which include bioclastic and reef carbonates as well as sandy limestone, shale and siltstone.

To the east of site 153, the two-layer structure is more resistive than to the northwest. The relatively resistive upper layer (350 m) is interpreted to correspond to the middle Devonian carbonate and evaporate dominated units including the Nyarling, Little Buffalo, and Chinchaga Formations. The conductive second layer (~300 m) corresponds to the Middle Ordovician (?)–Middle Devonian Mirage Point Formation which includes red and green shale beds, carbonate, sandy carbonate and gypsum (Glass, 1990; Rhodes, *et al.* 1984; Douglas & Norris, 1973) and the underlying Ordovician Old Fort Island Formation which consists of quartz sandstone, siltstone and shale. The greater thickness of the lower unit in the model than in the cross-section may be due in part to the divergence between the eastern end of the Corridor 1a and the location of the cross-section.

Precambrian Crustal Structure

Figure 5 shows the two-dimensional MT models for Corridors 1 and 1A which reveal a number of significant geoelectric structures. The existence and form of these structures has been confirmed using re-examination of the raw MT responses, multiple two-dimensional inversions, and one-dimensional and two-dimensional forward modelling and sensitivity studies (Wu 2001).

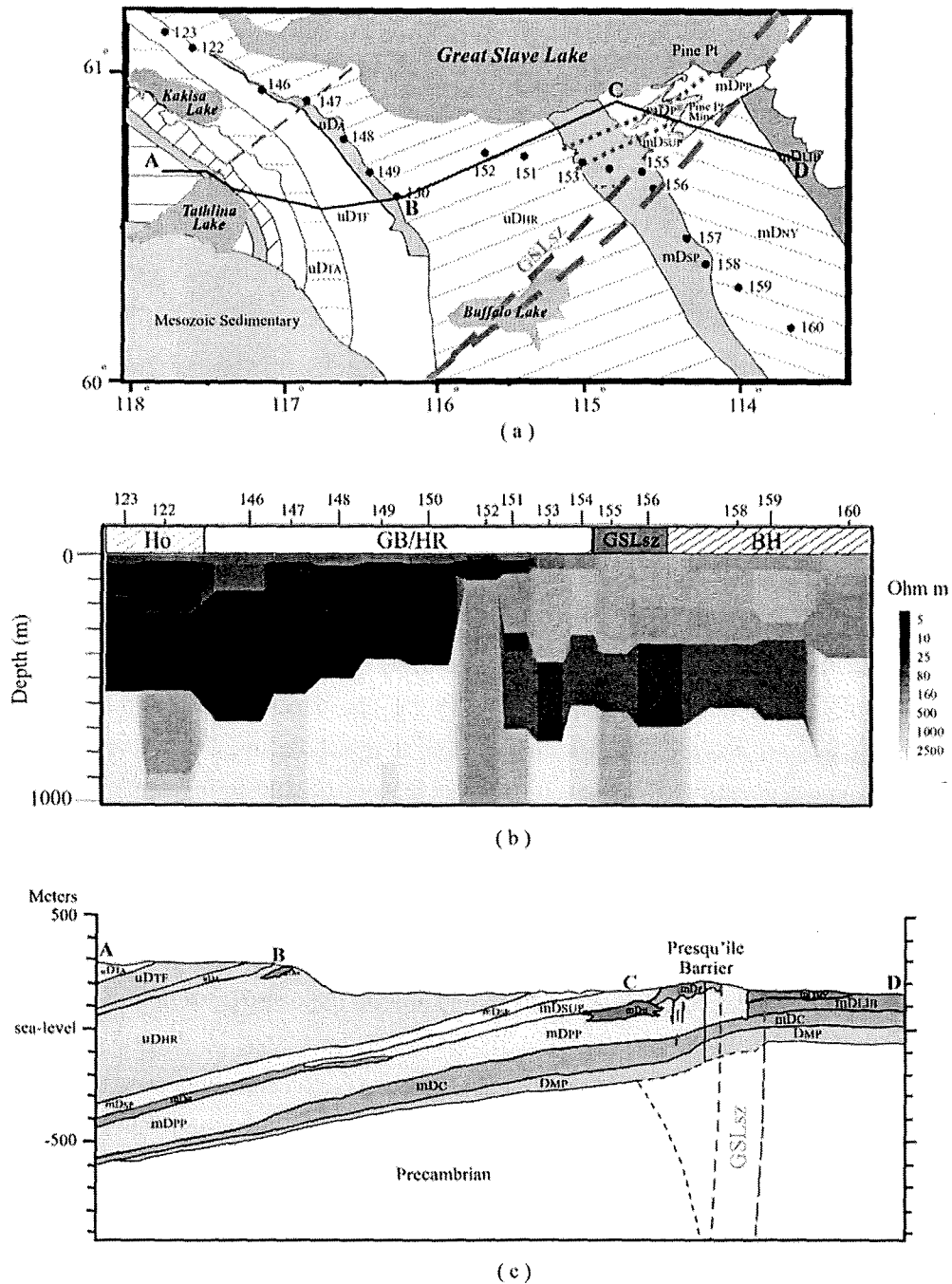


Figure 4 (a) Location of sites along Corridor 1A. The location of the GSLsz is shown by dashed lines and is inferred from magnetic field data. The dotted line shows the location of the Pine Point (Presqu'ile) Barrier. The mine symbol shows the location of the Pine Point. (b) Stitched 1D inversion models along Corridor 1A. (c) Geological cross-section (A-D) of Phanerozoic rocks. Modified from Douglas and Norris (1973) and Douglas (1973). Geological units: uDER: Escarpment Member, Hay River Formation (bioclastic and reefy limestone beds), mDC: Chinchaga Formation (laminated gypsum, salt, limestone and dolomite), DMP: Mirage Point Formation (dolomite, sandy dolomite, shale and gypsum).

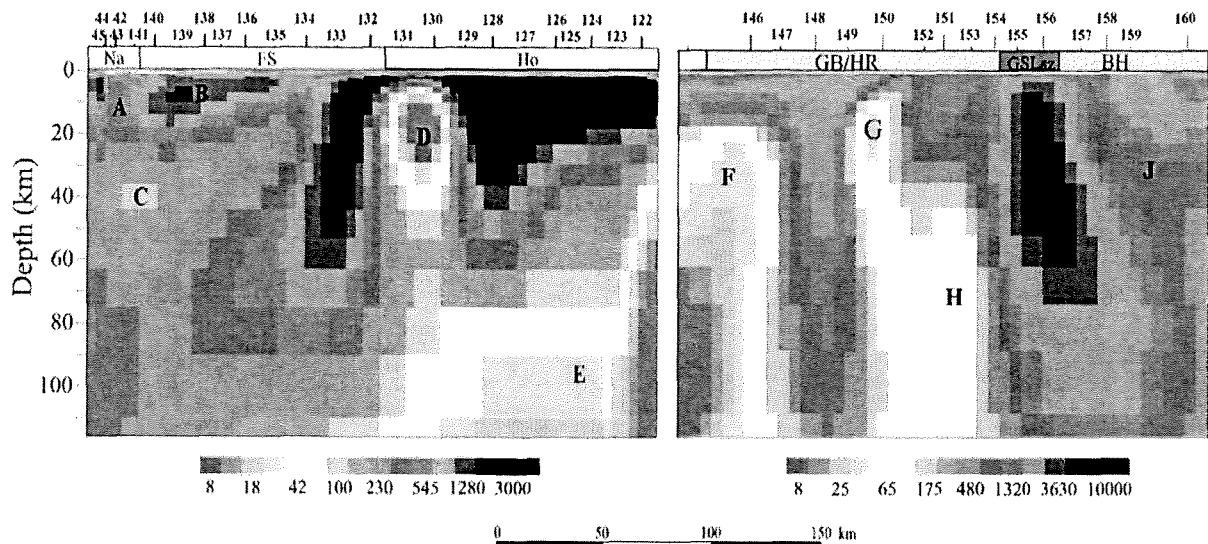


Figure 5 2D inversion models along Corridor 1 and 1A.

Fort Simpson Basin: The MT models image a relatively resistive ($>400 \Omega\text{m}$) body with a west-dipping base in the Fort Simpson terrane (labelled B in Fig. 5). The maximum depth to the base of this body, which extends at least 100 km along the survey line, is about 20 km. Seismic reflection data shows a package of west dipping (20° - 30°) reflections beneath stations 0-2500 corresponding to the location of MT sites 142-132 (Cook et al. 1998). The MT results support the interpretation of the seismic reflection data in terms of the Fort Simpson basin. This basin is interpreted to have formed as a result of lithospheric extension following collision of the Fort Simpson terrane with the western Hottah terrane at ~ 1.84 Ga. The relatively high resistivity of the rocks in the basin is consistent with the presence of the Proterozoic sedimentary rocks including sandstone, siltstone and argillite deposits (Cook et al. 1998; Hills et al. 1981).

Fort Simpson-Hottah conductor: The MT models reveal a clear conductive zone ($<100 \Omega\text{m}$) under sites 132 to 130 (labelled D in Fig. 5) at the boundary of the Fort Simpson and Hottah terranes. This zone extends from several kilometres depth to ~ 40 km depth. It is spatially correlated with a wedge-shaped zone defined by seismic reflection results that extends from several kilometres to ~ 25 km depth between stations 1800 and 3200 (Fig. 6).

This zone has been interpreted to be a structural boundary between the Fort Simpson terrane with the western Hottah terrane. According to Snyder (2000), the older Hottah terrane (1.95-1.91 Ga) collided into the eastern flank of the younger Fort Simpson terrane (1.8 Ga) creating wedged shape geometry. The upper crust of the Fort Simpson terrane was detached and thrust over the Hottah terrane and the ocean lithosphere that separated the Fort Simpson terrane from the Hottah terrane was subducted beneath the crust of the Hottah terrane.

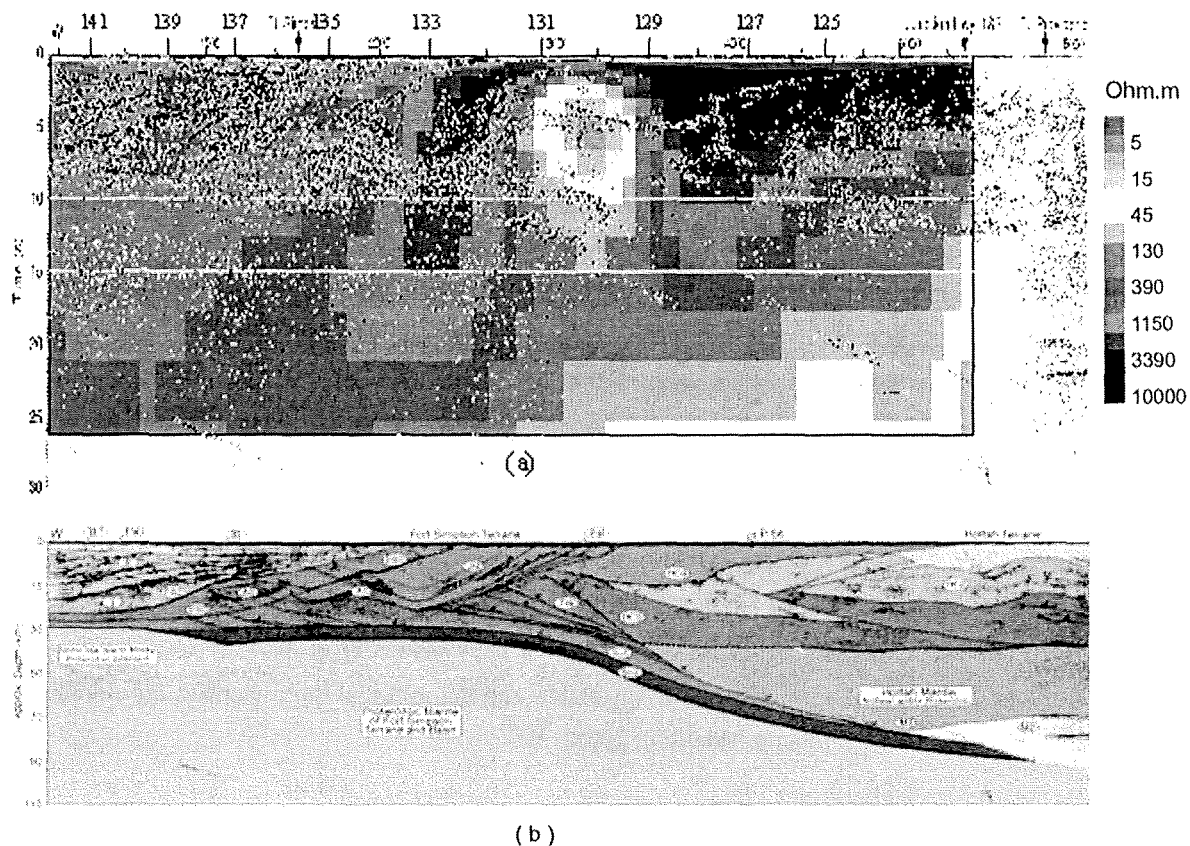


Figure 6 (a) Seismic reflection profile and 2D MT inversion model from Fort Simpson to Hottah terrane along Corridor 1. The figure shows seismic data overlying the resistivity model. (b) Geological interpretation of seismic reflection data (Cook et al., 1998).

The enhanced conductivity is spatially restricted to the crustal depths, to the point of the wedge, and to the Hottah side of the boundary. This geometry suggests the enhanced conductivity is directly related to metamorphism and deformation of Hottah rocks during the collision. Increased temperature and pressure may have produced graphitic schists and gneisses from sedimentary rocks in the Hottah crust. Fluids released by the subducting oceanic lithosphere may also have contributed to the processes that lead to enhanced conductivity.

Hottah-Great Bear Conductor: There is a conductive zone ($\sim 30 \Omega\text{m}$) beneath the boundary of the Hottah terrane and Great Bear Magmatic Arc with its top at about 20 km depth (labelled F in Fig. 5). The Great Bear Magmatic Arc is interpreted to have been a product of eastward subduction of oceanic lithosphere beneath the Hottah terrane at 1.84-1.87 Ga. Cook *et al* (1999) suggest the Great Bear Magmatic Arc is relatively thin ($\sim 3\text{-}4.5$ km) and lies above either Hottah crust or imbricated rocks of the Coronation margin.

Although the precise nature of the rocks beneath the Great Bear Magmatic Arc is not known, it is known that they are deformed rocks of the Hottah-Slave transition (Cook, *et al.*, 1999), for example, deformed and metamorphosed Coronation Supergroup, a continental margin (back-arc) depositional prism. Therefore, the enhanced conductivity in the middle and lower crust beneath the boundary between the Hottah terrane and Great Bear Magmatic Arc may be due to either graphite or conductive minerals concentrated during the deformation and metamorphism of Coronation Supergroup rocks.

There is also conductive crust beneath the central Great Bear Magmatic Arc (labelled G in Fig. 5) with its upper surface at around 10 km depth. Because this structure is further from margin of the Great Bear Magmatic Arc its source is less clear than that of the conductor to the northwest. However, it can again be attributed to electronic conduction in interconnected metasediments of either Hottah terrane or Coronation Supergroup.

Great Slave Lake shear zone: The MT inversion models reveal that the GSLsz forms a crustal-scale resistive zone ($>5000 \Omega\text{m}$, labelled I in Fig. 5) that is spatially correlated with a magnetic low. The GSLsz comprises greenschist to granulite facies mylonites. Its high resistivity is interpreted to be due to the resistive nature of the granitic protolith of the mylonites and the fact that the mylonites are dominated by rocks deformed in the ductile regime (Wu *et al.* 2002).

Buffalo Head Terrane The relatively high resistivity in the Buffalo Head terrane (labelled J in Fig. 5) is associated with the metaplutonic and subordinate felsic metavolcanic rocks, forming the terrane. The lower resistivity at mid to lower crustal depths in the east of the Buffalo Head terrane may be interpreted as the westwards extension of the Taltson Magmatic Arc beneath the Buffalo Head terrane. Additional MT soundings in the east of the Buffalo Head terrane and Taltson Magmatic Arc are needed to examine the extension of the Taltson magmatic zone in more detail.

Lithospheric Mantle Structure

The MT results show a region of enhanced conductivity at a maximum depth of 100-200 km depth across the mantle in the Hottah, Great Bear Magmatic Arc and Buffalo Head terranes. This depth is interpreted to be the base of the lithosphere. The source of the enhanced conductivity would be partial melt in the asthenosphere. High heat flow (109 mW/m^2) have been measured across the Hottah, Fort Simpson and Nahanni terranes by Lewis and Hyndman (2001) and along the Presqu'île Barrier to the west of the study area ($130\text{-}200 \text{ mW/m}^2$) by Majorowicz *et al.* (1989). Although there may be an influence from local topographic features and region fluid flow on these results, the values can be plausibly explained by a relatively shallow lithosphere-asthenosphere boundary.

The enhanced conductivity at shallow mantle depths beneath the boundary of the Hottah Terrane and Great Bear Magmatic Arc (labelled E in Fig. 5) and beneath the eastern Great Bear Magmatic Arc

(labelled H in Fig. 5) is more localized and is therefore interpreted to represent increased conductivity within the lithosphere rather than a decrease in the depth to the asthenosphere. Seismic reflection results show delamination structures extending to 100 km depth in the mantle beneath the Great Bear Magmatic arc providing evidence that lower crustal rocks were emplaced in the mantle during subduction (Fig. 6, Cook *et al.* 1999). This subduction was associated with collision and accretion of the Hottah terrane to the western margin of the Slave craton between 1.9 and 1.88 Ga. The source of the enhanced conductivity in the mantle beneath the western Great Bear Magmatic Arc may be either hydrogen or carbon introduced into the mantle through the subduction process.

The truncation of the mantle conductor beneath the eastern Great Bear Magmatic Arc at the GSLsz suggests significant movement of the mantle lithosphere as well as the crust. This motion could have removed higher conducting lithosphere from the southeast side of the fault. The ductile movement on the GSLsz (2.03 to 1.95 Ga) predates most of the orogenic activity associated with the collision of the Hottah terrane and Slave Province (1.9 to 1.88 Ga). Therefore, either the high mantle conductivity beneath the eastern Great Bear Magmatic Arc cannot be attributed to processes associated with the subduction of Hottah lithosphere, or else the truncation of the conductor occurred subsequent to the ductile motion on the GSLsz. It is possible that the conductor was truncated by the McDonald Fault, which runs parallel to the GSLsz, and formed during post-collisional convergence of the Slave and Rae provinces at ca. 1800 Ma (Ritts & Grotzinger 1994).

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Reprocessing of the LITHOPROBE data from the Southern Cordilleran Transect: Shallow reflections on Lines 7-8-9

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Reprocessing of Lithoprobe Southern Cordillera lines 7-9 has been undertaken to enhance shallow features that may be correlated to surface features delineated from detailed mapping. New mapping results in this region are interpreted with surface structures and strata having little post-Paleozoic relief (e.g., Thompson et al., 2001). In contrast, large (>10 km or so) structures that were previously observed throughout much of the crust in this region have been interpreted as Mesozoic-Tertiary contractional and extensional structures (e.g., Cook et al., 1992). Reprocessing of the original field data was accomplished at the LITHOPROBE Seismic Processing Facility. In contrast to the earlier processing done in 1988, the current work has incorporated refraction statics and detailed velocity analysis with cross-dip effects included. Post-stack processing includes FX filtering, coherency filtering and phase-shift migration. For migration, 90% of the laterally varying stacking velocity was used. The results of reprocessing have so far produced the following:

- 1) Shallow east-dipping reflections on line 9 that can be traced upward to ~300 ms (about 600-700 m depth) on the east flank of the Vernon antiform near Lumby. These reflections appear to become listric near about 1.0-2.0 s;
- 2) Subhorizontal reflections juxtaposed with the east dipping features in (1) in the central segment of line 9 at about 1.0-2.0 s;
- 3) West-dipping reflections that can be traced from about 300 ms on the west flank of the Monashee complex near Cherryville (lines 7 and 8) and flatten near 1.0-1.5 s west of there;
- 4) Very strongly reflecting shallow features at the central and west end of line 8 and at the west end of line 7.

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