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and

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

Pacific Geoscience Centre, Sidney, British Columbia
February 22-25, 2001

FRIDAY, February 23, 2001

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Preface

This volume includes the abstracts and short articles submitted for the 2001 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 both the southern and northern Cordillera. 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*

Continental arc-backarc evolution: the 2650-2710 Ma volcano-sedimentary sequences in the central Slave Province

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Introduction

The Slave Province is a complex Archean craton composed of N-trending basement gneisses, plutonic suites, mafic to felsic greenstone belts, and sedimentary successions (Fyson, 2000) that record 1.5 b.y. of Earth's earliest history, from the 4.03 Ga Acasta gneisses (Bowring and Williams, 1998) to ca. 2.52-2.6 Ga Late Archean stitching plutons (Davis and Bleeker, 1999). The Slave Province contains a relative abundance of exposed older basement that has left a geochemical imprint on younger 2722-2640 Ma greenstone belts (Cousens, 2000; Corcoran and Dostal, 2001). In addition, numerous unconformities between (1) basement-greenstone belts, (2) basement-molasse sequences, (3) basement-flysch successions, (4) flysch and molasse sequences, (5) sedimentary platformal and molasse sequences, and (6) basement - sedimentary platformal sequences (Henderson, 1985, 1998; Mueller et al., 1998; Mueller and Corcoran, 2001) indicate autochthonous basement - supracrustal belt - sedimentary cover relationships. Although there was significant deformation during the evolution of the Slave craton, the unconformities indicate that individual deformational events must have moved the sequences as large coherent blocks. Initial thought-provocative Slave subdivisions by Kusky (1989) with older (>2.8 Ga) Anton and Sleepy Dragon basement terranes and a younger (<2705 Ma) Contwoyto terrane with the Hackett River arc, have been modified to (i) N-trending west and east volcanic domains (Fyson, 2000) and (ii) prominent basement-controlled domains (Isachsen and Bowring, 1997). An exposed basement high in the central part of the Slave Province, represented by the 3.22 Ga Augustus plutonic complex (Northrup et al., 1999), the 2.95 Ga tonalitic gneiss at Beniah Lake (Isachsen and Bowring, 1997), the 2.93 Ga tonalitic gneiss at Patterson Lake (Bleeker et al., 1999a), and the 2.8 Ga gneiss at Sleepy Dragon Lake of the Sleepy Dragon Complex (Henderson et al., 1987), has been referred to as the central Slave basement complex (Bleeker et al., 1999a).

Numerous greenstone belts straddling the central Slave basement high have been the focus of our studies and include Point Lake, Beniah Lake, Beaulieu and Cameron River volcanic belts. Emphasis was initially placed on the physical volcanology and sedimentology of the belts or segments thereof (Corcoran, 2000; Mueller and Corcoran, 2001), concomitant with an in-depth geochemical analysis of the volcanic rocks from specific regions (Corcoran and Dostal, 2001), porphyry stocks (Simard, 2000), and sedimentary rocks in the molasse (strike-slip) basins (Corcoran and Mueller, 2001). The final stage of the project concentrated on U-Pb zircon age determinations to elucidate the timing of greenstone belt formation and pluton emplacement. Preliminary U-Pb-zircon data bracket greenstone belt formation and associated sedimentation between 2650-2710 Ma (Mueller et al., 2000; this study).

Point Lake volcanic belt

The Point Lake volcanic belt is a complex 2-4 km-thick volcano-sedimentary assemblage in which subaqueous mafic volcanic and felsic volcanic/volcaniclastic units are interstratified with the epiclastic turbidite-dominated Contwoyto Formation (Corcoran and Dostal, 2001). The mafic tholeiitic Peltier Formation is composed of massive to pillowed flows, pillow breccia and pillow fragment breccia with local horizons of shale and synvolcanic dyke systems that collectively represent an ocean floor setting upon which seamounts formed (Corcoran, 2000). The 2.69 ± 4 Ga dykes cutting the 3.22 Ga Augustus granite (basement) are the presumed feeders to this seafloor sequence (Northrup et al., 1999). Interstratified felsic volcaniclastic units in the Peltier formation yielded an age of 2673 ± 4 Ma (Northrup et al., 1999), and are comparable in age to the 2683.3 ± 8.6 - 8.2 Ma volcaniclastic deposits of the Beuparlant Formation (preliminary data from Mueller et al., 2000). The latter is interstratified with the deep-water, felsic-intermediate volcaniclastic Samandre and flysch-type Contwoyto formations (Henderson, 1998), suggesting contemporaneity between these various depositional units. An age of 2650.0 ± 3.2 Ma for a feldspar porphyry stock constrains the formation of the Contwoyto sedimentary rocks. The porphyry, emplaced along a splay of the pan-Slave Beniah Lake fault, intrudes and therefore post-dates deposition of the turbiditic flysch deposits (Contwoyto Formation). The clastic Contwoyto Formation in the study area is part of the same depositional event as the felsic volcanic-volcaniclastic units of the Samandre and Beuparlant formations, and is thus time-equivalent. An AFM diagram, REE patterns, and multi-element diagrams of the feldspar porphyry, indicate a calc-alkaline signature with Nb depletions (Simard, 2000) consistent with subduction-related processes. The feldspar-phyric porphyry stock plots as a volcanic arc granite in the discrimination diagram of Pearce et al. (1984), also supporting a subduction-related origin.

The Peltier Formation is composed of E-MORB or N-MORB-type basalts formed on continental crust in a backarc basin, whereas the felsic-intermediate units display a calc-alkaline signature (Corcoran and Dostal, 2001). The Sm-Nd isotope geochemistry of these basalts suggests crustal contamination (Northrup et al., 1999; Corcoran and Dostal, 2001) and the basement-cover relationship in the field lends additional support. Numerous unconformities including: (1) an erosional unconformity between the Contwoyto Formation and the ca. 2.6 Ga clastic molasse-type, Keskarrah Formation, (2) an erosional unconformity between the basalts of the Peltier Formation and the Keskarrah Formation, (3) an unconformity between the Contwoyto Formation and Augustus granite, and (4) an unconformity between the Keskarrah Formation and Augustus granite, support the geochemical inference of crustal contamination.

Beniah Lake volcanic belt

The Beniah Lake volcanic belt is a poorly documented area that was initially investigated for early > 2.8 Ga Slave evolution, emphasizing quartz-arenites and associated mafic and felsic volcaniclastic deposits. Recent mapping by Corcoran (Ph.D thesis in progress) with U-Pb zircon age determinations has shown a complex structural relationship between an early platformal volcano-sedimentary sequence and the younger greenstone belt. Although the volcanic succession, composed of predominantly massive and pillowed flows with abundant N-trending gabbro dykes, was initially considered part of deep-water sequence, detailed facies

mapping of newly identified interstratified felsic volcanoclastic units has shown a shallow-water, wave-influenced setting of pyroclastic origin. The large, composite 0.50-2 m-thick volcanoclastic bedforms grades upsection into a pillowed flow, consistent with a marginal arc in which paroxysmal volcanism congested the coastline. A similar scenario was documented for the Raquette Lake Formation of the Cameron River volcanic belt (Mueller and Corcoran, 2001). Abundant, prismatic, pale pink to colourless zircons yielded a preliminary age of 2668 ± 13 Ma, similar to the ages of the Point Lake and Beaulieu River belts. A calc-alkaline, synvolcanic porphyry stock intruding mafic, possibly gabbroic dykes, and aligned along an early NNW-SSE trending lineament, yielded an age of $2708.8 \pm 5.5/-5.1$ (preliminary data from Mueller et al., 2000). Some of the interstratified volcanoclastic and chert units may be contemporaneous with this synvolcanic porphyritic granite. The age is comparable to the Yellowknife supracrustal sequence proper and indicates that pre-2700 Ma felsic volcanism may be widespread. The adjacent tonalitic gneiss exposed over large segments of the Beniah Lake area is under renewed geochronological investigation in the vicinity of the synvolcanic porphyry stocks.

Beaulieu River volcanic belt

The 2-3 km-thick Beaulieu River volcanic belt, studied in detail near Stanley Lake, is a felsic-mafic sequence in which mafic basaltic and intermediate andesitic flows of calc-alkaline affinity are associated with tholeiitic basalts. Abundant brecciated pillowed flows and stratified hyaloclastite have been observed, and Corcoran (2000) has interpreted the mafic sequence, characterized by vesicle-rich flows as a shallow-water segment of a seamount. Mapping in 1999 revealed an extensive, up to 250 m-thick felsic unit composed of subaqueous massive volcanoclastic debris rich in euhedral 1-2 mm-size quartz crystals, as well as massive to brecciated quartz-feldspar-phyric flows or small domal structures. Recognition of flow-banding in the brecciated flows supports this interpretation. The volcanoclastic beds, considered to be of pyroclastic origin based on abundant liberated broken and euhedral quartz crystals, have an age of $2676.2 \pm 7.3/-3.4$ Ma, whereas the quartz-feldspar-phyric flows were slightly older, with a 2685.5 ± 2.9 Ma age (preliminary data from Mueller et al., 2000). The Sleepy Dragon complex represents the basement and is exposed to the west of the belt, where a local platformal quartz arenite sequence is preserved between the volcanic and plutonic rocks.

Cameron River volcanic belt

The Cameron River volcanic belt is considered the western counterpart of the Beaulieu River volcanic belt (Lambert, 1988). This volcanic belt is divided into a 3-4.2 km-thick subaqueous volcanic sequence with (1) a lower 1-3 km-thick mafic pillow basalt sequence interstratified with felsic flows and minor volcanoclastic, shale and siltstone horizons, (2) a middle dacitic-rhyolitic, 400 m-thick dome-flow sequence and (3) an upper 400-1400 m-thick, andesitic pillow and pillow breccia sequence with minor rhyolitic breccia and volcanoclastic horizons (Lambert, 1988). This volcanic sequence grades into the turbiditic flysch-type Burwash Formation towards the east and changes laterally in the SSE into the subaqueous to subaerial Raquette Lake Formation (Henderson, 1985; Mueller and Corcoran, 2001). Extensive dykes swarms in the Sleepy Dragon basement, considered as feeder dykes to the subaqueous flows by Lambert (1988) and Lambert et al. (1992), yielded an U-Pb age of 2687 ± 1 Ma

(Bleeker et al., 1999b). To the south of the Raquette Lake Formation, a felsic volcanic unit at Turnback Lake has been dated at $2663 \pm 7/-5$ Ma (Henderson et al., 1987).

The bimodal, 200 m-thick, volcano-sedimentary sequence is interpreted to represent a remnant arc segment floored by continental crust. The formation overlies the gneissic Sleepy Dragon Complex unconformably. A complex stratigraphy traceable for 7 km along strike displays three volcanic and three sedimentary lithofacies that constitute the volcano-sedimentary succession. The volcanic units are (1) a mafic volcanic lithofacies composed of subaqueous pillow-pillow breccia, and subaerial massive to blocky flows, (2) a felsic volcanic lithofacies representing felsic flows that were deposited in a subaerial environment, and (3) a felsic volcanic sandstone lithofacies interpreted as shallow-water, wave- and storm-reworked pyroclastic debris derived from explosive eruptions for which Bleeker et al. (1999b) obtained an age of 2683 ± 1 Ma. The sedimentary units are represented by (1) a conglomerate-sandstone lithofacies consistent with unconfined debris flow, hyperconcentrated flood flow and talus scree deposits, as well as minor high-energy, stream flow conglomerates that formed coalescing, steep-sloped, coarse-clastic fan deltas, (2) a sandstone lithofacies, interpreted as hyperconcentrated flood flow deposits that accumulated at the subaerial-subaqueous interface, and (3) a mudstone lithofacies consistent with suspension sedimentation in a small restricted lagoon-type setting. The debris in the conglomerate and sandstone beds originates from either the extrabasinal Sleepy Dragon complex or the intrabasinal felsic and mafic flows.

Conclusions

Our results favour a depositional setting in which an extensive arc - backarc system developed unconformably on a > 2.8 Ga continental crust. The arc system formed on the leading edge of a continental margin and resembles modern analogues such as Japan or New Zealand. The geometry of the Slave Province with a juvenile arc-backarc system to the east and older supracrustal rocks and basement to the west suggests an overall west-dipping subduction zone during the formation of this continental arc system. The autochthonous basement-cover relationships, covering a timespan of 600 m.y., suggest several stages of deformation during Archean supercontinent formation and disruption. The basement acted locally as a topographic high and hence, represents a source of detritus, as documented by abundant plutonic debris in some belts. The ages of the various greenstone belts overlap and formation between 2690-2670 Ma appears to be prevalent, based on recent and known age determinations.

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Kimberlite and xenoliths from the Southern Slave - a terrain with the highest diamond potential in the Slave craton.

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Introduction

Recent petrological studies of the Slave craton (NWT, Canada) show the underlying lithosphere to be uniquely heterogeneous and to have distinct petrological domains (Grütter et al., 1999; Griffin et al., 1999). Three northeast trending lithospheric domains (Fig. 1) that comprise the Slave mantle have different compositions (Grütter et al., 1999); both the northern and central domains are stratified (Kopylova et al., 2000; Griffin et al., 1999). However, at present little is known about the Slave mantle underlying its SE part. This paper presents the first xenolith-derived data on the composition and the thermal regime of the SE Slave lithosphere and characterises its kimberlitic rocks. We studied hypabyssal rocks of the 5034 pipe within the Kennedy Lake (KL) kimberlite cluster and its mantle xenoliths, as well as xenoliths from the Hearn pipe.

The Kennedy Lake kimberlite cluster (Fig. 1) consists of 5 pipes: the 5034, Tesla, Tuzo, Hearn and Wallace pipes. All the kimberlite bodies are significantly diamondiferous, the grade of 1.71 ct/t is reported for the 5034 pipe, 1.80-2.33 ct/t - for the Hearn pipe, 2.20 -for the Tuzo pipe (Mountain Province Mining News Releases, 10 July, 22 July, 10 Aug. 1999). The 5034 kimberlite pipe, and, perhaps, all of the Kennedy Lake kimberlite bodies have been emplaced during Cambrian time (Heaman et al., 1997). Thus, this cluster of kimberlites represents the oldest known kimberlite occurrence on the Slave Craton. The 5034 kimberlite comprises two igneous phases. One phase is a fresher hypabyssal rock, and another is represented by a diatreme-facies altered pyroclastic rock rich in granitoid xenoliths. This study is based on samples from the hypabyssal phase of 5034, which, also, is the major host of xenoliths in the Kennedy Lake cluster (H. Grütter, pers. comm.).

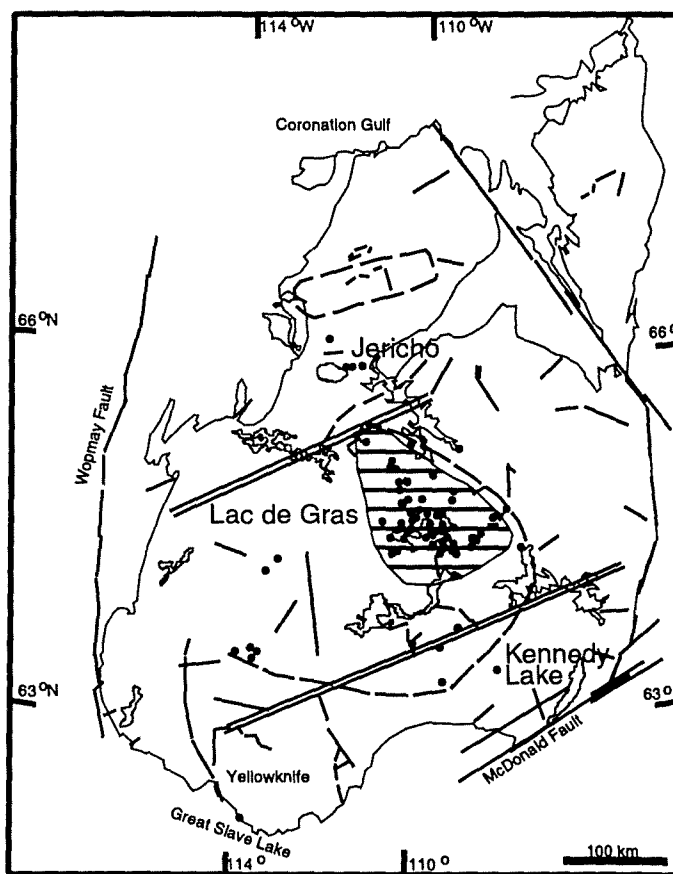


Fig. 1. Schematic map of the Slave craton, showing location of kimberlite pipes (black dots). Double lines designate boundaries between northern, central and southern lithospheric domains mapped by distinct compositions of garnet in kimberlite concentrates (Grütter et al., 1999). Also shown are the minimum extents of a shallow ultradepleted layer (horizontal lined pattern) and of the deeper Archean lithosphere (dashed outlines).

Kimberlite

The most common rock is a macrocrystal monticellite-bearing serpentine phlogopite kimberlite containing 30%-40% macrocrysts (Ol and Phl) and 70% groundmass. The groundmass typically contains phlogopite plates (20-55 vol.%), euhedral monticellite (5-20 vol.%) and olivine (10-20 vol.%), subhedral and euhedral spinel (<2 vol.%), subhedral perovskite (<2 vol.%) and anhedral to skeletal apatite (<2 vol.%), set in a late stage serpentine and chlorite (25-40 vol.%). Other accessory phases include anhedral calcite (0-10%), pectolite (0-10%) and barite. Diopside forms reaction coronae upon calcite-rich xenoliths and occurs as euhedral prismatic crystals in the groundmass. Samples containing diopside (up to 10%) do not bear significant amounts of monticellite. Crystallization of diopside in the 5034 kimberlite is thought to result from contamination of magma by xenoliths of calc-silicate country rock. Pectolite, a rare mineral in kimberlites, may have grown as a late-stage mineral in contaminated kimberlites (Scott-Smith, 1983) or may have crystallized from a residual alkaline Na-Ca-rich magma saturated in CO₂ and H₂O between 150 °C - 210 °C (Lo-Huann-Jih et al., 1998).

Another variety of hypabyssal kimberlite of the 5034 pipe is aphanitic; it is represented by thin dykes and by igneous rims of nucleated autoliths within macrocrystal hypabyssal samples. Aphanitic (<5% macrocrysts) kimberlite has a phlogopite-poor groundmass made up of euhedral olivine (25 vol.%), monticellite (10 vol.%), spinel (<1 vol.%), perovskite (<1 vol.%), apatite and a common late stage pectolite (up to 15 vol.%) or chlorite-serpentine (up to 50 vol.%). This kimberlite should be classified as a monticellite-bearing serpentine kimberlite.

Olivine present in the kimberlite is forsterite Fo₈₈₋₉₃; monticellite has 0.17-0.21 mol.% Fe. Phlogopite shows a discontinuous evolution to tetraferriphlogopite (TFP); homogeneous cores of phlogopite microphenocrysts are commonly mantled by late groundmass TFP enriched in Fe and depleted in Al- and Ti. Large spinel crystals commonly exhibit a trend of Ti- and Fe- enrichment, associated with Cr- Mg- and Al-depletion and, thus, roughly evolve from chromite cores to magnetite rims. Smaller spinel grains are typically homogeneous magnetite. The trend observed in these spinels is similar to Magmatic Trend 2 reported for lamproitic or orangeitic spinel (Mitchell, 1995). Perovskite is Sr- and REE-poor (5-6 wt.% (REE)₂O₃).

Macrocrystal hypabyssal kimberlite of 5034 pipe should be classified as a Type I kimberlite based on the presence of monticellite which is prohibited in Type II kimberlite or lamproites (Mitchell, 1995). However, a high abundance of phlogopite and distinct compositional trends of mica and spinel in the 5034 kimberlite is unusual for a Type I, non-micaceous kimberlite. A phlogopite-rich character of the 5034 kimberlite is similar to that of the Siniyskaya and Bezmyannaya (Siberia), Matsoku and Wesselton Type I kimberlite pipes (Zubarev, 1989) and several pre-Cretaceous Type I Southern African kimberlites (H. Grutter, Pers. Comm.). Chemical trends observed in mica and spinel of pipe 5034 are typically found in Type II kimberlites but occasionally reported for Type I kimberlite as well (Mitchell, 1995).

Mantle xenoliths

Petrography

Among our collection of 38 xenoliths, the most common rock is a coarse peridotite (74%), containing either garnet (69%) or spinel (31%). Other types of xenoliths are porphyroclastic peridotite (16%), mosaic peridotite (5%) and orthopyroxenite (5%). No eclogite is found. An absence of spinel-garnet peridotite indicates an unusual gap in sampling of certain mantle depths by a kimberlite host .

Coarse-equant peridotites are typically composed of olivine (50-90 vol.%), orthopyroxene (10-40 vol.%) and lesser amounts of clinopyroxene (0-10 vol.%), garnet (up to 10 vol.%) or spinel. The rock appears to result from an early crystallization of subhedral garnet and olivine followed by crystallization of subhedral to anhedral orthopyroxene, and late anhedral clinopyroxene and spinel. Garnets always crystallize prior to pyroxenes and are commonly poikilitically included in large anhedral olivines.

Most (90%) coarse peridotite is harzburgite, containing less than 5 vol.% clinopyroxene. Lherzolite is scarce and usually contains no more than 10 vol.% clinopyroxene. Other minor rock types are dunite, orthopyroxenite and wehrlite. Interestingly, spinel peridotites are commonly clinopyroxene-free, whereas garnet-bearing rocks can contain 0-10% clinopyroxene. This may indicate greater chemical depletion of shallow mantle in the Southern Slave similar to the Northern Slave (Kopylova et al., 2000).

Deformed peridotites have porphyroclastic or mosaic-porphyroclastic (cf. Harte, 1977) textures. Porphyroclastic rocks are garnet-bearing peridotites containing 10-80% neoblasts of olivine (50 to 70%) and orthopyroxenes (30% to 50%).

In mosaic-porphyroclastic peridotites olivine and orthopyroxene neoblasts typically comprise more than 95 vol.% of the rock, whereas garnet, olivine and pyroxene porphyroclasts are scarce. Neoblasts in mosaic peridotites are markedly smaller than those in porphyroclastic rocks. Mosaic peridotite typically exhibits a laminated texture, as neoblasts form thin lenses/bands associated with modal variations in olivine and orthopyroxene. Orthopyroxene bands are finer-grained than olivine stripes.

Mineral chemistry

Mineral phases in garnet peridotites were analyzed with a automated CAMECA SX-50 microprobe (UBC) with the same set of standards and analytical conditions that were used for studies of mantle xenoliths from the Jericho pipe (Kopylova et al., 1999).

There is no compositional difference between olivines from coarse and deformed peridotites (Fo₉₁₋₉₃ with 0.38 wt% NiO). Typically, olivine is homogeneous and porphyroclasts have compositions similar to neoblasts. In some deformed samples (2-7, 15-2) olivine rims are enriched in Fe (Mg#=0.87-0.90).

Orthopyroxene is homogeneous, with Mg# between 0.92 and 0.94 and a very low Al content, 0.21 to 0.44 wt.% Al₂O₃. It is poorer in Al than all mantle orthopyroxene reported for the Slave craton so far. Its low Al content is an indicator of its deep origin. Like olivine, orthopyroxene in coarse or deformed peridotite is undistinguishable in composition.

Clinopyroxene is classified as Cr-diopside with Cr-content ranging from 1 to 2.24 wt.% Cr₂O₃ and Mg# = 0.91-0.95. All chemical characteristics of clinopyroxene in coarse and deformed rocks are similar except for Ti; Ti content above 0.07 wt% was present only in coarse peridotite. Two thirds of peridotite samples have zoned clinopyroxene and compositional variations between grains. The characteristic compositional evolution involves a depletion in Al, Cr, Fe, Na ± Ti with a concomitant enrichment in Ca and Mg. Recrystallization into low-Cr, jadeite-poor clinopyroxene is controlled by lower T and P. Such late-stage clinopyroxene is typical for many pipes in the Slave craton and in Siberia (Kopylova et al., 1999a) and was ascribed to secondary crystallization during eruption (Boyd et al., 1997).

Garnets are fairly homogeneous in composition and can be classified as Cr-pyrope. On a Ca-Cr diagram garnet plots along the "Iherzolitic" trend reflecting the presence of clinopyroxene in the KL harzburgites (although in a low abundance). Equilibration with clinopyroxene is typical for garnet found in harzburgitic mantle of the Slave craton, as shown by studies of the Jericho (Kopylova et al., 1999a) and the Lac-de-Gras mantle (Pearson et al., 1999; MacKenzie and Canil, 1999). There is a strong positive correlation ($K=0.942$) between Ti and Na in garnet. Ti and Na are thought to be introduced to the peridotitic mantle by late metasomatic fluids on the basis that Ti does not correlate with any other element in garnet, but correlates with Ti and Na in coexisting pyroxenes.

In conclusion, the highly depleted mineral chemistry of all Kennedy Lake peridotite, its Mg-rich compositions of olivine and orthopyroxene are typical of low-T suites of cratonic peridotite which are interpreted as samples of lithosphere (Boyd, 1987; Harte & Hawkesworth, 1989; Boyd, 1998).

Thermobarometry

Equilibrium P-T conditions were computed for representative mineral compositions averaged for 3-6 analyses typically measured in mineral cores. Analyses of exotic Fe-rich olivine rims, secondary Cr, Al, Na-poor clinopyroxene were excluded from the dataset.

Equilibrium P-T estimates for analysed xenoliths were calculated with 2-Px and Ol-Gar geothermometers and Opx-Gar geobarometers usually recommended for xenoliths-bearing kimberlites (Pearson et al., 1994; Finnerty and Boyd, 1987). Equilibrium temperatures for harzburgite lacking clinopyroxene can be estimated by Ol-Gar geothermometers. However, a comparison of Ol-Gar T estimates (O'Neill & Wood, 1979) with 2-Px T estimates (Brey & Kohler, 1990; Finnerty & Boyd, 1987) at depth for 4-phase peridotite (Opx+Cpx+Ol+Gar) shows their inconsistency. The BK temperatures are up to 200° higher than the OW temperatures; the latter estimate also do not correlate with the FB geothermometer. We therefore could not use the OW Ol-Gar temperature estimates for clinopyroxene-free rocks in a single P-T array with 2-Px temperature estimates. The resulting P-T arrays were computed by two methods. Firstly, we used a combination of the FB geothermometer (Finnerty and Boyd, 1987) with the MG geobarometer (McGregor, 1979). Secondly, a combination of the BK geothermobarometers estimates (Brey & Kohler, 1990) was employed. Both sets of resulting P-T's (Fig. 2A and B) yield an essentially similar P-T array for the Kennedy Lake peridotite. In the FB-MG solution, data points are spread over a larger interval of pressure and temperature, whereas the BK solution is shifted towards higher P and T's and is characterized by a larger scatter of temperature estimates at any given pressure.

The Kennedy Lake peridotite formed at $T= 950-1250^{\circ}\text{C}$ and $P= 55-72 \text{ Kb}$ in a cold cratonic mantle deep within a diamond stability field. Regardless of the geothermobarometric method used, they plot at a slightly higher pressure and lower temperature than xenoliths from other Slave kimberlites, i.e. Jericho and Lac de Gras pipes (Fig. 2). This suggests a colder mantle below the SE Slave than below the NE and Central Slave at the time of its sampling by host magmas. At Kennedy Lake, porphyroclastic and mosaic-porphyroclastic peridotites formed at equilibrium P-T conditions similar to those of coarse peridotites. Here, a suite of low-T peridotite of lithospheric affinity may have been occasionally sheared, deformed. Our thermobarometric estimates indicate that the lithosphere of the SE Slave craton, sampled by the Kennedy Lake pipe is at least 230 km deep. This minimum lithospheric thickness is greater than that of the NE Slave (160-190 km, Kopylova et al., 1999b) and of Central Slave (200 km, Pearson et al., 1999). The thicker lithosphere and colder thermal regime makes the SE Slave a terrain with the highest diamond potential in the Slave craton.

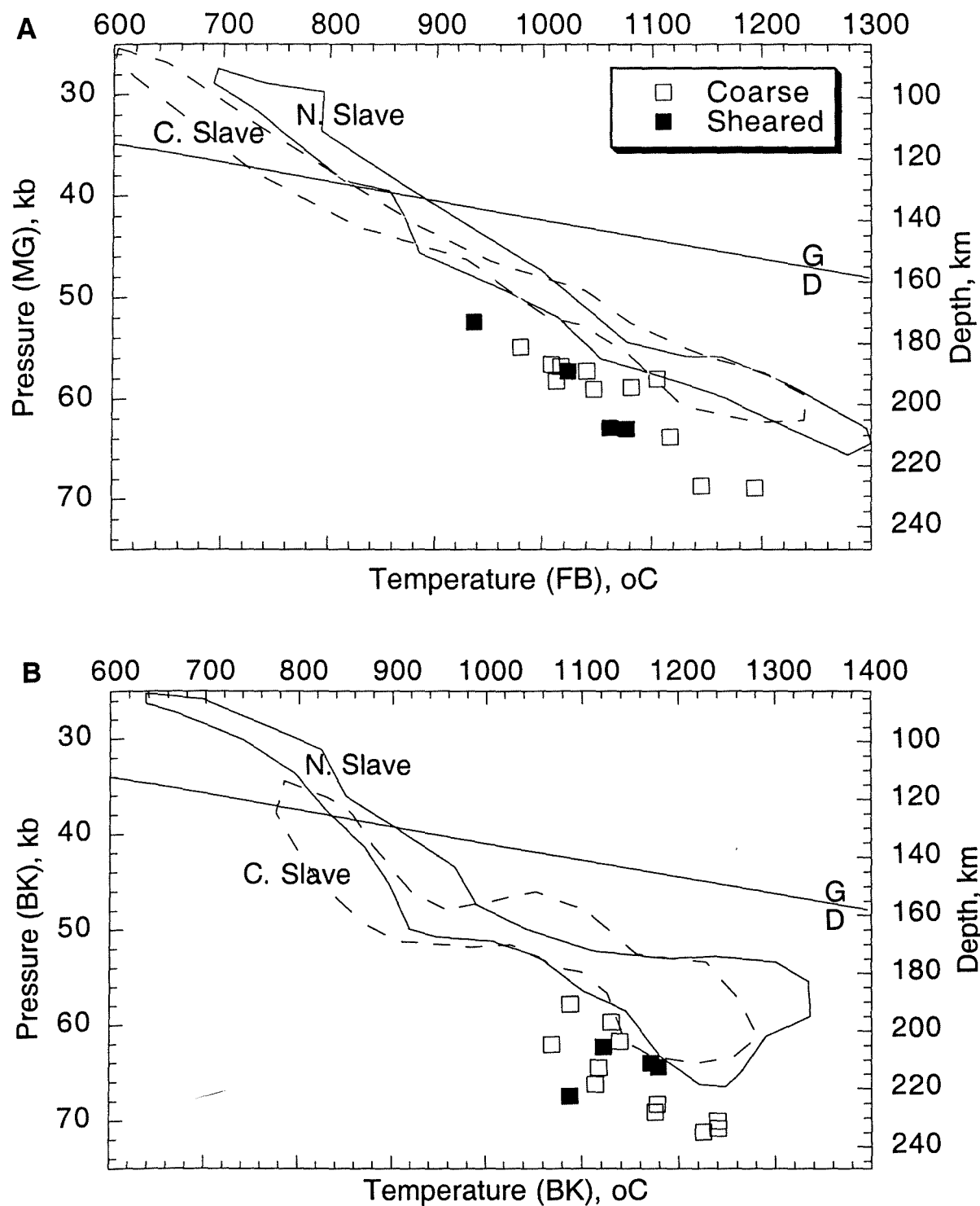


Fig. 2. Pressure-temperature estimates for Kennedy Lake mantle xenoliths in comparison with other Slave kimberlites according to (a) Finnerty and Boyd (1987) and MacGregor (1974), (b) Brey and Kohler (1990). A P-T field for the Central Slave (dashed outline) is based on Lac de Gras data (Pearson et al., 1999), a P-T field for the Northern Slave is based on the Jericho data (Kopylova et al., 1999a). Graphite-diamond equilibrium is from Kennedy and Kennedy (1976).

Kimberlite from the 5034 pipe sampled lithospheric mantle during the Cambrian, whereas kimberlites from Jericho and Lac De Gras pipes sampled NE and C. Slave domains during Jurassic and Tertiary time. The apparent differences in thermal regime and the lithospheric thickness between the SE portion and other domains of the Slave craton could primarily reflect its lateral heterogeneity. In this case, the SE lithospheric domain of the Slave has the coldest and deepest cratonic root. However, a possible temporal evolution of the whole Slave mantle cannot be excluded. The KL pipe gives us the earliest "look" into Slave cratonic mantle. Perhaps, the lithosphere below the entire craton was colder and thicker in the Cambrian, and a later evolution may have led to the lithosphere thinning and to hotter geothermal regimes in Jurassic-Tertiary time.

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Rheological Constitution of the Slave Mantle Lithosphere: Insights from Peridotite Xenoliths.

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The rheological character of the lithospheric mantle underlying the Slave craton is established using data derived from mantle-derived peridotite xenoliths. Specifically, we use grain size measurements and crystallographic preferred orientation (CPO) analysis in conjunction with previously published thermobarometric data to deduce the structural state of the lithospheric mantle over a depth range of 70 to 200 km. Mantle xenoliths (>200) for the study were derived from the diamond-bearing Jericho pipe and include coarse-grained peridotite, porphyroclastic (mylonitic and protomylonitic) peridotite and foliated and non-foliated eclogites. Similar to other cratons, the coarse grained peridotites are indicative of ambient lithospheric mantle which is equilibrated with a steady-state geotherm, whereas the mylonitic peridotites represent disturbed mantle lithosphere and falls off the steady-state geotherm.

Fabric analyses and microstructural observations indicate that dislocation creep was the primary strain-accommodating mechanism at all depths, including within the fine-grained mylonitic shear zones; diffusion creep, if active, was not the dominant mechanism of deformation.

Constraints on the strength of the Slave uppermost mantle were determined by calculating differential stresses using the olivine paleopeizometer of van der Wal et al (1993). Maximum and minimum differential stresses of 120 MPa and 3 MPa respectively are recorded by the Slave mantle peridotites. As anticipated, differential stress decreases with depth through most of the lithosphere.

Strain rates recorded in the Slave upper mantle peridotites were calculated using the thermobarometric data of Kopylova et al, (1998) in conjunction with the rheological parameters and the empirical flow law of Chopra and Paterson (1981,1984). Peridotites from the steady-state geotherm are adequately described by this empirical flow law, in that they describe a unique curve in terms of steady-state stress, strain rate and temperature. We conclude that the microstructures associated with the coarse grained peridotites are representative of the Slave (Archean) lithosphere.

In contrast, the high temperature mylonitic peridotites that fall off the steady-state geotherm have bimodal grain sizes consisting of coarse-grained fractions and fine-grained dynamically recrystallized grains. The fine-grained dynamically recrystallized olivine matrix is indicative of higher differential stresses (up to 120 MPa) and higher strain rates (10^{-5} to 10^{-8} s^{-1}). We interpret the high temperature porphyroclastic peridotite as samples of the lithosphere-asthenosphere boundary. They are a record of the transient, high stress, high strain rate deformation associated with localized, non-coaxial deformation at the lithosphere-asthenosphere boundary. Shear zone formation at the base of the lithosphere occurs as a result of the relative motion between the Slave craton and the underlying asthenosphere.

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Petrogenesis of the Coronation Sills and its relationship to the Neoproterozoic break-up of Rodinia

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The Franklin Igneous Event (FIE) represents the second largest mafic intrusive event in the Canadian Shield and spans nearly 2500 km across the Arctic Archipelago. The FIE consists of the Natkusiak flood basalts, Coronation Sills, Borden dykes on Baffin Island, dykes within the Brock Inlier, and possibly the Hottah Dyke Swarm. Regional U- Pb dating and paleomagnetic studies of dykes and sills on Victoria Island, Baffin Island and the Coronation Gulf suggest the FIE occurred 723 Ma ago just north of the equatorial paleolatitude. The Coronation Sills are typically 30- 100 m thick, exposed continuously for 100s of km along the shore of Coronation Gulf and hosted within carbonates, clastics and evaporates of the Neoproterozoic Shaler Supergroup. The sills are medium grained with subophitic texture and consist of calcic clinopyroxene ($\text{Wo}_{38}\text{En}_{45}\text{Fs}_{17}$), plagioclase (An_{60}) and minor amounts of olivine (Fo_{51}) and Fe- Ti oxides. Modal composition of olivine shows a systematic variation across the sills where phenocrysts steadily increase from the upper contact to almost the lower contact before decreasing rapidly at the base suggesting they have been affected by flow differentiation. Concentration of olivine is also reflected by a notable increase of Mg and Ni. Compositionally, the sills are tholeiitic basalts and are variably fractionated due to crystallization of plagioclase, clinopyroxene and olivine as inferred from Pearce Element Ratios. The REE pattern exhibited by the Coronation Sills is similar to **E-type MORB with a slight enrichment of the MREE**. The sills have a restricted La/Yb range (3.4 to 2.4) and match those of the Natkusiak Flood basalts. The distribution of other incompatible trace elements is also similar to that of E-type MORB and the Natkusiak flood basalts. Chemical compositions of Franklin dykes and sills from across the Arctic Archipelago show a distinct spatial variation with respect to the inferred source (plume) on Victoria Island. The intrusive bodies located further from the source region are chemically more evolved than those proximal to the source. Geochemical data imply that the FIE rocks developed from a source magma similar to enriched mid- ocean ridge basalt and represent a plate separation event. Using plate reconfigurations and other dyke swarm events emplaced during the Neoproterozoic, it is suggested the FIE was one magmatic event that contributed to the break-up of Rodinia. The timing and nature of dyking events in Australia, South China, Canadian Arctic, Newfoundland, and Norway suggest that Northern Rodinia began to break-up around 820 Ma and continued at least until 615 Ma in a manner similar to the opening of the modern Atlantic Ocean.

High heat flow and high crustal temperatures along the SNORCLE transect

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High heat flow has been measured along all of the east-west corridor of the SNORCLE transect west of the Slave province. Some of the new values calculated using bottom hole temperatures from wells penetrating shale sections are not well determined. However, the best determined results are the highest, with an average heat flow of 109 mW/m². Heat flow in the Cordillera of northern B.C. and the southern Yukon is very high, similar to the southern Canadian Cordillera. However, the heat flow is surprisingly high across the Hottah, Fort Simpson and Nahanni terrains, as Majorowicz (1996) first suggested. This might imply high crustal temperatures, reaching 800-1000° C, limited lithospheric strength, and a deformation style involving the whole, (thin) lithosphere. For the region east of the deformation front, we favor the alternate explanation of high crustal heat generation.

In the southern Canadian Cordillera, high heat flow is in contrast to lower values on the covered craton of the foreland to the east. In the north, high heat flow exists in the Cordillera and to the east on part of the craton which has been relatively stable for 1.1 Ga. Excess heat from the asthenosphere is expected beneath the northern Cordillera, which was a backarc region, similar to the southern Cordillera. But seismic velocities and other temperature indicators don't imply extremely high temperatures along the SNORCLE transect under the western craton. Therefore, a different, shallow source of heat is required in the craton.

High heat flow from Proterozoic cratons has also been observed in Australia where it is attributed to high heat generation in the upper crust (Sandiford and Hand, 1998). High heat generation from enriched upper crustal rocks eroded off the surface of the Slave province is a possible source of the extra heat within the

craton. The heat generation of basement grains from well cuttings, carefully selected for dating, is $3.2 \mu\text{W}/\text{m}^3$. However, the extraordinarily large heat generation may come from the small grains of accessory minerals that are preferentially washed out of the cuttings. The -120 mesh fraction of Slave samples with high heat generation is enriched by a factor of two in Thorium content, compared to the whole rock, so our measurements on cuttings may not represent whole rock data.

Seismicity of the Central and Southern Canadian Cordillera

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Introduction

This article summarizes the seismicity of the central and southern Canadian Cordillera - defined here as the region inland from the active plate boundaries extending along the British Columbia coast in the west to the interior platform in the east, and between the Canada-U.S. border in the south, and the British Columbia-Yukon border in the north (Figure 1). There is a marked contrast in both the rate of seismicity, and the maximum size of earthquakes observed in the central and southern Cordillera compared to the adjacent cordilleran regions both to the north and to the south (Figure 2). Here we present an overview of the seismicity, a description of earthquake monitoring and the earthquake history of the region; we discuss earthquake location and depth control, and examine the types of the earthquakes and their relationship to the tectonics of the region.

Earthquake Monitoring

The history of earthquake monitoring in this region is relatively short. The first station in western Canada began operating in 1898 at Victoria in southwest British Columbia. This was the only station in the Canadian Cordillera until 1951. There were, however, seismic stations in adjacent regions that could be used to locate Cordilleran seismicity. Seismic stations at Sitka, Alaska (opened in 1904) and Saskatoon, Saskatchewan (SAS) (opened in 1942) provided recordings of earthquakes of magnitude 5 and above within the central and southern Canadian Cordillera. Seismic monitoring improved in the Cordillera in the mid-1950's, when seismic stations were operated for a short period at Banff (BAN), Alberta and Lillooet (LLL), British Columbia. In the early 1960's, permanent Geological Survey of Canada (GSC) network stations began operating at Penticton (PNT), in the southern interior, and at Edmonton (EDM), just to the east of the central Cordillera (Figure 2). Completeness at the magnitude 4 level was not achieved for the Cordillera until 1971. In the early 1980's additional stations provided by BC Hydro began operating (e.g., MCC), and additional GSC stations were deployed along the northern coast between 1985-1990. Today the completeness threshold is approximately magnitude 2-2.5 for the Cordillera, with the exception of southwest BC where it is less than 1.

Seismicity Patterns

The majority of the seismicity in western Canada is associated with active plate tectonics along the western margin of the Cordillera (e.g., the Cascadia subduction zone and the Queen Charlotte transform fault). As is evident in Figure 2, the rate of seismicity drops off very quickly moving inland from the coast. The level of seismic activity increases on the eastern side of the Cordillera, in the region of the Rocky Mountain Trench and the Foreland Fold and Thrust Belt, and then decreases to very low levels across the Canadian Prairies (interior platform). There are

also significant north-south variations in the seismicity patterns. The seismicity in the central portion of the Canadian Cordillera, eastward of the Queen Charlotte transform fault, is significantly lower than the Cordilleran seismicity to the south (the region eastward of the Cascadia subduction zone), and lower than the seismicity of the northern Canadian Cordillera (eastward of the collisional tectonics in the Gulf of Alaska).

The largest earthquakes recorded in the central and southern Cordillera have occurred within the Foreland Fold and Thrust Belt. These are the magnitude 6 earthquake of 1918 (Rogers and Ellis, 1979) located near McNaughton Lake, and the magnitude 5.2 earthquake near Prince George (Rogers et al., 1990). The maximum size of earthquake ($M=6$) in the central and southern Canadian Cordillera is small (Figure 2) compared to the northern Canadian Cordillera, where thrust earthquakes as large as $M=6.9$ have been recorded (Wetmiller et al., 1988), and compared to the northern U.S. Cordillera where earthquakes of $M=7.3$ and larger have been recorded (for a summary and additional references see Doser, 1989).

There is no ongoing seismicity associated with the volcanic belts of British Columbia, including the prominent Mt. Edziza volcanic belt in northwestern British Columbia.

In addition to natural seismicity, just to the east of the Cordillera there are regions where earthquake activity appears to be associated with oil and gas extraction. Notably, the Fort St. John area of northeastern British Columbia (Horner et al., 1994) and near Rocky Mountain House, Alberta (Wetmiller, 1986).

The uncertainty in epicentral locations and focal depth for earthquakes within the Cordillera has varied considerably through the years, as the distribution of seismic stations has changed. It is important to note that even today, there is little depth control for these earthquakes. With the exception of those few earthquakes that occur within 10-15 km of a seismic station in the Cordillera, the focal depths can only be constrained, for example, to "the upper crust" (or about 0-15 km depth). Lower crustal earthquakes have been detected in the vicinity of the Cascadia subduction zone of southwest British Columbia, where very low heat flow (30-40 mW/m²) results in a thicker seismogenic zone, and earthquakes occur to about 30 km depth. The only subcrustal earthquakes detected within the Cordillera are those associated with subduction of the Juan de Fuca plate in southwestern British Columbia. Uncertainties in epicentral locations today range from about ± 10 km to about ± 50 km in both latitude and longitude, depending on distribution of seismic stations around the epicentre and the size of the earthquake. Such large uncertainties make it very difficult to associate seismic activity with any specific structure or fault. This will improve however, as additional seismic stations are deployed and location methods continue to improve.

Focal Mechanisms

To date, there are very few focal mechanisms determined for earthquakes in the central and southern Cordillera. This is due to the low magnitudes of the earthquakes (generally less than magnitude 4), the sparse station distribution, and the techniques that have been traditionally used (first-motion solutions). Currently however, new centroid moment tensor (CMT) programs are being installed and tested at the Pacific Geoscience Centre, and this will allow for the determination of focal mechanisms for many more earthquakes (as small as magnitude 3) using waveform modelling. Mechanisms for the largest events in this region (in the fold and thrust belt) show thrust faulting in a roughly east-west sense. This concurs with the major geological formations in the area, and is similar to the style of faulting in the fold and thrust belt of the northern Canadian Cordillera, where earthquakes as large as magnitude 6.9 have occurred (Figure

2). Initial mechanisms obtained using the CMT programs for smaller earthquakes ($M=3.5-4$) in the fold and thrust belt of the southeastern Cordillera show similar thrust mechanisms. Immediately to the south of the Canada-U.S. border, there appears to be a change from a compressional tectonic setting in the southern Canadian Cordillera, to an extensional regime in the northern U.S. Cordillera. Note that the largest events in the northern U.S. Cordillera are normal faulting earthquakes (Figure 2).

Summary

In summary, the seismicity of the central and southern Cordillera is low to moderate. The largest events have occurred in the fold and thrust belt of the eastern Cordillera. Focal mechanisms show thrust faulting and E-W oriented compression. This contrasts with the extensional regime observed just south of the Canada-U.S. border in the northern U.S. Cordillera, but is similar to the large events recorded in the northern Canadian Cordillera. Also noteworthy is the observation that seismicity within the Cordillera is lowest in the region inland from the Queen Charlotte transform fault, and highest in the regions that are inland from subduction or compressional regimes (the Cascadia subduction zone in the south, and the Yukatat block between northwest British Columbia and Alaska in the north). This suggests (see Adams and Clague, 1993) that Cordilleran seismicity is related to reactivation of existing faults caused by the transmission of crustal stresses from the active plate boundaries along the west coast.

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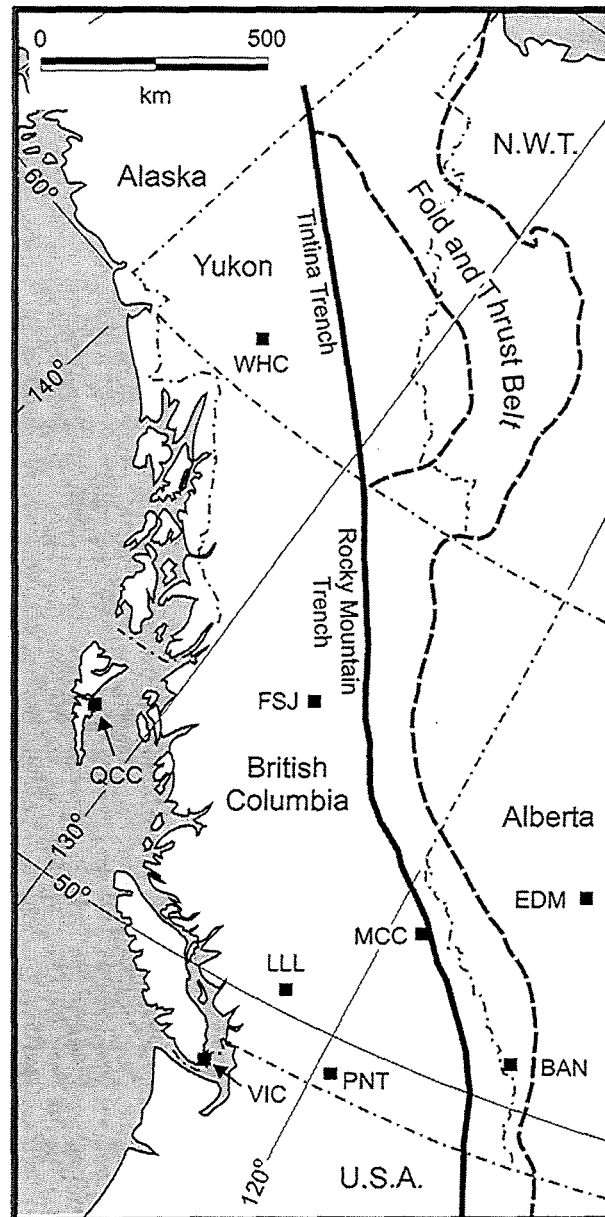


Figure 1. Major tectonic elements of the Central and Southern Canadian Cordillera and the location of some key seismograph stations (squares).

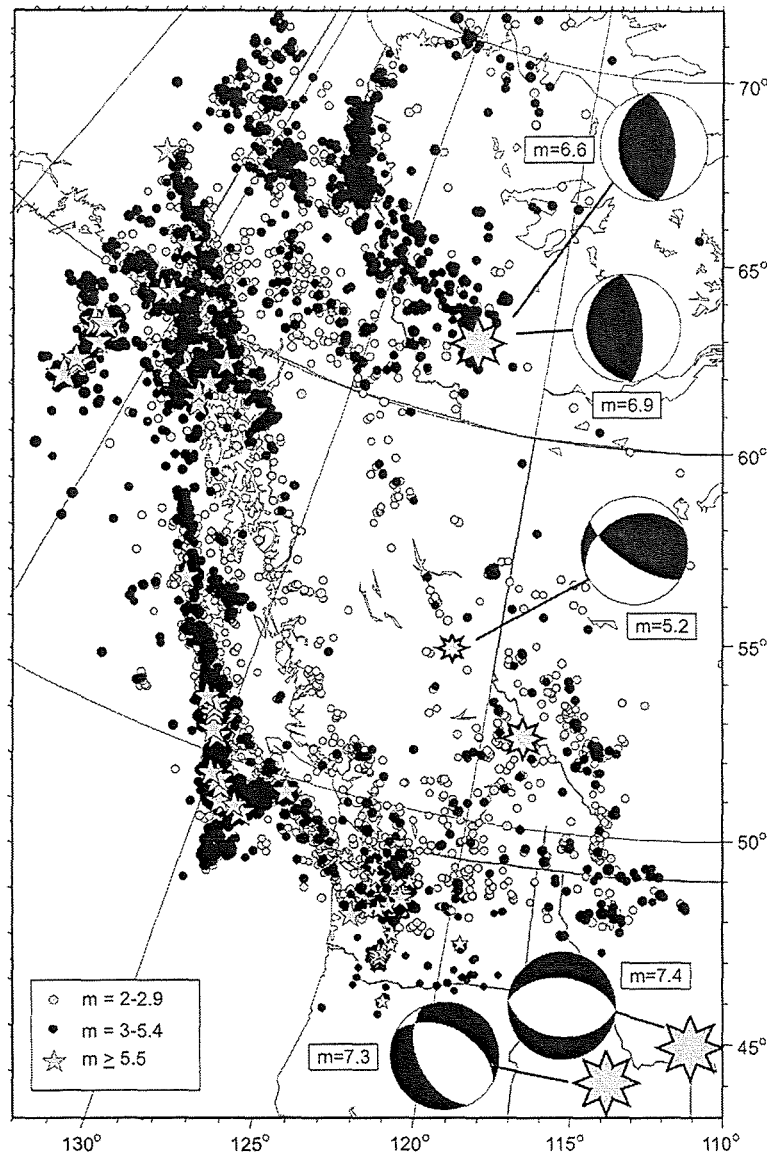


Figure 2. Background seismicity (1980-2000) of the Cordillera and adjacent regions. Seismicity is not complete for the regions south of 46° and west of 145°. The largest earthquakes in the central and southern Cordillera, the northern Canadian Cordillera, and the northern U.S. Cordillera are shown by eight-point star symbols. Focal mechanisms and magnitude for the largest Cordilleran events discussed in text, are shown.

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Flexural rigidity of the Lithosphere in the Canadian Cordillera

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The way the earth responds to applied geological loads provides important information on the long-term thermal, physical and mechanical properties of the lithosphere. These properties need to be specified in geodynamic models of loading, such as mountain building, formation of sedimentary basins and continental rifts. Lithospheric strength is often expressed as flexural rigidity or its equivalent, the effective elastic thickness, which is mainly a function of the temperature gradient, composition, and state of stress of the lithosphere. For the continental lithosphere the pattern of effective elastic thickness is complex. However, it provides important constraints on thermal and mechanical state of the lithosphere. In this work, the spatial distribution of effective elastic thickness of the lithosphere in the Canadian Cordillera is estimated from the stochastic relationship of topography and Bouguer gravity. A two-dimensional maximum entropy based spectral estimator is used to calculate the topography-gravity coherence function. This estimator allows for a higher resolution of elastic thickness than classical Fourier spectra and makes it possible to analyse tectonic provinces of dimension only a few hundred kilometers. The wavelength of transition from high coherence to low coherence of topography and gravity is directly related to the flexural rigidity. A significant variation of flexural rigidity can be observed for the Canadian Cordillera with much higher rigidity and effective elastic thickness values in the adjacent Precambrian shield. Generally the distribution of flexural rigidity shows good agreement with other thermally related geophysical data, like age of the lithosphere, regional heat flow, seismicity, seismic properties from active source experiments, and stress field.

LITHOPROBE Seismic Reflection Profiling of the Northern Canadian Cordillera: First Results of SNORCLE Profiles 2 and 3

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ABSTRACT

The Northern Cordillera in western Canada is underlain by a westward tapering wedge of layers that can be followed from outcrops of Proterozoic strata in the Foreland Belt to the lowermost crust of the Intermontane Belt. The layering was discovered on the new SNORCLE regional deep seismic reflection profiles along line 3 in the Yukon and lines 2b and 2a in northern British Columbia. East of the Tintina fault, the wedge layering dominates most of the crust with a relatively thin non-reflective zone above. As the upper seismically quiet zone thickens westward, the prominently layered wedge thins and tapers to zero at the reflection Moho about 40 km southwest of Dease Lake on line 2a, and near White Pass at the Yukon-Alaska border on line 3. These layers are correlated with thick (> 15 km) middle Proterozoic strata exposed in and underlying the Muskwa anticlinorium in northeastern British Columbia, and are thus interpreted as parautochthonous layers of Proterozoic strata that are correlative with strata of the Fort Simpson Basin visible on SNORCLE line 1. The detachments that carry deformed rocks of the Mackenzie Mountains and northern Rocky Mountains must therefore be subhorizontal either within or at the top of the layering.

The only significant geological break in the reflective zone coincides with the Tintina fault located near Ross River on line 3 and about 25 km west of Watson Lake on lines 2a and 2b. Other strike-slip faults, such as the Teslin, Kutcho, and Thibert faults do not penetrate the deeper layers.

A consequence of the interpretation of thick Proterozoic rocks beneath most of the northern Cordillera is that, with the exception of much of Stikinia, accreted rocks mapped on the surface are confined to the crust above the wedge. For example, in the Yukon east of the Teslin zone, klippen of Yukon-Tanana terrane are only a few km thick where they were emplaced upon pericratonic strata of the Cassiar Platform and Selwyn Basin, but may thicken to include much of the crust west of the Teslin zone. Along line 2a in northern British Columbia, accreted rocks of Stikinia are interpreted as a crustal-scale east-tapering wedge that overlies, at least in part, the deep layering. Bowser Basin sedimentary rocks and at least the upper part of Stikinia were deformed into the east-verging Cretaceous Skeena fold belt.

The reflection Moho exhibits variable character and undulates in a series of broad open folds with widths of about 150 km. On the west end of line 3, near White Pass, a mantle reflection dips eastward from about 15.0 s to about 20.0 s (about 47 to 69 km depth) and the Moho above shallows by 1.0-1.5 s (about 3.5-5 km). The mantle reflection is most easily interpreted as a frozen subduction surface and, although it is possible that the Moho rise is related to the subduction, a similar Moho geometry is observed on line 2a where the deep layering tapers and there is no subduction surface.

INTRODUCTION

Acquisition of ~1895 km of deep reflection data in northern British Columbia and the Yukon provides the first regional reflection profiles of the Northern Cordillera to establish a link between structures identified on SNORCLE Line 1 across the western portion of the Canadian Shield, deformed supracrustal rocks of the northern Cordillera, accreted rocks of the orogenic interior, and the active margin on the west coast. Objectives of the survey were thus to: 1) determine the western limit of pre-Cordilleran North American crust beneath the Cordillera, 2) determine the deep structure in the vicinity of major strike-slip faults such as the Tintina fault, and 3) determine the deep structure of accreted terranes and their relationship to North American rocks. Although many details are yet to be worked out, some of the key regional patterns are already evident.

ACQUISITION AND PROCESSING

Data were recorded along three profiles (Figure 1) between August, 1999 and February, 2000. Line 3 follows the south Canol road for about 650 km from MacMillan Pass in the Mackenzie Mountains, across the Tintina fault at Ross River, the Teslin zone at

Johnson's Crossing, and ends at White Pass on the Yukon-Alaska border. Line 2b follows the Alaska Highway for 570 km from Ft. Nelson, B. C., across the deformation front near Steamboat, the exposed Proterozoic strata of the Muskwa anticlinorium near Stone Mountain, the Tintina fault near Watson Lake and ends about 60 km west of Watson Lake. Line 2a follows the Campbell highway and Highway 37 for about 675 km from 40 km north of Watson Lake, across the Tintina fault near Watson Lake, the pericratonic strata of the Cassiar terrane, accreted rocks of the Stikine terrane and overlying Bowser Basin rocks in the Skeena fold belt, and ends in the Coast Mountains near Stewart, B. C. Lines 2b and 2a were recorded as separate lines to permit an overlap of 40-50 km in the vicinity of the Tintina fault west of Watson Lake.

Acquisition parameters (Table 1) were such that the final processed stacks include 32.0 s of data and nominal stack fold of 192. Processing (Table 2) was accomplished at the LITHOPROBE Seismic Processing Facility and required approximately one year to complete.

DESCRIPTION OF THE RESULTS

General Characteristics

The data quality is generally very good with reflections visible throughout most of the crust. In contrast to SNORCLE line 1 (Cook et al., 1999), reflections from the mantle are confined to a zone of dipping events on the western end of line 3; elsewhere, the mantle appears to be seismically transparent at this stage of the analysis. Hence, the transition from the reflective crust to the non-reflective mantle is easily identified as the reflection Moho.

Westward Tapering Wedge Layering

The dominant pattern within the crust along both profiles is a zone of prominent reflections that can be followed from the east, where it is visible throughout most of the crust, to the west where it thins and near the Moho. Near the Muskwa anticlinorium along line 2b, layers of the wedge rise to the surface correlate to exposed Proterozoic rocks (Figure 2) and have similar character to layers observed on the western end of SNORCLE Line 1 (Cook et al., 1999). They dip west from the Muskwa anticlinorium to the middle and lower crust (coinciding more or less with elevated seismic velocities; Welford et al, in press) and become diffuse between the Liard hot springs and Coal River, probably due to degradation of data quality from surface effects. Nevertheless, some reflections are visible

that allow correlation to improved data near Coal River. West of there, to Watson Lake, the layers comprise much of the crust and are more deformed, but can confidently be related to outcrops of Proterozoic strata. The layering is disrupted at the Tintina fault near Upper Liard, but west of the Tintina fault, similar layering tapers westward from a thickness of about 8.0-9.0 s (about 24-27 km) on the east to its disappearance about 40-50 km south of Dease Lake (Figure 3).

A similar zone of prominent reflection layering is visible along line 3 from MacMillan Pass to the Tintina fault at Ross River. Here, the layering is confined to the crust between about 4.0 s and 12.0 s, or about 12-36 km depth, and the upper crust is relatively non-reflective. As along line 2, the layering is disrupted at the Tintina fault, but can be followed west of there. The wedge layering tapers westward to the Teslin zone, where it has a thickness of about 5.0-6.0 s and then appears to thin west of there to only about 1.0 s thickness (about 3.0-3.5 km) near White Pass.

The crust above the tapering wedge is generally less reflective. On the east of line 2b, this zone is relatively thin, but thickens significantly west of the Tintina along line 2a. Along line 3, this zone is about 4.0 s thick near MacMillan pass, and thickens only slightly to about 5.0-6.0 s near the Teslin zone. West of there reflections in this zone are more numerous and complex.

The Crust-Mantle Transition and Mantle Reflections

The crust-mantle transition is visible as a downward loss of reflectivity along most of the length of the profiles. In general, it occurs between 11.0 and 12.0 s (about 33-36 km), although from the Western Canada Sedimentary Basin to the Tintina fault, it is diffuse and only clearly recognizable as a distinct reflection between Coal River and Watson Lake. West of the Tintina fault along line 2a, however, the reflection Moho is more prominent, and outlines broad ~150 km arches with crests located near the Tintina fault, and about 40 km north of Dease Lake. East of Stewart, B. C., the Moho appears to rise westward beneath the Coast Complex as predicted by Hammer et al. (2000).

The reflection Moho is visible along the entire length of line 3, and, although it is somewhat diffuse within about 50-60 km of MacMillan Pass and in the vicinity of the Tintina fault, it also exhibits arches with amplitudes up to 2.0 s (about 6-8 km) near Carcross and about 25 km east of Johnson's Crossing.

Sub-Moho reflections are rare. The only important exception occurs at the western end of line 3. Here, a suite of reflections is visible dipping from 15.0 s (about 47 km depth) on the west end of the line 3 to about 20.0 s (about 69 km depth) some 45 km east of there.

INTERPRETATIONS

The Westward Tapering Wedge

The geometric interpretation of crustal structure hinges on the nature of the westward tapering wedge. There are three essential ties to regional features that lead to an interpretation of these layers as Proterozoic strata: 1) line 2b crosses exposed Neo- and Mesoproterozoic strata of the Muskwa anticlinorium in northeastern B. C. providing a direct tie to outcrop, 2) line 3 crosses exposed Neoproterozoic rocks of the Selwyn basin providing a tie to outcrop for layers above the westward tapering wedge, and 3) line 2b crosses exposed Neoproterozoic strata east of Watson Lake.

Because reflections from within the westward tapering wedge are deep, they do not rise to the surface along line 3. Along line 2b, however, where the profile crosses exposed Mesoproterozoic strata, the layers are arched and project to outcrop in the Muskwa anticlinorium (Figure 2). Although some of the details of correlation are yet to be worked out, it is clear that the layering correlates to Proterozoic strata on the east and that, although individual stratigraphic horizons can not be traced from east to west, the geometric wedge of layers can. Accordingly, the wedge layers to the west may include layered strata, imbrications, and perhaps some basement upon which the strata were deposited.

Reflections from within and above the quiet zone can be correlated with known strata at the surface. For example, along line 3 near Macmillan Pass, a zone of west dipping reflections can be followed from about 0.5 s where they appear to be truncated by subhorizontal reflections, to about 4.0 s where they flatten into the prominent layering below. The surface rocks in the vicinity of these layers are gently folded subhorizontal Paleozoic strata that are only about 1.5-2 km thick (Gordey and Irwin, 1987).

Neoproterozoic rocks are exposed farther west in anticlinal culminations (e. g., near Sheldon Lake and Twin Creek), thus leading to an interpretation with the Paleozoic rocks overlying Neoproterozoic and older strata with angular unconformity. A stratigraphic link to outcrop is also available along line 2b between Coal River and Watson Lake where Neoproterozoic Windermere strata are exposed in the vicinity of the profile. As along line 3, the reflectivity from 0.0-3.5 s is less prominent than below about 4.0 s.

The interpretation of the westward tapering wedge as North American Proterozoic rocks implies that Precambrian North American crust and lithosphere are present in the subsurface as far west as the disappearance of the wedge – about 40 km southwest of Dease Lake along line 2a, and at least as far west as White Pass on line 3 - and thus underlie much of the northern Cordillera. As a result, Mesozoic detachments beneath the Mackenzie Mountains and northern British Columbia must flatten within or below the wedge such that the surface rocks, including accreted terranes, were carried above them.

The layers within the wedge are internally deformed, sometimes severely. Near Twin Creek on line 3, for example, subhorizontal layers between 9 and 12 s are truncated against east-dipping layers. A similar geometry is visible east of Watson Lake on line 2b and between 4.0 and 8.0 s along line 2a west of the Tintina fault. The magnitude of the relief on these structures (~10 km in some areas) is inconsistent with them being later than Paleozoic because the relief is not reflected in the younger rocks at these locations. Accordingly, there is evidence on these data for large-scale Proterozoic deformation preserved at depth beneath the Cordillera.

The distance from the Fort Simpson ramp to the western limit of the wedge provides an estimate of the width of the parautochthonous Proterozoic basinal strata underlying the orogen. Along a northeastward projection from line 3, for example, the Fort Simpson ramp has been detected east of the Mackenzie Mountains near Fort Norman (Mitchellmore and Cook, 1994), a distance of about 800 km from White Pass. Along a cross section commensurate with line 2, the Fort Simpson ramp has been detected near Fort Nelson, B. C. (Cook and van der Velden, 1993), a cross-strike distance of about 350-400 km from Dease Lake. Whether the difference between these estimates represents differences in the shape of the Proterozoic margin (e. g., with a large recess corresponding to the Mackenzie Mountains), or whether it represents different amounts of removal of the Proterozoic strata by younger deformation is uncertain at this time.

Strike-Slip Faults

Although the amount of displacement along the Tintina fault is uncertain in this region, it may be several hundred km, and if so, reconstruction could place the west side of the Tintina fault that is visible along line 3 against the east side of the Tintina fault that is visible along line 2. Nevertheless, the middle and lower crustal layering is observed on both lines 2 and 3 west of the Tintina, implying that the layering is extensive on both sides

of the fault along strike as well as across strike. Furthermore, the observation that the Tintina fault penetrates the deep layering requires that the layering is older than youngest displacement on the Tintina. Some other strike slip faults, including the Teslin (line 3), Kutcho (line 2a), and Thibert (line 2a), do not appear to penetrate the westward tapering wedge layering and are therefore interpreted to be transported above it.

Accreted Terranes

The identification of Proterozoic North American rocks in the subsurface constrains the position of most of the accreted terranes to the crust above the westward tapering wedge layers. Along line 3, for example, allochthons of the Yukon-Tanana terrane east of the Tintina fault and in klippen such as the St. Cyr klippe west of the Tintina must be less than a few kilometers thick. West of the Teslin zone, however, the thinning of the lower crustal wedge permits allochthons in the upper crust to thicken, although they are still confined to the crust.

Along line 2, the lower crustal wedge can be followed westward beneath the Cassiar, Cache Creek and northeastern 40-50 km of the Stikine terranes. All of these are therefore confined to the crust above the wedge. Southwest of Dease Lake, the Skeena fold belt includes Paleozoic to Cretaceous sedimentary and volcanic rocks of Stikinia and the Bowser Basin that were deformed into an east-verging thrust and fold belt during Cretaceous contraction (Evenchick, 1991). Identification of the Proterozoic tapering wedge may provide the geometric and tectonic link between the Cretaceous deformation of the Skeena fold belt and the broadly coeval deformation in the Foreland Belt of the northern Rocky Mountains speculated by Evenchick (1991).

CONCLUSIONS

Preliminary results from the 1999-2000 SNORCLE reflection survey along lines 2 and 3 provide evidence that Proterozoic North American strata are present at depth beneath much of the northern Cordillera. Along both lines 2 and 3, the North American layering has the form of a westward tapering wedge that thins and disappears near the Moho. At the west end of line 3, the wedge is very thin, whereas along line 2 it terminates about 40-50 km southwest of Dease Lake. The Tintina and possibly the Kechika faults penetrate these wedge layers, whereas most others appear to be confined to the crust above the wedge layering and thus the wedge layering must be older than that youngest movement on the Tintina. Similarly, accreted rocks, with the exception of much of Stikinia, are also confined to the crust above the layering.

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FIGURE CAPTIONS

Figure 1. Map of the SNORCLE transect region with locations of the three profiles.

Figure 2. Detail of Profile 2b in the vicinity of the Muskwa anticlinorium to illustrate the correlation of the layered reflections with Proterozoic strata. Cross section is scaled from Thompson (1981). Shaded area represents Phanerozoic strata on both the geologic section and the seismic section. Note that the seismic section turns north along strike and thus does not match the faulting characteristics on the west side of the geological section.

Figure 3. Segment of profile 2a near Dease Lake illustrating the tapering of the wedge layering near the Moho (arrow).

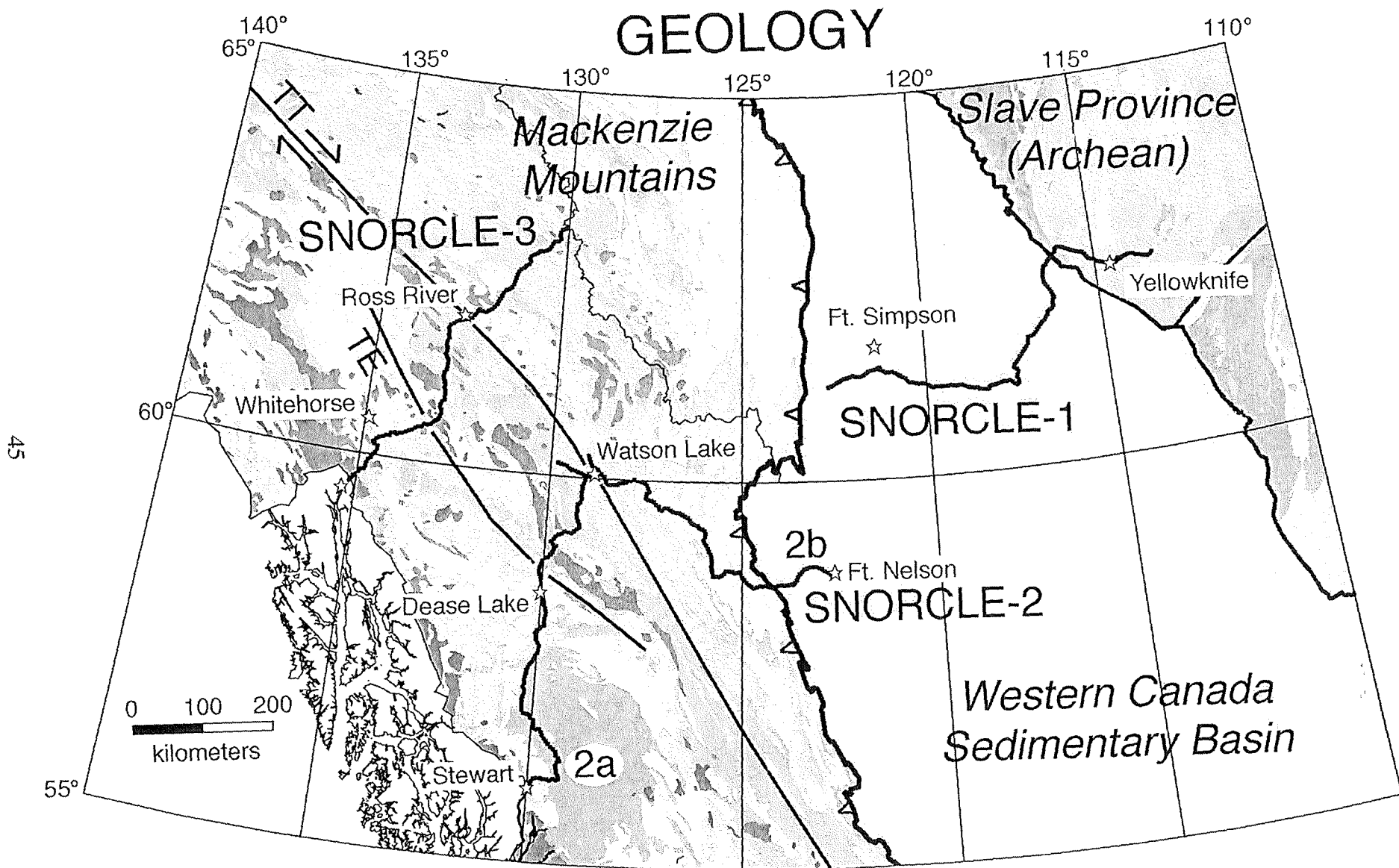


Figure 1

Line 2b - Muskwa Anticlinorium

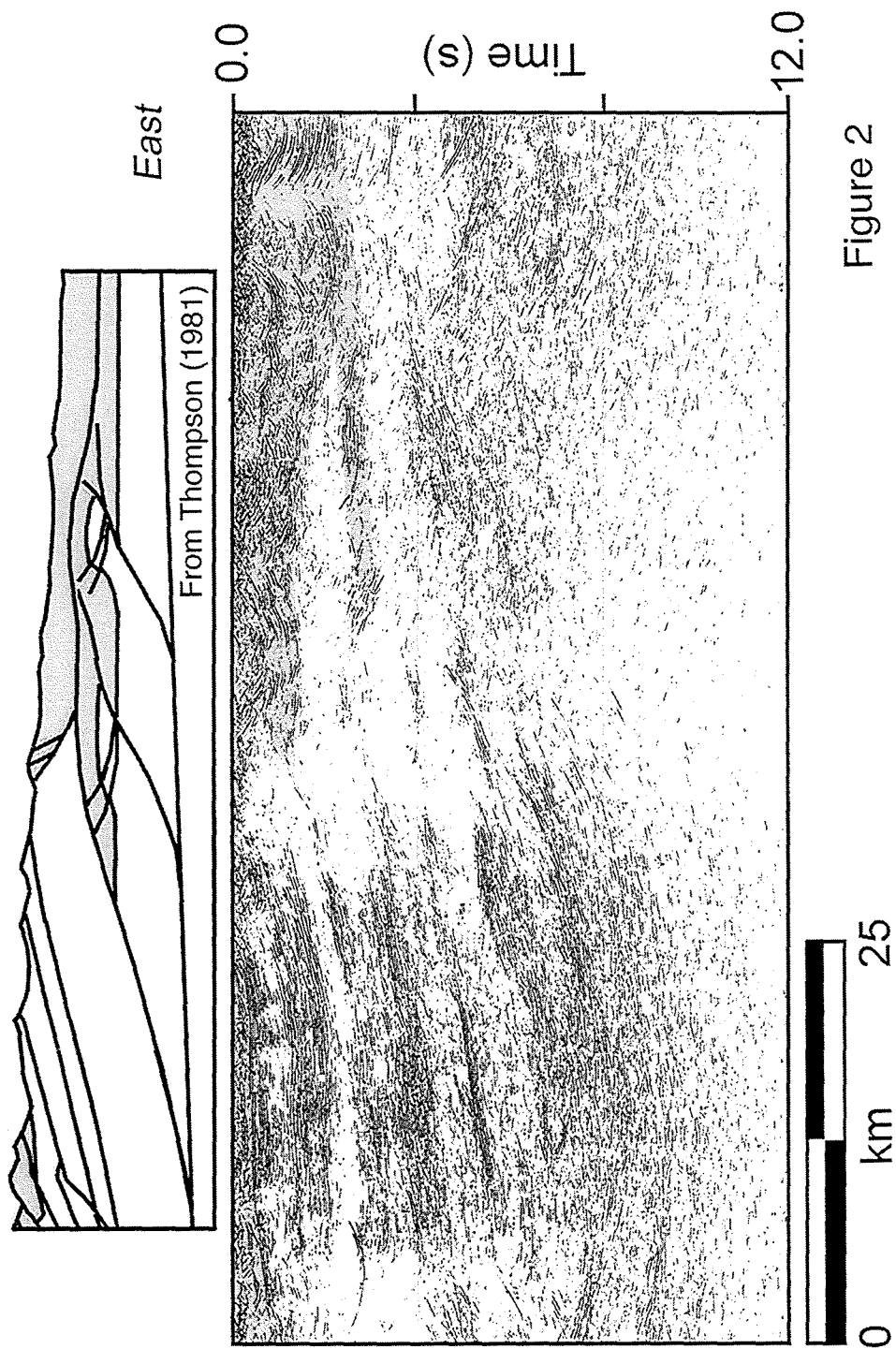


Figure 2

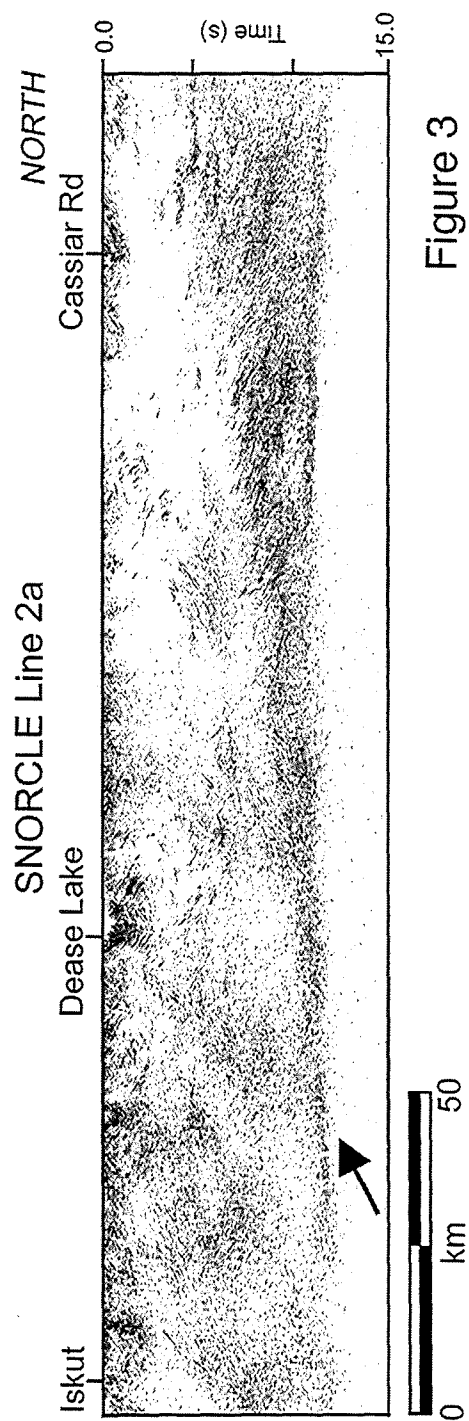


Table 1. Data Acquisition Parameters

Parameter	Value
Number of Vibrators	4 or 5
Number of sweeps	5 or 4
Sweep frequencies	10-80 Hz (linear)
Sample rate	4 ms
Sweep length	20 s
Record length	32 s (uncorrelated)
Number of channels/record	576
Receiver station spacing	50 m
Number of geophones	9 per station
Vibrator point spacing	75 m
Geophone frequency	10 Hz
Nominal fold	192
Instrument	I/O 2000, 24 bit

Table 2. Data Processing Parameters

Parameter	Value
Diversity-type stack of uncorrelated records	
Extended crosscorrelation	32 s record length
Crooked line geometry	25 m x 2000 m bins
Notch filter, where necessary	
Trace edits, first break picks	
Refraction statics computation	two layer
Velocity analysis	local constant velocity stacks
Application of first break mutes	
Automatic gain control	800 ms window
Gapped deconvolution	32 ms gap, 100 ms operator
Statics application	
Normal moveout correction on Common midpoint gathers	
Residual statics computation	4.0-8.0 s window
Cross-dip correction	
Trim statics	4.0-12.0 s window, +/- 12 ms
Stack	Nominally 192 fold
Trace energy balance	8-16 s window
Combine stacks	Redatum to 750 m (2a and 2b), 1200 m (3)
Migration	phase shift
Coherency filter	
Plot	

Seismic characterization of two major tectonic zones in the Canadian Cordillera: the Teslin and Tintina zones along SNORCLE Line 3

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In order to complement the standardized Lithoprobe processing of the entire SNORCLE seismic data set, we have undertaken more specialized processing that concentrates on two major tectonic zones crossed by SNORCLE Line 3. The elements of this new processing that may produce differing results are: (1) the choice of a new CDP line that more closely matches the Canol Road, (2) the use of a limited range of source-receiver offsets, 1500-7500 m in this case, (3) a revised velocity analysis partly based on new near-surface tomography models, (4) an additional post-NMO top mute, and (5) large-scale displays both migrated and depth converted using the new velocity models. The standard processing seeks to provide the best image of the entire lithosphere crossed by the profile. Our specialized processing is meant to optimize reflections within the uppermost few kilometers of the crust. In theory, this high-resolution processing should image all the major features initially revealed by the standard processing but also some additional features that are either weaker or less laterally coherent.

The tomographic velocity models are derived from the first P-wave arrivals observed on the reflection profile receiver array. Each SNORCLE 3 vibrator source point was recorded out to 12 km in both directions along the Canol Road. This range of offsets typically provides information about velocities down to 3-5 km depending on the velocity structure. The dense ray coverage available provides very reliable, high-resolution velocity models that do not depend strongly on the skills (i.e. prejudices) of the modeller. The accompanying figures show the resultant velocity models for the Teslin and Tintina tectonic zones where crossed by SNORCLE Line 3. Velocities range from 1.5 to 7.5 km/s. The tomography algorithm used is that included in the ProMax processing package.

Tintina Tectonic Zone. One section of SNORCLE 3 that was looked at in detail includes VPs 14500 to 15250, from just NE of the Pelly River ferry crossing to near the road summit within the St. Cyr Range. The crustal scale section shows a major disruption of lower crustal reflectors within this interval and suggests faults that cut the entire crust. Within the uppermost three kilometers of the 'high-resolution' section a few synformal reflectors are observed locally, both within the Tintina 'graben' and in the Lapie River valley to the SW. More generally within the uppermost ten kilometers reflectors are very weak and difficult to resolve. The few stonger, more coherent ones of this group dip to the SW at moderate (20-50 degrees) dips (Figure 1). One set of reflections dips NE at

2.5 km depth beneath the Lapie River bridge; these reflections project to the surface at about VP 15050. At 10-20 km depths the set of NE dipping reflections observed by the standard processing also appear on the 'high-resolution' section.

Strong variations observed in the tomographic velocity model do not correlate well with individual reflections. Velocities less than 2.0 km/s are common within the uppermost 500 m between Ross River and the Lapie River bridge, that is, within the Tintina 'graben' (Figure 1). Directly below this layer the velocities range from 4.7 to 6.8 km/s over lateral distances of only a few kilometers. Along geological strike from these anomalies lie elongate outcrops of Slide Mountain rocks of amphibolite to granulite metamorphic grade and outcrops of Tertiary volcanic rocks (Gordey & Makepeace, 2000). These rock units would be expected to have relatively high (>6.3 km/s) and low (<5.5 km/s) P-wave velocities, respectively. These velocity variations are resolved with reasonable confidence down to 2.0 km depth. The straightforward interpretation is that buried extensions of the mapped tectonic slivers of Slide Mountain rocks occur directly beneath the town of Ross River and 8.5 km to the SW. In between and to the SW lie either slivers of Tertiary volcanics or more recent alluvial deposits. Beneath the Lapie River bridge velocities are 6.2 km/s at 1 km depth and remain at >5.8 km/s further to the SW and at greater depths. These are presumably representative of the Cassiar terrane rocks making up the St. Cyr Range.

In general, magnetotelluric conductor anomalies have an anti-correlation with these velocities (Ledo et al., 2001a). A shallow conductor was located at 2-3 km depth just SW of Ross River where one of the low-velocity slivers occurs. Another shallow conductor lies SW of the Lapie River bridge and dips SW to depths of 10-15 km; this may correlate with some of the SW-dipping reflections at 6-10 km depth noted above.

Teslin tectonic zone. The part of SNORCLE 3 processed as a high-resolution study of the Teslin zone includes VPS 18250 to 19450, from where the Canol Road turns to the SSW near Murphy Creek to near Little Teslin Lake. The crustal-scale section shows prominent lower crustal reflectivity is continuous throughout this section and only a few, isolated E-dipping reflections occur in the upper third of the crust. Within the uppermost five kilometers of the 'high-resolution' section a few clear E-dipping reflectors are observed; these correlate with those observed in the crustal-scale section. One reflection can be traced to within a few hundred meters of the surface a few kilometers west of Johnson's Crossing near where SNORCLE 3 joins the Alaskan Highway. Another reflection traces a double antiform structure overlying the above reflection, before it too dips to the east at moderate (20-30 degree) dips. The majority of the weaker, less coherent reflections in this shallow section also dip moderately to the east.

Only one strong lateral variation in P-wave velocity is observed on the tomographic velocity model along this part of SNORCLE 3 (Figure 2). Relatively high velocities (>6.5 km/s) occur beneath and west of Johnson's Crossing at 1-3

km depth (not resolved beneath 3 km). Relatively low velocities (<4.5 km/s) overlie the high ones, but also occur for 5 km further to the east, up to the surface trace of the Teslin fault (Gordey & Makepeace, 2000). East of the Teslin fault the uppermost several kilometers of the crust has moderate velocities of 5.0-6.2 km/s. Lithologies are not tightly constrained here, but the high velocities west of Johnson's Crossing may correlate with mafic volcanic or carbonate rocks of the Cache Creek terrane. The moderate velocities to the east appear to correlate with a Cretaceous(?) granitic pluton as do high conductivities (Ledo et al., 2001b). It is not possible to resolve the overall dip of this distinct velocity boundary with information from only the uppermost few kilometers, but it does occur beneath the surface trace of the Teslin fault where a sharp change in rock type and metamorphic grade has been mapped (Gordey & Makepeace, 2000). A suggestive correlation between this lateral velocity gradient and the E-dipping reflections noted above indicates a moderate eastward dip to the Teslin fault within the upper crust.

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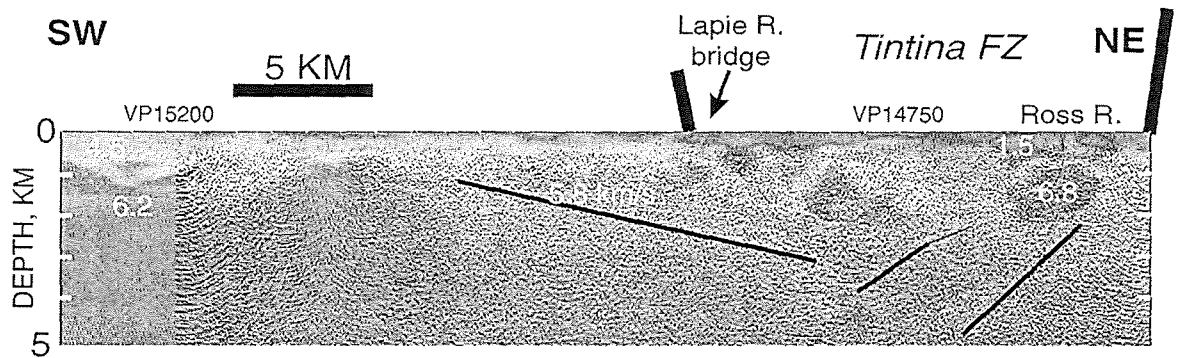


Figure 1. 'High-resolution' reflection section of the Tintina tectonic zone portion of SNORCLE 3. Greytones indicate P-wave refraction velocities as determined by tomographic inversion of first break picks. Black lines highlight a few of the more coherent reflections although these can only be observed on larger scale versions of the section. Both the reflections and velocity variations are consistent with a series of tectonic slivers within a 'flower' structure.

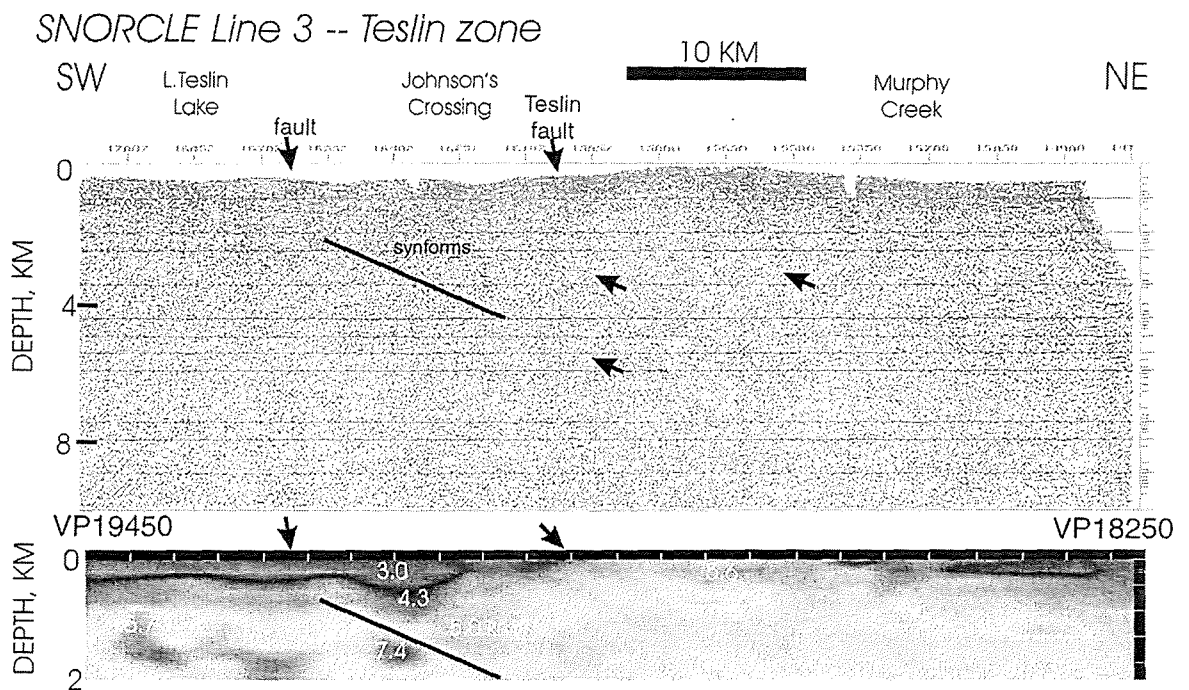


Figure 2. 'High-resolution' reflection section of the Teslin tectonic zone portion of SNORCLE 3. Greytones in lower section indicate P-wave refraction velocities as determined by tomographic inversion of first break picks. Black arrows highlight a few of the more coherent reflections. Note that the reflection section has elevation statics applied, the velocity model does not; this produces a discrepancy in depth labels.

Some geological constraints on Cordilleran structure imposed by Line 3 reflections – MacMillan Pass to White Pass

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SNORCLE Line 3 was oriented almost ideally across the northern Canadian Cordillera and provided a 650 km transect of the orogen from the foreland thrust-fold belt to the plutonic-metamorphic core in the Coast Belt. It followed the Canol Highway from the Northwest Territories-Yukon boundary at MacMillan Pass to Johnson's Crossing, along the Alaska Highway from Johnson's Crossing to Jake's Corner, along Yukon Highway no. 8 from Jake's Corner via Carcross to the Yukon-British Columbia boundary, and from there to the Canada - USA boundary at White Pass.

A reflection Moho at fairly constant depths of about 11 - 13 s (two-way travel time, ~ 33 - 39 km depth) delineates the base of crustal reflectivity. The mantle beneath it appears seismically transparent except for an east-dipping reflection at the western extremity of Line 3; this contrasts with the generally more reflective mantle beneath the Shield and pre-Cordilleran craton in Line 1.

One of the most spectacular and likely most significant features uncovered by the reflection survey is the deep "westward tapering wedge" (WTW), a westward-thinning strongly reflective layered zone prominent in the middle and lower crust. The top of the WTW is ~ 11 - 12 km deep at MacMillan Pass

and the layering is interpreted as a combination of stratigraphy, Proterozoic deformation structures, and possibly thin basement of the ancient margin (see Cook et al., this volume).

The Tintina fault system, seen as a near-vertical zone of poor reflections through the entire crust, does not offset the wedge in a dip-slip sense. The layered WTW can be traced as far west as White Pass, where the top of wedge is visible at ~ 30 km depth. If the WTW on Line 3 is the same material as seen on Line 2 that can be correlated with exposed Proterozoic strata in the Muskwa Ranges, these layers can be followed to within ~ 150 km of the present continental edge. The Cordillera in the Yukon appears to be underlain to by craton-derived ancient North American strata and possibly basement, to within ~ 40 km of the Denali fault.

Any palinspastic restoration of the ancient margin thus requires that sutures between the North American plate and overlying rock assemblages interpreted as oceanic accreted terranes must root to the west of White Pass. Similarly, the basal detachment of the Cordilleran orogen must root to the west of that point, whether it is at the top of the westward tapering wedge or lower in the crust.

To the east of the Tintina fault system, the WTW is overlain by a thinner and seismically more transparent zone. Weak reflections characterize this quiet zone of the upper crust between MacMillan Pass and the Tintina fault. They likely image layering and thrust-fold belt structures in Paleozoic and underlying latest Proterozoic strata of the former Selwyn Basin. Antiformal fold structures with several kilometres of wavelength and amplitude are visible in layering that can be traced westward from rocks of the latest Proterozoic Hyland Group near MacMillan Pass. Those reflections are truncated by the uppermost reflections in the crust, which are correlated with overlying lower and middle Paleozoic strata that are interpreted to be < 2 km thick, recording older deformation and a likely extensive regional angular unconformity (see Cook et al., this volume).

Where it is crossed by the seismic profile, the “flap”, i.e., that part of the Yukon-Tanana terrane on the east side of the Tintina fault system, is underlain by a number of shallow reflections, suggesting that it is not more than 1 – 2 km thick.

To the west of the Tintina fault system, the upper crust is similar, i.e., is generally seismically transparent and shows no significant break in the crust. The accreted terranes that dominate the surface geology are either seismically similar to the underlying parautochthonous strata, or are too thin to be distinguished in the upper 1 – 3 km of the profile. Reflections that underlie the western part of Line 3 above the WTW, particularly between Carcross and White Pass, are complex and need further interpretation.

For example, it cannot be determined whether the upper and middle crust of the Cordillera is part of “Stikinia”, as suggested by

the evidence on Line 2 (see Evenchick et al., this volume). As seen on Line 2, Stikinia would form an easterly tapering wedge between the overlying Cache Creek terrane and the underlying WTW and (or) crystalline basement.

Approximately 20 km to the southeast of Line 3, the Cache Creek terrane is thrust over rocks of Stikinia along the Nahlin fault, but the nature and dip of the contact along the seismic profile are unclear. Similarly, rocks in the narrow belt of “Stikinia” that is crossed by Line 3 have uncertain relationships with metamorphic rocks of the Nisling terrane to the west. From local field evidence, the contact may be partly depositional. A number of steep structures parallel to the contact are mapped a few kilometres to the east within Stikinia, and these may reflect a steeper tectonic contact. One of these structures, the Llewellyn fault, is crossed by the seismic profile but is not apparent in the reflection data as a significant structure.

In addition, the interpretation of an eastward-tapering tectonic wedge of Stikinia along Line 3 would require that it (Stikinia) is tectonically juxtaposed with rocks of the Nisling “terrane” to the west. The Nisling terrane rocks, which are cratonic or pericratonic metamorphic rocks of likely Proterozoic age, occur at the surface at the western extremity of Line 3. Their possible North American margin affinity places an additional constraint on the nature of the underlying crust.

If the Nisling rocks are not of North American affinity, they nonetheless represent cratonic crust to the west, which must either overlie or underlie part of Stikinia along a moderately dipping contact, because no steep fault of crustal dimension is suggested by the reflection data.

Highlights of SNORCLE Line 2A - the Cassiar-Stewart Highway (Highway 37, 37A)

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SNORCLE Line 2A (Figs 1,2), acquired in late 1999 and early 2000, complements the other two LITHOPROBE cross-Cordillera transects (SNORCLE Line 3 and the Southern Cordillera Transect) by providing the most complete cross section of terranes (Cassiar, Slide Mountain, Quesnellia, Cache Creek, and Stikinia). Most of the major terranes along this transect are well exposed and the natures of their boundaries are relatively clear. Most significantly, the transect crosses the largest terrane, Stikinia, which is absent in the Southern Cordillera Transect, and less well defined in the northern transect (SNORCLE Line 3).

Seismic data obtained from the SNORCLE Line 2A transect along highways 37 and 37A, from Watson Lake Yukon Territory, to Stewart British Columbia, show the following crustal characteristics:

1. A Moho at depths of about 33-39 km (11-13 sec. two way travel time) beneath Ancestral North America and most of Stikinia.
2. A westerly tapering wedge of Proterozoic strata and/or crystalline basement that ends at the Moho below Stikine River, and is most clearly defined between Stikine River and the Kechika Fault.
3. At the Northern Rocky Mountain Trench the entire crust is disrupted by a near vertical zone of poor reflectivity. The Kechika Fault trace is above a similar, but less striking zone.
4. The tapered wedge of Proterozoic strata and/or crystalline basement appears not to be penetrated by the Kutcho and Thibert faults, the latter of which has been considered to be a

fundamental suture zone involving Stikinia, Cache Creek, Quesnellia, and Ancestral North America.

5. The Cache Creek and Slide Mountain (Sylvester Allochthon) terranes appear, as previously inferred, as thin flaps or nappes lying on Stikinia and Ancestral North America, respectively.
6. Stikinia is characterized by gently to moderately northerly and southerly dipping reflections. They are similar in nature (density, variation in dip, continuity) throughout the crust with the possible exception of the region between Stikine River and about 57°N, which has more diffuse reflections near the Moho.
7. Strata of the Bowser Basin are strongly reflective; large scale contractional structures of the Skeena Fold Belt are apparent in Bowser Basin strata *and* Stikinia.
8. The Moho rises to its shallowest depth, ca. 30 km, at the southwest end of the transect.

Most terranes are well defined by reflections, and apparent dips of reflections are consistent with field relationships near major boundaries. Significant strike slip faults (Fig. 2; Northern Rocky Mountain Trench, Kechika, Thibert, and Kutcho) coincide with lateral changes in reflection character, and/or zones of more diffuse, less continuous reflections. A better understanding of the ages and structural histories of these faults would aid interpretation of their significance. From the gross geometries described in points 2, 4, 5, and 6, it appears that Stikinia north of Stikine River forms an easterly tapering wedge between the overlying Cache Creek terrane and underlying westerly tapering wedge of Proterozoic strata and/or crystalline basement. Structures above the Stikinia wedge have been considered to be associated with the suture between Stikinia and terranes to the east. Their position, more than 90 km northeast of the southwest-most point of the westerly tapering wedge of Proterozoic strata and/or crystalline basement, suggests that they were transported northeastward relative to the Proterozoic and/or crystalline wedge. The top of the westerly tapering wedge may represent this suture and/or a contractional fault connecting structures associated with the Skeena Fold Belt with those of the Omineca and Foreland belts. In these scenarios the wedge of Stikinia is a tectonic sliver between Proterozoic strata and/or crystalline basement below, and accreted terranes above.

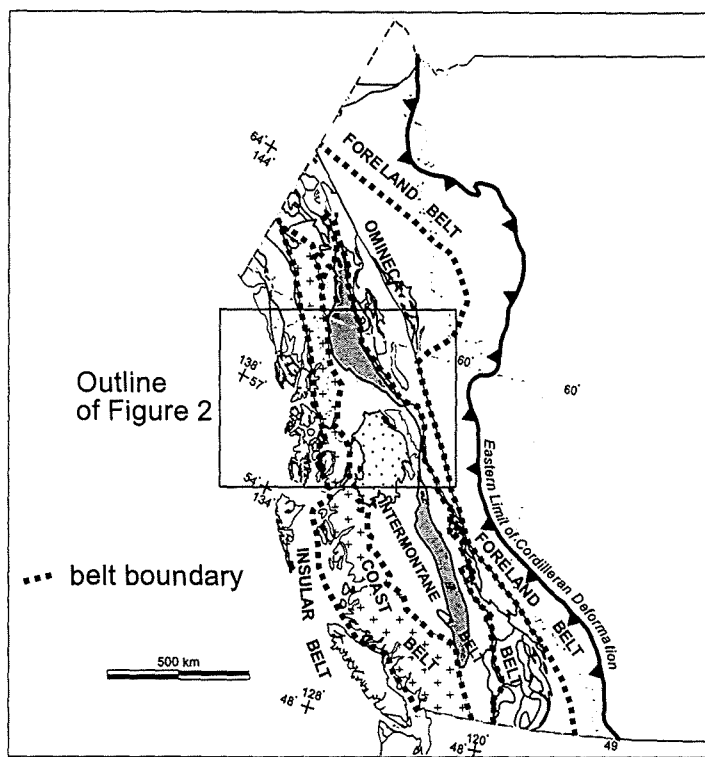


Figure 1. Major terranes, morphogeological belts, and location of Figure 2. Modified from Wheeler et al., 1991.

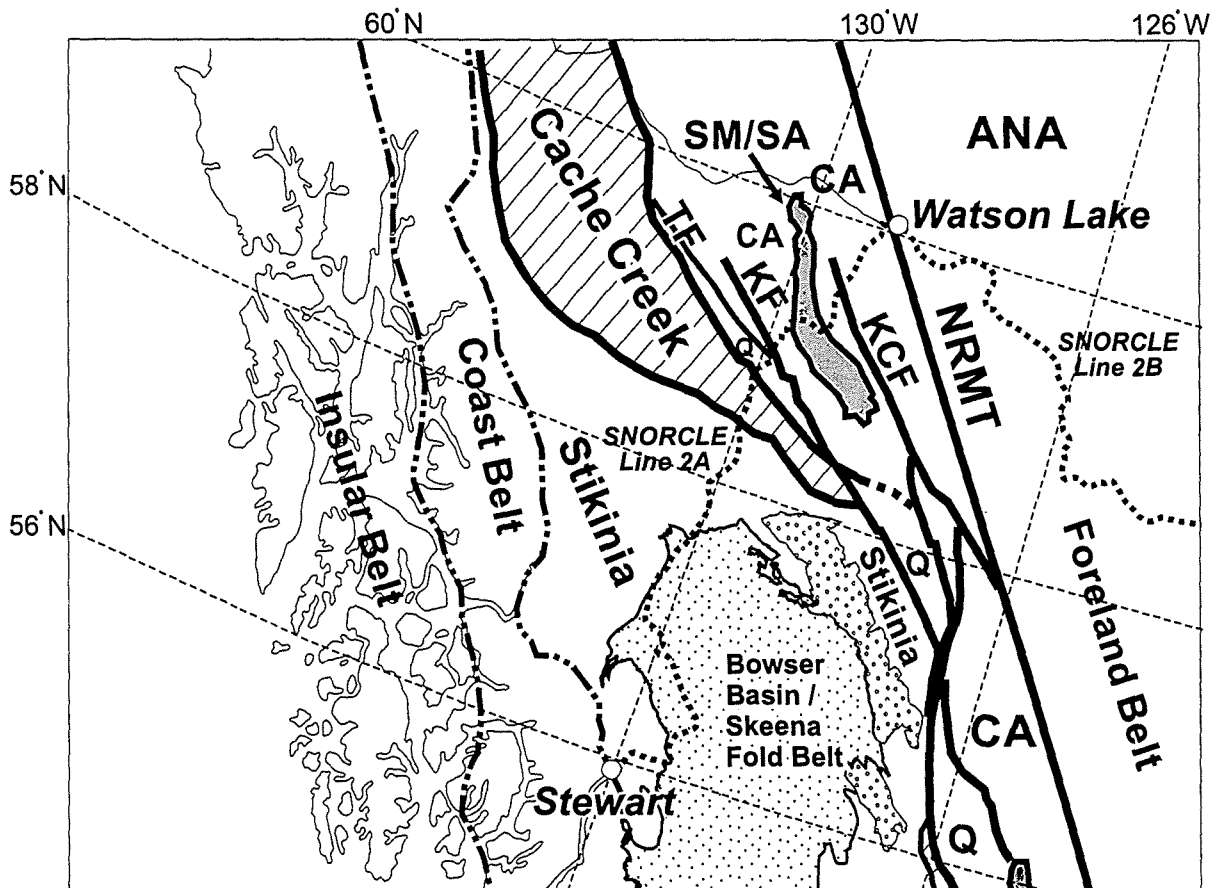


Figure 2. Location of SNORCLE Line 2A, and major features discussed in abstract (map is modified from Wheeler et al., 1991). Heavy lines are major faults, dash-dot lines are the boundaries between Insular, Coast, and Intermontane belts. Abbreviations: ANA Ancestral North America; CA Cassiar Terrane; SM/SA Slide Mountain Terrane, Sylvester Allochthon; Q Quesnellia; KCF Kechika Fault; KF Kutcho Fault; TF Thibert Fault

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Trans-Cordilleran Comparison between the SNORCLE Seismic Datasets: Near-Vertical to Wide-Angle Profiles of the Lithosphere along Corridor 2

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Introduction

The northern Canadian Cordillera has been traversed by two independent but coincident seismic experiments. In 1997, the SNORCLE Refraction Experiment (SNORE) acquired refraction and wide-angle reflection (R/WAR) data along the three SNORCLE corridors. These explosive-source data permit 2-D modelling of lithospheric velocities and of prominent impedance contrasts detectable from near-vertical to wide-angle incident angles. In 1996 and 1999, vibroseis reflection data were acquired along the same three corridors, providing relatively high-resolution images of impedance contrasts at near-vertical incident angles. These coincident near-vertical and wide-angle seismic datasets are complementary, providing constraints on structure and physical properties throughout the crust and into the upper mantle. This contribution pertains to Corridor 2.

The SNORCLE Corridor 2 seismic coverage was broken into two overlapping lines that took advantage of the major road corridors traversing the northern Cordillera. SNORE Line 21 and SNORCLE Line 2B cross ancestral North American lithosphere, following the Alaska Highway from near Fort Nelson, AB to west of Watson Lake, YK (Fig. 1). From east to west, the lines begin in the Proterozoic Fort Simpson magmatic arc terrane, cross the enigmatic Nahanni terrane, and continue across the Foreland belt - the margin of ancestral North America. The lines extend past the Tintina fault providing seismic coverage that penetrates into the Omineca belt.

SNORE Line 22 and SNORCLE Line 2A cross the western accreted Cordillera, following the Stewart-Cassiar Highway from Watson Lake, YK, to Stewart, BC (Fig. 1). From east to west, the seismic lines begin on the east side of the Tintina fault (ancestral North America), cross into the Omineca belt and continue south through the Intermontane superterrane reaching the boundary of Stikinia with the Coast belt plutonics. The SNORCLE Corridor 2 seismic data provide a direct link with the ACCRETE transect, effectively providing continuous seismic coverage across the Coast belt and Insular superterranes to the Pacific plate.

The combined analysis of these complementary seismic datasets will provide 2-D (in some locations 3-D) models of lithospheric structure and properties. Determining the sub-surface geometry of the allochthonous and autochthonous terranes and their relationship to ancestral North America is necessary to advance tectonic reconstructions. Refinement of models describing Cordilleran development may also lead to improved understanding of the fundamental processes (e.g., accretion, extension) influencing continental evolution.

Experiments

The preliminary migrated reflection images along Corridor 2 resulting from the SNORCLE vibroseis data are described by *Cook et al. (this volume)*. These data were acquired using a vibroseis source with sweeps of 10-80 Hz at 75 m shot intervals recorded by a 576 channel geophone array with 50 m group spacing. This provides a maximum 14.4 km offset in a symmetric layout. Nominal fold, or redundancy, is 192. The migration was carried out with one average velocity function along the line. The SNORE R/WAR data recorded arrivals from explosive sources (1000-3000 kg) using a fixed geophone array (~1 km spacing) with maximum offsets approaching 600 km. A description of the SNORE Line 21 experiment and an interpretation of velocity structure along it are given by *Welford and Clowes (2000)* and *Welford et al. (in review)*. SNORE Line

22, and in particular, its southern overlap with the ACCRETE transect, is described by *Hammer et al. (2000)* as part of their interpretation of the transition from the Coast belt to Stikinia. Preliminary interpretations of the complete Line 22 R/WAR data can be found in *Hammer and Clowes (1999)*. In both cases, the SNORE velocity structural models were primarily developed through travel time inversion and ray amplitude forward modelling using RAYINVR (Zelt and Smith, 1992). Gravity and magnetic modelling were also employed to provide additional constraints.

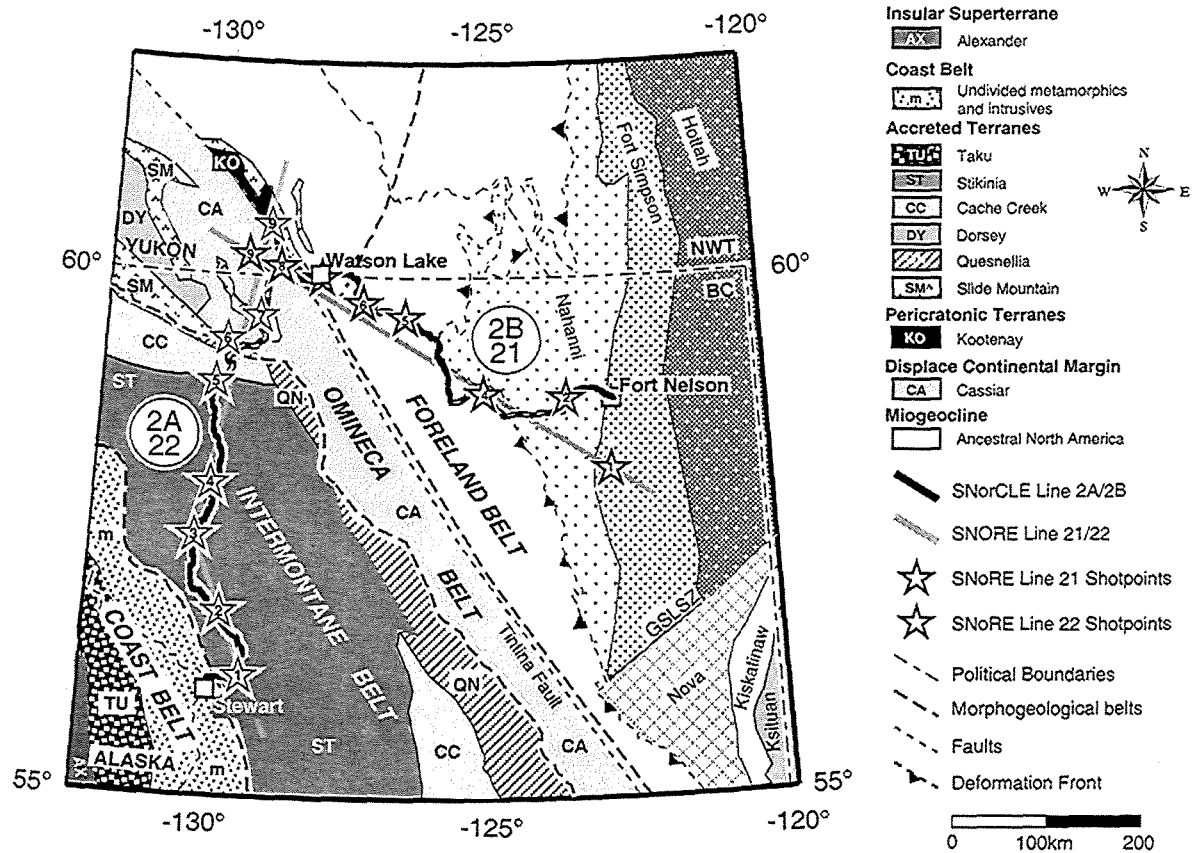


Figure 1: Terranes and geomorphological belts of the northwestern Canadian Cordillera (modified from *Wheeler et al. [1988]*). SNORCLE vibroseis reflection data were acquired along lines 2A and 2B (thick black lines); the processed vibroseis profiles approximately follow these corridors. The two profiles discussed in Fig. 3 are located by thin white dashed lines along the road corridors. The SNORE wide-angle data (Lines 22 and 21) were acquired along the same roads using explosive shots (stars) as energy sources. In order to model the large-offset SNORE data, they were processed along linear 2D profiles (thick grey lines).

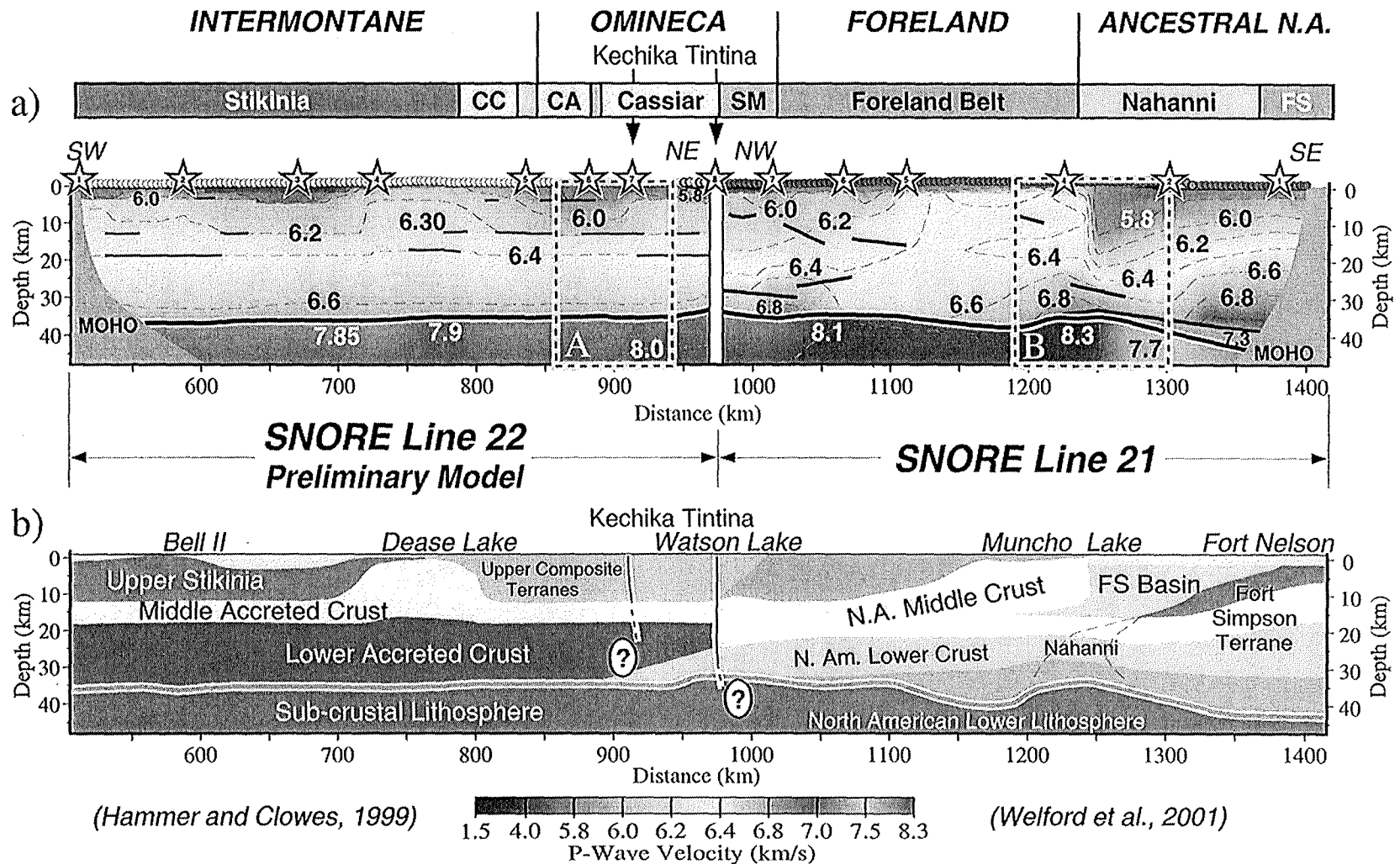


Figure 2.(a) Velocity structure model for the crust and upper mantle for SNORE lines 22 and 21 (modified from *Hammer and Clowes* [1999] and *Welford et al.* [2001]). Velocities are shaded and noted by numbers (km/s) and contours (dashed lines). The unconstrained model is masked out. Dashed boxes (A, B) demarcate the sections discussed in Fig. 3. Explosive shots indicated by numbered stars. (b) Interpretive section based on the velocity profiles. The Tintina and Kechika faults are noted. Terrane abbreviations are CC (Cache Creek), CA (Cassiar), SM (Slide Mountain), and FS (Fort Simpson). Vertical exaggeration 2.7:1.

The two seismic datasets sample the earth through different angular ranges and with different frequencies, thus providing complementary information about the structure and composition of the lithosphere.

- The SNORE R/WAR data are recorded to long offsets, sampling the upper crust and upper mantle with refracted energy that is used to model velocities that constrain lithology, temperatures and other physical properties. In addition, reflections with a broad range of incidence angles define lithospheric structures and are used to constrain the velocity model. Resolution varies with depth and location within the model, but velocities are typically constrained to within ± 0.1 to 0.2 km/s while spatial resolution is on the order of 1-3 km vertically, 10 km horizontally in the upper crust and up to 50 km in the upper mantle. Recorded frequencies are generally below 15 Hz, restricting the detection threshold for impedance contrasts to be greater than 100 m thickness at near-vertical incident angles. However, at such angles, with no redundancy, the amplitudes probably would be too small to observe. Wide-angle reflections used in the interpretation usually derive from substantive lithospheric boundaries.
- In contrast, the vibroseis multichannel seismic (MCS) reflection data is acquired using a short geophone array that restricts the reflected energy to near-vertical incident angles. However, the closely spaced receivers and source points provide redundancy to improve signal-to-noise ratios and much higher spatial resolution than the SNORE wide-angle dataset. Both vertical and horizontal resolution depend on velocity and frequency; horizontal resolution also depends on vertical travel time. In the upper crust (e.g., 4 s), vertical and lateral resolution are in the order of 50 m; in the mid-crust (e.g., 8 s), vertical resolution is similar but horizontal increases to about 100 m. At Moho depths (e.g., 12 s), vertical resolution decreases to about 100 m for a clear, strong reflection (not usually observed) with horizontal resolution to about 500 m. However, due to the need to convert travel times to depth, the lack of good velocity information until refraction values are considered, and the difficulty of identifying a specific "wiggle" (and hence travel time) with the Moho, its depth in kilometers can vary by up to 3 km. Velocities from MCS reflection data are only constrained within the upper 5-10 km by such data.

In summary, the near-vertical incidence dataset provides primarily detailed structural information (geometry); the R/WAR dataset provides a much coarser image of structure defined by velocity variations and contributes constraints for physical properties and lithologies.

Trans-Cordilleran Comparisons

The preliminary vibroseis reflection profiles along SNORCLE Corridor 2 are compared with the prominent features and interpretations of the R/WAR models. The final velocity structure model for SNORE Line 21 and a preliminary velocity model (middle and lower crust not finalized) for Line 22 are shown in Fig. 2, together with a schematic interpretation of these structures. *Cook et al. (this volume)* present and discuss the preliminary results from MCS lines 2B and 2A which were recorded along the same roadways. Some comparisons and observations about these two approximately co-located seismic surveys are noted below.

Foreland Belt / Ancestral North America (SNORE Line 21, SNORCLE Line 2B)

The lithospheric model for SNORE Line 21 (*Welford et al., 2001*) exhibits significant heterogeneities in velocities and crustal thickness (Fig. 2). The new reflection data provide images that constrain the geometry of lithospheric structures. Some of the principal observations and comparisons are:

- Western Canada Sedimentary Basin and Fort Simpson Basin
The R/WAR model clearly defines the Western Canada Sedimentary Basin (WCSB) and Fort Simpson Basin by a westward-dipping wedge of low velocities that extends to the edge of the Foreland Belt and depths approaching 20 km. The MCS data show coherent layered reflectivity over the same region with

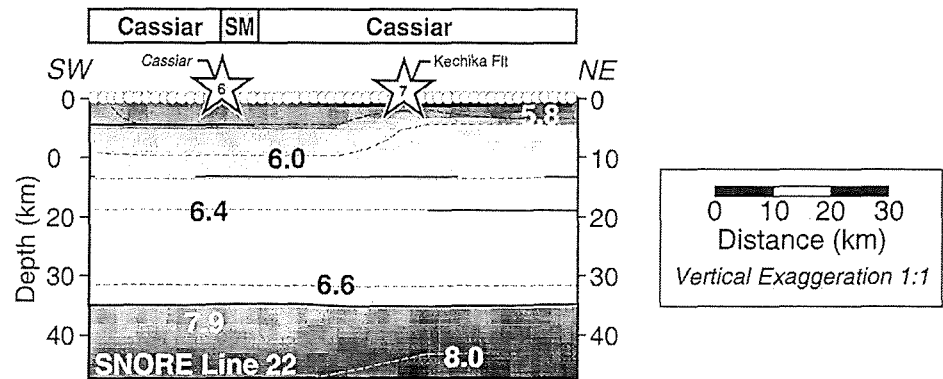
a distinct change in character at the approximate edge of the Foreland Belt. The two profiles correlate well.

- Low velocity and low density ramp below the eastern Foreland Belt (middle crust)
The velocity and density models indicate a subtle low velocity and low density ramp or tongue extending west from the sedimentary basins into the mid-crust below the eastern Foreland Belt. This feature correlates surprisingly well with a narrow zone of high reflectivity on the MCS data. On the velocity model, it is overlain by higher velocities, suggesting that the high velocity upper crust of the eastern Foreland Belt represents middle crustal rocks thrust over Proterozoic rocks of the Fort Simpson Basin.
- Eastern Foreland Belt
Data from shot points 4 and 5 of Line 21 generally were characterized as poor quality; arrival amplitudes were weak, suggesting either poor coupling of the geophones or strong attenuation within the upper crust. Approximately the same region along MCS Line 2B also exhibits a degradation in data quality.
- Lower crust
Throughout the length of Line 2B, the middle to lower crust exhibits strong reflectivity (with the exception of the zone of poor quality data and at the Tintina Fault as mentioned above). With a few modest exceptions, the velocity model shows reasonably uniform values from 6.4-6.6 km/s.
- Moho - the seismic crust-mantle boundary
Wide-angle Moho reflections (the PmP phase) along Line 21 showed the poorest quality of all the SNORE 97 R/WAR data. Along the eastern part of the line, no clear PmP phases were discerned and the interpretation showed a complex crust-mantle transition beneath the Nahanni and Fort Simpson terranes. Along MCS Line 2B, the Moho reflection is not distinct; it is generally indicated by a gradual cessation of reflectivity. A reasonably distinct Moho is only observed in the region immediately east of the Tintina Fault. In the region of the complex transition, the reflection Moho cannot be distinguished and incoherent reflected energy persists as deep as 14 s (~40 km). The refraction interpretation also indicates an increase in crustal thickness to 40 km in this region.

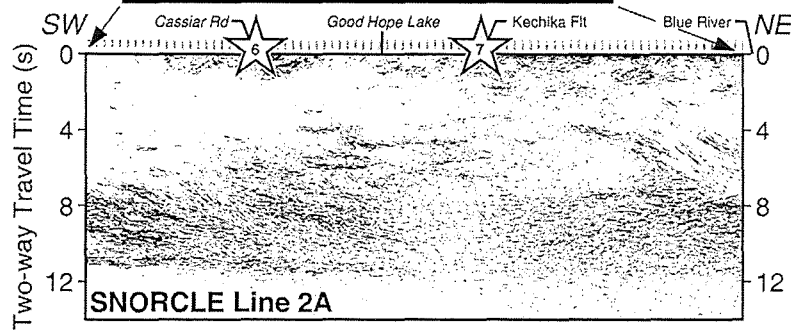
Tintina Fault Zone

The Tintina fault zone represents the surface expression of the division of ancestral from accreted North America. The SNORE R/WAR models for Lines 21 and 22 (*Welford et al., 2001; Hammer and Clowes, 1999*) show a prominent near-surface "notch" or zone of very low velocities which has been interpreted as the location of the Tintina Fault about 25 km west of the position normally drawn on maps, or as a splay fault associated with the Tintina system (much like the San Andreas has multiple strands in the San Francisco area). Also, about 25 km east of the low velocity zone on Line 21 and below the map position of the fault, the velocity contours show a distinct change in the middle crust, a change that could be associated with velocity differences across a crustal strike-slip fault. However, in general, the wide-angle data do not indicate prominent changes in velocity structure in the lower crust below the Tintina Fault. The reflection data do show a clear truncation of reflections that is attributed to the fault and a vertical zone of almost no reflectivity extending throughout the crust; the location extends from about 10-20 km west of the map position of the Tintina Fault. Both the refraction and reflection results indicate a slight thinning of the crust in the region below the fault. This is an interesting result because further southeast of this study area, where the Tintina Fault has a clear expression on the gravity map, it is defined by a narrow increase in gravity relative to values to the NE and SW, an increase that could be represented by a thinning of the crust at the fault (*Welford et al., 2001*).

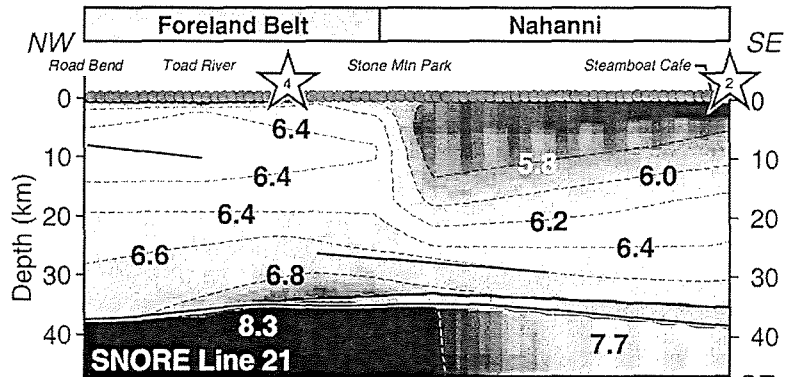
Ai)



Aii)



Bi)



Bii)

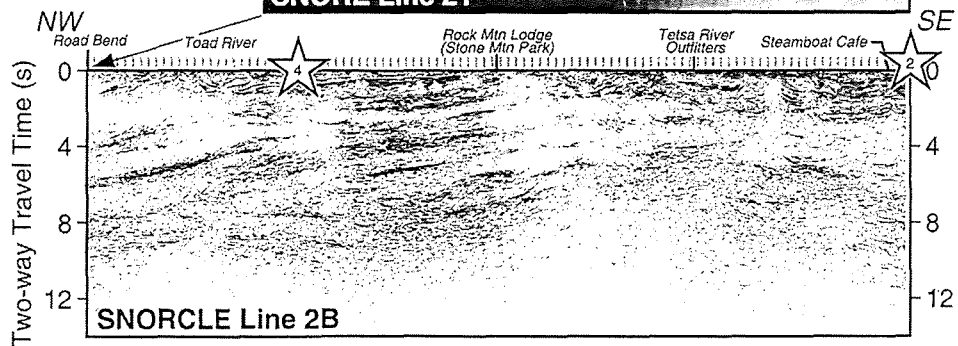


Figure 3: Comparison of (i) wide-angle velocity structure models (SNORE) with (ii) near-vertical incidence, migrated vibroseis profiles. (A) Profiles through the Omineca belt. (B) Transition from the Foreland belt to ancestral North America. Stars represent SNORE shotpoints; see Figs. 1, 2 for profile locations. Differences in profile lengths are due to the vibroseis sections following the road while the velocity profiles approximate a straight section plane (see Fig. 1). Vertical exaggeration 1:1.

Accreted North America (SNORE Line 22, SNORCLE Line 2A)

The preliminary crustal velocity model for SNORE Line 22 (*Hammer and Clowes, 1999*) exhibits significant lateral heterogeneity in the upper crust. The lower crustal velocities are currently modelled as being relatively homogeneous; however, these results are still preliminary and are being refined. The reflection data again provide images that constrain the geometry of lithospheric structure. Some comparisons and observations about these co-located seismic surveys are noted below.

- Omineca belt - upper crust

The R/WAR data require a complex, laterally variable structure throughout the upper crust of the Omineca belt. A prominent region of low velocities (~5.6 km/s) corresponds to the only part of the MCS reflection image where there is a lack of reflectivity. From the refraction data, an abrupt change in velocity across the Kechika fault zone is modeled; this is not required to continue through the crust. In contrast, the reflection data show a narrow zone of poor reflectivity extending to the Moho and bounded by strong reflectivity. In the lower crust, the bounding reflectivity has opposite vergence, indicating a crustal penetrating feature

- Intermontane belt (Stikinia) - Bowser Basin and upper crust

The SNORE velocity model requires relatively slow velocities (3.4-5.6 km/s) in the uppermost crust. These velocities extend to about 4 km depth and are interpreted to define the Bowser basin assemblage overlying Stikinia. A similar depth was derived from interpretation of magnetic data (*Lowe et al., 1998*). The reflection data show strong, undulating upper crustal reflectivity. One prominent band of reflections appears to extend to the surface at the northern end of the mapped Bowser Basin and continues to the south from times of 1-2.5 s in an undulating manner for at least 150 km. We associate this feature with the deformed base of the Bowser assemblage and/or deformation characteristic of the Skeena fold belt, which involved the upper crust of Stikinia (*Evenchick, 1991*). The region above the band of reflections generally corresponds with the region of low velocities.

- Lower crust

Modeling of the lower crustal structure is not yet complete for SNORE Line 22. The initial models require relatively slow velocities (6.65 km/s) that increase slightly toward the northeast. The most prominent large-scale feature on the MCS image is a tapering wedge of strong reflectivity that is observed at the northern end of the profile from times of 4 s to 11.5 s. This strong reflective band extends southward for more than 250 km with its upper surface becoming progressively deeper and the Moho forming its lower surface. At about the location of the Stikine River, the zone has died out. We interpret this prominent feature as representing a wedge of parautochthonous layers of Proterozoic strata that are correlative with strata of the Fort Simpson Basin (based on ties with Line 2B). This interpretation suggests that North American rocks extend below the accreted terranes almost to the Coast Belt. Cook et al. (this volume) describe the feature in more detail. The preliminary velocity model from the R/WAR data does not include variations that outline a similar feature.

- Moho - the seismic crust-mantle transition

In contrast with SNORE Line 21, strong wide-angle Moho reflectivity (PmP phase) is observed throughout Line 22, defining a clear boundary at ~35 km (with the exception of slight thinning below the Tintina Fault as discussed above). The reflection Moho is also very distinct for all of Line 2A except in the northern Omineca belt. In its time representation, it shows long wavelength undulations from 11 s to 12.5 s (variations of up to 4 km), which are different from the preliminary refraction model. The shallowest Moho travel times correspond to a region where the upper crustal velocities are low, so that a time-to-depth conversion would accentuate the undulations rather than compensate for them. However, the deeper reflection Moho below the Bowser Basin could be attributed to the low upper crustal

velocities described above. The more poorly defined reflection Moho for the northern 100 km of Line 2A is consistent in character and travel times with Line 2B, and approximates well with the depth to Moho on the velocity model.

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ELECTROMAGNETIC IMAGES OF THE TINTINA FAULT

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INTRODUCTION

The northern Rocky Mountain Trench – Tintina strike-slip fault (TTF) is a 2000-km long transcurrent fault along which an estimated 450-km or more of late Cretaceous to Oligocene dextral movement. Little is known about the subsurface expression of this feature. It is located within the northern Cordillera of western North America. The Cordillera comprises a region of oceanic and island-arc terranes accreted to North America since the neoproterozoic, with most of this westward growth occurring the last 200 Ma (Coney et al., 1980). Rocks to the northeast of the TTF are mainly sedimentary and represent the ancient North American margin. Rocks to the southwest of the TTF are mostly young, mainly igneous and metamorphic, and represent numerous accreted terranes.

The SNORCLE-Northern Cordillera MT experiment has three principal objectives. The first involves the determination of internal variation within the lithosphere. The second thrust concerns the characterization of the regional conductivity properties of the Northern Cordillera in relation to the major tectonic units. The third specific goal is the determination of the subsurface geometry and character of the Tintina Fault. In this work, we present results from the modelling and initial interpretation of three profiles crossing the TTF.

MAGNETOTELLURIC DATA

To determine the subsurface geometry and character of the TTF three regional MT surveys crossing the fault at different locations (Fig. 1) were carried out in the fall 1999. Profile 200 was measured along the Alaska highway and highway 37 (BC) close to the town of Watson Lake (Yukon), profile 300 along the Canol road close to Ross River (Yukon) and profile 400 along the Ketz River Mine road (Yukon). It is important to

note that between profiles 200 and 300 there are more than 400-km separation along the Tintina Fault. The Tintina Fault is covered by a thick layer of sediments (~2km) along profile 200, and sparser thickness along the other profiles.

The period range of the recorded data is between 0.0014 s and 10000 s, although in this work we are going to consider only the data between 0.01 s and 1000 s. Depicted in Fig 1C is the real induction vectors at $T=30$ s. The presence of a conductive structure in the middle of the three profiles can be deduced from this data.

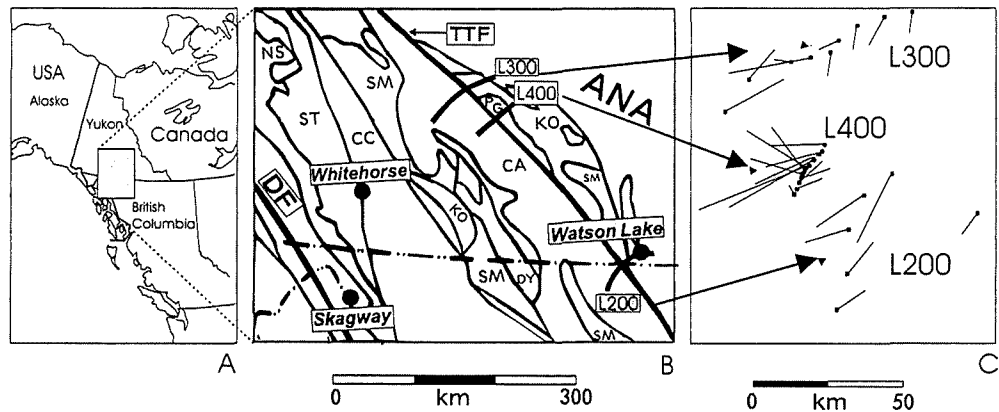


Figure 1. A: Map of western North America, shaded area corresponds to the location of figure 1B. B: Geological sketch of Southern Yukon and Northern British Columbia, showing the locations of the 3 MT profiles. TTF: Tintina Fault; DF: Denali Fault; ANA: Ancestral North America; CA: Cassiar terrane; CC: Cache Creek terrane; DY: Dorsey terrane; KO: Kootenay terrane; NS: Nisling terrane; PG: Pelly Gneiss terrane; SM: Slide Mountain terrane; ST: Stikinia terrane. MT profiles, L200: line 200; L300: line 300; L400: line 400. C: Real part of induction vectors (Parkinson's criterion, pointing to the conductors) for $T=30$ s.

To retrieve the strike of the regional structures and the regional impedance tensor we applied the Groom and Bailey method following the scheme of McNeice and Jones (2001). Figure 2 displays the strike directions as rose diagrams determined from the data at each site in the period band 0.01 – 1000 s in five decade-wide period bands. The individual site estimates of strike are weighted by the error misfit to the GB distortion model. The average GB regional strike for profile 300 is -50° and for profile 200 and 400 is -45° . The geoelectric strike determined from the analysis of the data are consistent with the main geological structures observed. The consistency of strike among the three profiles is also apparent in the data.

Figure 3 shows the apparent resistivity data for six selected sites, two of each of the profiles. The similarity between the data is conspicuous, in spite of the separation between sites (more than 400 km between profile 300 and profile 200).

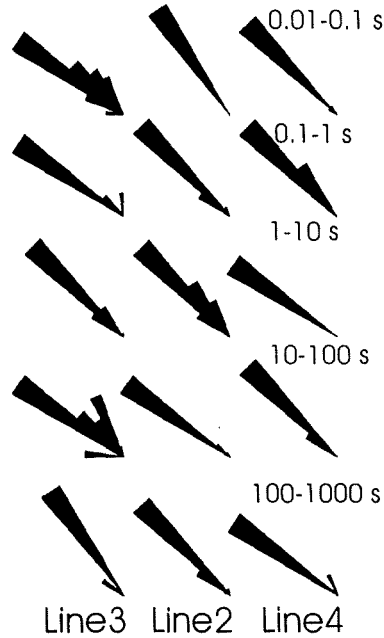


Figure 2. Rose diagrams giving the variation of the best fit for the regional azimuth to the Groom and Bailey distortion decomposition model for five different period bands between 0.01 and 1000s.

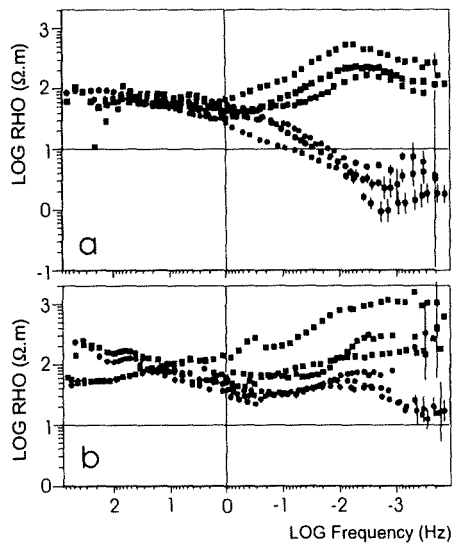


Figure 3. Apparent resistivity data, squares: TM-mode; circles: TE-mode.

2D INVERSION AND INTERPRETATION

For each profile, simultaneous 2D inversions of the TM and TE apparent resistivity and phases were undertaken using the algorithm of Rodi and Mackie (2001). We adopted a RMS figure of 5% higher than the minimum possible, fitting the apparent resistivity data to within 5% and the phases to within 1.4° on average. During the inversion, neither structural features nor conductivity discontinuities were imposed. Although the geomagnetic transfer functions corresponding to the TE mode were not used during the inversion procedure, the final models reproduce their main features. The resulting models are shown in figure 4, and the comparison between the data and model responses are portrayed in Fig. 5 as pseudosections of phases.

The similarity between the three models is noticeable, with the existence of high conductivity structures at several depths and scales. All three models exhibit conductivity structures ($\sim 1\text{-}10\ \Omega\cdot\text{m}$) below the surface trace of the TTF up to 5 km depth (label A in fig. 4) that can be associated with a fault system. To the SW, at deeper levels (below 5 km) an important conductivity structure ($\sim 1\ \Omega\cdot\text{m}$) is present (label B in fig. 4). In line 300, the extension to the SW of this structure is limited by the data located to the SW (not showed here) and it is not an artifact introduced by the inversion algorithm. To the NE another conductive structure is resolved (label C in fig.4), although the lack of sites to the NE in line 400 does not allow a good resolution of this structure. Finally, another striking feature of the three models is the occurrence of a low conductivity at deeper levels below the surface trace of the TTF (label D in fig.4).

The causes of the high conductivity zones in the models are not known at this point, although the origin of them could be related to the strike-slip movement of the fault and/or processes after the movement was finished. To gain some insight on the possible causes it is useful to review other magnetotelluric studies done in strike-slip faults. A high resolution study of the San Andreas Fault at shallow depths ($z < 3\ \text{km}$ depth) was undertaken by Unsworth et al. (1997). The main result is the presence of a high conductive zone in below the surface trace of the fault that has been associated with the presence of saline fluids. Another example is the crustal study of the Fraser River fault in BC done by Jones et al. (1992). This study reveals that the fault penetrates the entire crust with a subvertical geometry. And the cause for the enhanced conductivity at mid-crustal depths was associated to the presence of graphite generated by deeply penetrating surface waters.

Stanley et al. (1990) conducted a MT survey in Alaska crossing the Denali fault, another strike-slip fault in the Northern Canadian Cordillera and in Alaska. Their work resulted in the discovery of a thick unit of highly conductive rocks underlying parts of the Yukon-Tanana terrane extending to depths of greater than 20 km beneath relatively thin sheets of crystalline rocks. These authors concluded that the conductive units probably are Mesozoic flysch. However, Mathez et al. (1995) proposed the presence of interconnected graphite films to explain the observed conductivity values. A recent interpretation of Pb, Sr, Nd and O isotopic data of Mesozoic and younger granitic rocks close to the Denali fault (Aleinikoff et al., 2000), suggests that the high conductivity detected by MT soundings is not caused by the presence of flysch at depth, but may, in fact be due to interconnected carbonaceous films coating grain surfaces. Abraham et al. (2000), through the isotopic study of recent alkaline lavas, concluded that the Tinitina Fault is a high angle feature that penetrates the upper mantle and separates two distinct lithospheric roots.

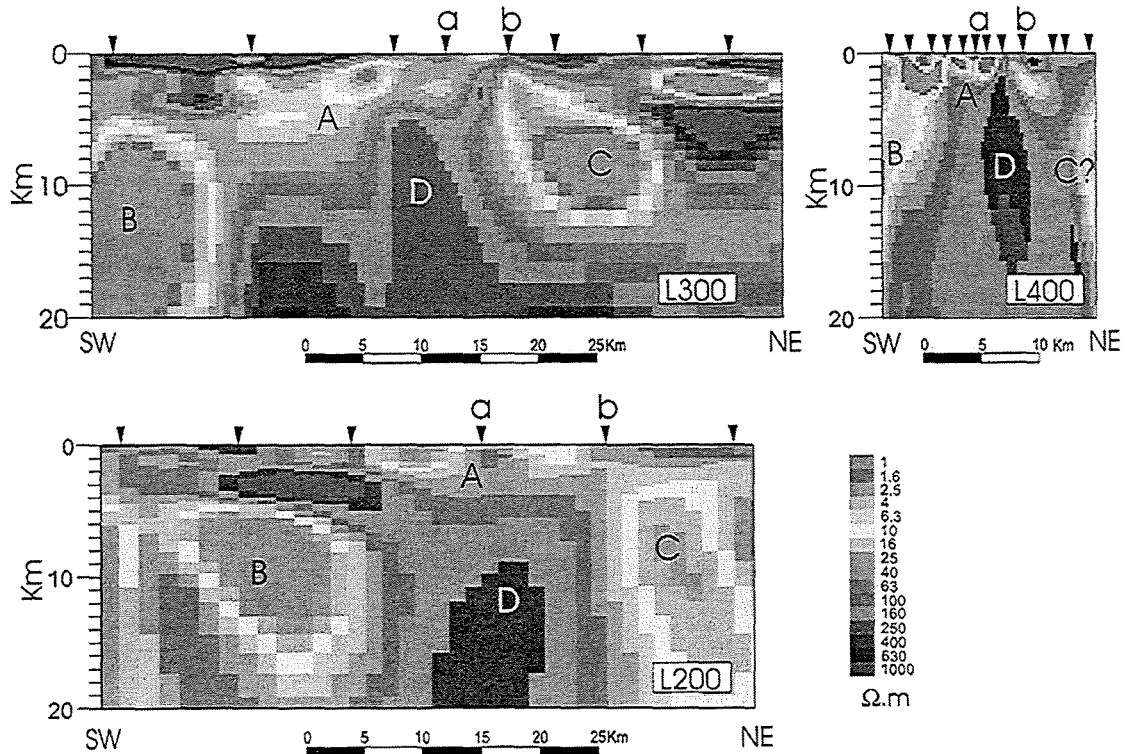


Figure 4. 2D MT resistivity models obtained by inversion using both TE- and TM-mode resistivities and phases. Triangles show the position of the MT sites, a and b correspond to the location of sites presented in figure 3.

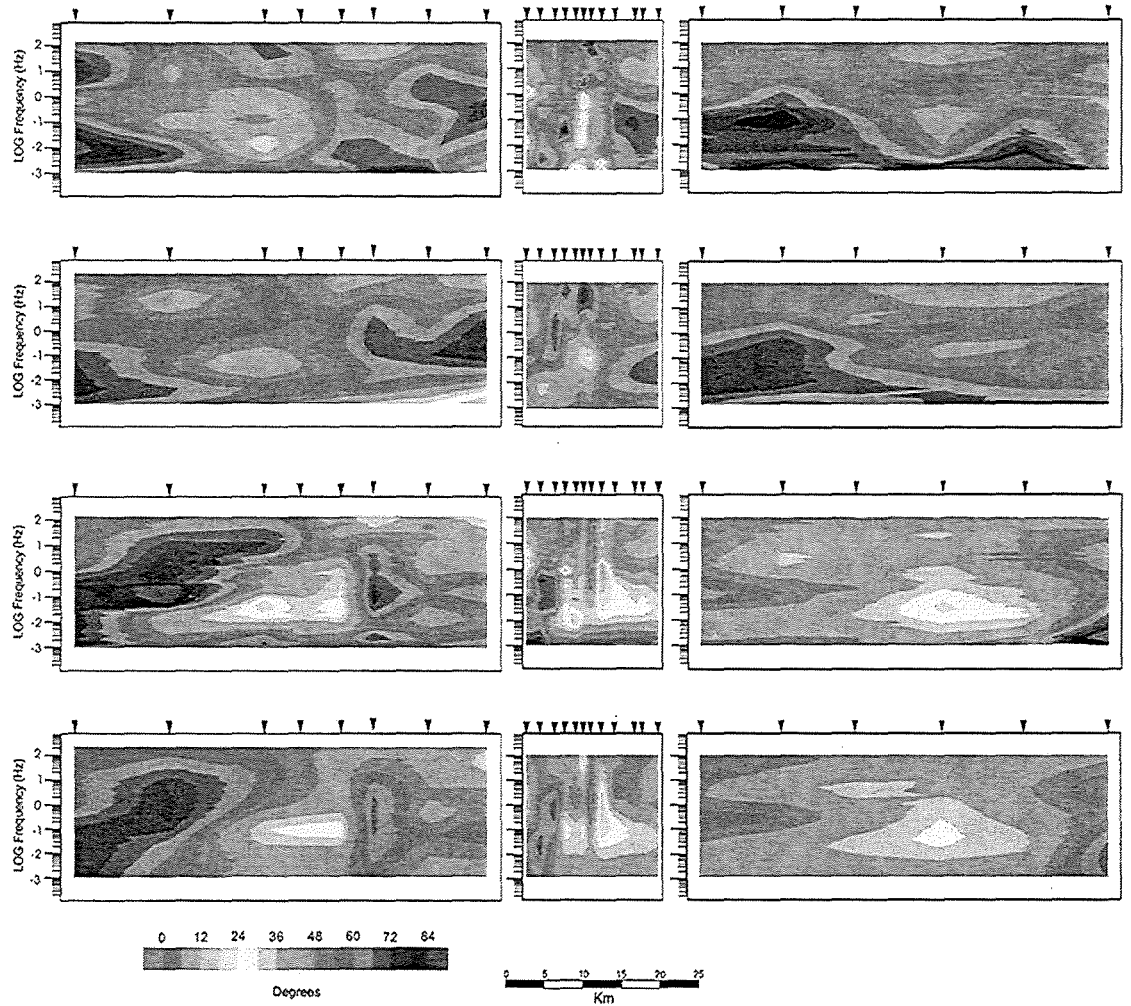


Figure 5. Comparison of phases from both TE- and TM-mode for the data (top) and model responses (bottom) for each profile.

CONCLUSIONS

We have shown that the electromagnetic response of the Tintina Fault is homogeneous along-strike for more than 400 km. This homogeneity has been observed both on the raw data and in the electrical conductivity models obtained. In the upper part

of the models ($z < 5$ km depth) the fault is imaged as a system of vertical contacts, suggesting that the TTF is formed by a system of faults more than for a single fault. At deeper levels, the presence of a high conductivity anomaly ($\sim 1 \Omega.m$) has been mapped.

Acknowledgements

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**IMPROVED STRATIGRAPHIC CORRELATIONS IN
THE INTER-TILLITE INTERVAL, WINDERMERE SUPERGROUP
(NEOPROTEROZOIC), MACKENZIE MOUNTAINS, NW CANADA**

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In Northwestern Canada, the Windermere Supergroup contains an exceptional record of the late Neoproterozoic opening and early evolution of the proto-Pacific Ocean. Indeed, the Windermere Supergroup contains one of the thickest, best-preserved, and best-exposed successions of terminal Neoproterozoic age to be found anywhere in the world. Over the past two decades, the lithostratigraphy of this succession has been progressively refined, to the point where formation-level correlations are generally very well constrained. However, an interval in which correlations remain problematic is the succession bounded below by the tillites of the Shezal Formation and above by the tillite-bearing Ice Brook Formation and the overlying "cap carbonates" of the Ravensthorpe Formation (formerly the "tepee dolostone"). This interval contains two formally defined formations, the lower Twitya Formation, a siliciclastic-dominated unit deposited in slope to shelf settings, and the upper Keele Formation, deposited on a carbonate platform. In some sections, the Keele Formation is capped by an informal unit of shelf siliciclastic rocks (the "Keele clastic wedge"). The total succession is up to 2 km thick. These formations record deposition during the interglacial period between two of Earth's most severe glaciations, and their association with tillites and cap carbonates potentially lends them significance in debates surrounding the "snowball Earth" hypothesis. In addition, the upper part of the Twitya Formation contains simple Ediacaran fossils that are, at present, the oldest-known metazoan fauna.

The chief difficulty encountered in the study of these rocks lies in correlating between relatively proximal and relatively distal sections. At more proximal localities (e.g., near Stoneknife River), the Twitya Formation records overall shallowing, from deep-marine environments (certainly below wave-base, and possibly on the continental slope) to shelf depths (with excellent hummocky cross-stratification). The contact with the overlying Keele Formation is gradational. The Keele Formation is dominated by a backstepping series of upward-shoaling packages with organic-rich, locally slumped basal beds and oolitic tops. A subaerial exposure surface of uncertain regional extent separates these beds from siliciclastic-dominated strata of the overlying "Keele clastic wedge". In proximal sections, the "Keele clastic wedge" is variably conglomeratic, sandy, or shaley, is locally calcareous, and contains numerous, closely spaced exposure surfaces and/or flooding surfaces.

By contrast, a much more distal section at Bluefish Creek (in which the Twitya Formation contains the Ediacaran fossils referred to above) is entirely siliciclastic and lacks any beds that could be assigned to the Keele Formation. Instead, the complete succession between the two tillite-bearing formations has been mapped as Twitya Formation. The succession consists of a lower part (roughly half the total thickness) dominated by slump-folded siltstone and minor sandstone, probably deposited in a slope setting. The upper part lies erosionally upon the lower. It is characterised by thick-bedded, massive to normal-graded conglomerate to conglomeratic sandstone, interpreted as thick-bedded turbidites, and lesser slumped siltstone and sandstone of probable turbidite origin. A distinctive feature of the conglomeratic facies is the presence of pebbles to cobbles of dolostone and limestone, as well as quartz and quartz arenite. This part of the succession is interpreted to represent deposition in a series of submarine channels. The assemblage of simple Ediacaran fossils comes from an interval of interbedded shale and sandstone within this part of the succession.

If the entire succession at Bluefish Creek is, in fact, equivalent to the Twitya Formation of more proximal sections, then the conglomeratic upper part would probably correlate with the relatively shallow-water upper part of the proximal Twitya. However, in proximal sections, the Twitya Formation contains very few carbonate beds. In view of the prevalence of carbonate clasts in the conglomerates at Bluefish Creek, such a correlation is problematic. We consider that the carbonate clasts are probably reworked material derived from the Keele platform during the exposure interval that preceded deposition of the "Keele clastic wedge." The conglomeratic upper part of the Bluefish Creek succession is thus more likely to correlate chronostratigraphically with the "Keele clastic wedge" than with the Twitya Formation.

Our observations and interpretations raise a number of issues. First, the typical Keele Formation is entirely absent at Bluefish Creek. This presumably reflects basinward thinning of the unit, related in part to its backstepping (transgressive) character. In addition, the erosional nature of the base of the conglomeratic interval suggests that erosion may have removed the thin, distal Keele, at least locally. It may be that the conglomeratic succession at Bluefish Creek represents the fill of a submarine canyon, an hypothesis best tested by regional mapping of the conglomeratic interval. A second issue relates to the age of the simple Ediacaran fossils preserved in the conglomeratic interval at Bluefish Creek. The beds in which these fossils occur probably do not correspond in age to the Twitya Formation as mapped in more proximal localities, but rather post-date deposition of the typical Keele Formation. Nevertheless, they remain the only sub-tillite Ediacaran fossils yet reported. Finally, the complexity of the sequence-stratigraphic record preserved in the "Keele clastic wedge" in proximal sections has implications for the Ice Brook glaciation. The presence of numerous, closely spaced sequence boundaries in this part of the succession suggests that the arrival of the glaciers at this locality was preceded by a protracted build-up phase with rapid climatic fluctuations.

could this cgl be
Keele and associated w/
rifting event?
→ Beaulieu Valley
→ my Hamill date 75 2nd glacial -

Cross sections of the allochthonous units west of the Cassiar batholith and the Sylvester allochthon: Pancake tectonic stratigraphy above the Cassiar Platform

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Recent detailed mapping of allochthonous rocks west of the Cassiar batholith, previously included within the Dorsey Terrane, has defined three tectonically distinct entities, and has delineated the relationships between them over large areas. Although structures within the units may in places show steep attitudes, major tectonic boundaries are generally very gently dipping, except in areas of local flexure, such as the Sylvester Allochthon on limbs of the McDame synclinorium and the western margin of the "Dorsey Terrane" east of the Seagull Batholith in southern Yukon. This configuration suggests that the top of the cratonic Cassiar Platform may extend westward at shallow depths as far west as the Teslin Fault.

The three subdivisions of the "Dorsey Terrane" are, from base to top, the Ram Creek assemblage, the Dorsey assemblage, and a late Paleozoic (and possibly older) stratigraphic sequence comprising the Swift River succession, Screw Creek limestone and Klinkit succession. The Ram Creek assemblage is an early to late Mississippian arc, probably built on continental basement. The Dorsey assemblage contains both continentally-derived coarse siliciclastic and basinal sediments, overlain by lower Mississippian arc strata. Its lower parts contain mafic to ultramafic sills with N-MORB to E-MORB petrochemical signatures. This suite is comparable to the mafic intrusions in the Lake George subterrane of the Yukon Tanana Terrane in Alaska, to early Paleozoic intracontinental igneous suites in the Selwyn Basin, and to sills in the Windermere Supergroup; they record early rifting of the continental margin. The entire Dorsey assemblage is intruded by early Mississippian arc-related plutons. Garnet-grade metamorphism is pervasive in it, and upper amphibolite facies metamorphic assemblages are in places well-preserved in its lower part. The Dorsey assemblage was emplaced on top of the Ram Creek assemblage along an easterly directed, post-mid-Permian, pre-Early Jurassic thrust fault (*see Stevens and Harms, 2000*).

The Swift River succession, lowest unit in the late Paleozoic stratified sequence, is dominated by chert and argillite, with subordinate siliciclastics and minor volcanic rocks. Present (structural) thickness is more than 4 km. Its base rests above the Dorsey assemblage. The area of the contact is strongly transposed, although no more so than rocks above it or, in particular, below in the thoroughly tectonised Dorsey assemblage. Nevertheless, the degree of deformation precludes confident identification of this contact as depositional, and indirect evidence suggests that it may be tectonic. The oldest age of the depositionally overlying Screw Creek limestone is only slightly younger than the youngest dated pluton in the Dorsey assemblage, leaving less than 10 Ma for the accumulation of a thick sequence of hemipelagic sediment. The top of the Swift River succession is overlain unconformably by the Screw Creek limestone; conglomerates and grits near the contact contain not only chert and quartz sandstone, but also plutonic (plagioclase and orthoclase) and metamorphic (quartz-sericite schist, muscovite, tourmaline) debris. Perhaps this marks accretion between the Swift River and Dorsey assemblage. In any event, they were together by Early Jurassic time, as both are cross-cut by the Nome Lake batholith. In the upper part of the stratigraphic sequence, the Screw Creek limestone is of Viséan to Bashkirian age. It is overlain by and interfingers with the late Paleozoic Klinkit succession, which contains pyroclastic and epiclastic rocks derived from a late Paleozoic (late Mississippian through Permian?) arc. The Screw Creek limestone and overlying arc strata resemble the Harper Ranch subterrane of Quesnellia.

Both regional distribution of units and exposed contacts between them show the overall gentle attitudes of the tectonic stratigraphy. Near the headwaters of the Cottonwood River, both the base of the Ram Creek assemblage on top of the miogeoclinal Earn Group, and the base of the Dorsey

assemblage on top of the Ram Creek, are essentially flat. In north-central Jennings River area, the allochthonous panels only dip significantly west near their eastern margin. Throughout most of the area dips are very gentle, and westerly-vergent recumbent folds dominate the stratified late Paleozoic sequence. In the southern part of the "Dorsey Terrane", the Dorsey assemblage outcrops in a horseshoe surrounding overlying units of the Swift River and Klinkit assemblages: this suggests a gentle synclinorium like the one occupied by the Sylvester allochthon.

The Sylvester allochthon consists of three major stacked tectonic units. 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 the Rapid River tectonite. The Rapid River tectonite contains siliciclastic and basinal sediments and mafic igneous rocks, metamorphosed in garnet amphibolite facies and intruded by early Mississippian and Early to mid-Permian plutons. It strongly resembles the Dorsey assemblage in terms of protoliths, structural history, and metamorphism; the intrusive bodies are an important element common to both .

The linkage between Rapid River tectonite and Dorsey assemblage allows construction of a speculative 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. In this section, the allochthonous units are exposed in two flat-bottomed synclinoria, with the region of the Cassiar Batholith an antinclinorium between. The combined stacking order is, from lowest and most inboard, to highest and most outboard: 1) Slide Mountain Terrane; 2) Paleozoic arc and carbonate strata, ranging from Early Mississippian in the Ram Creek assemblage to mid-Permian in the Sylvester Allochthon 3) Early Mississippian and Permian arcs superimposed on continental basement, the Rapid River Tectonite and Dorsey Assemblage; and 4) late Paleozoic basinal, arc and carbonate strata in the central part of the "Dorsey Terrane". 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 these cross sections 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.

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LATE PALEOZOIC IGNEOUS ACTIVITY ALONG THE WESTERN EDGE OF ANCESTRAL NORTH AMERICA, NORTH-CENTRAL BRITISH COLUMBIA

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Bedrock mapping along the westernmost edge of the Cassiar Terrane in north-central British Columbia has documented felsic volcanism of Early Mississippian age hosted in Earn equivalent rocks (Big Creek Group - Gilliland Tuff) and a Permian mafic dominated volcanic package in the Lay Range Assemblage. Recent age dating undertaken to better constrain these packages has resulted in the recognition of a new pulse of felsic volcanism of Late Permian age also hosted in the Big Creek Group. This new age data places further constraints on tectonic models for the evolution of this part of the Cordillera.

The Early Mississippian calc-alkalic felsic volcanics of the Gilliland Tuff are interpreted to have formed in a west-facing arc along the western margin of Ancestral North America. In an effort to more precisely determine its age, re-sampling of the type locality for U-Pb age dating was undertaken. Results are pending.

During the summer of 1999, in an effort to track volcanic rocks of the Gilliland Tuff northwards in the southern Wrede Range, semi-continuous lenses of felsic tuff up to 50 metres thick were delineated within the upper part of the Big Creek Group. While initial mapping suggested they were of Gilliland Tuff affinity, subsequent age dating determined them to be of Late Paleozoic age and therefore clearly not correlative with the Gilliland Tuff.

The felsic volcanics contain a strong penetrative cleavage, but still maintain recognizable fragmental textures and rip-up clasts of surrounding carbonaceous sediments suggesting their eruption on the sea floor. Whole rock and trace element geochemical data are limited, but indicate a calc-alkaline signature. A thin package of intermediate and mafic volcanics found immediately along strike yielded a single sample with trace element geochemistry of alkaline affinity.

The felsic tuff horizon was sampled for U-Pb dating at two localities. An interpreted crystallization age of 254 ± 5 Ma for one of these samples is based on the weighted mean of $^{207}\text{Pb}/^{206}\text{Pb}$ ages for two multigrain fractions of elongate zircon; one of these gave concordant results at the interpreted age of the rock and the other is inferred to have undergone minor Pb loss. Results of four analyses of stubby grains from the same sample are discordant and suggest the presence of inherited zircon. A best fit line projected through three of these from the concordant fraction at ca. 254 Ma gives an upper intercept of 2.5 ± 0.3 Ga, providing an estimate for the age of inherited zircons in the discordant fractions. The second sample yielded only enough zircon for four small multigrain fractions. Weakly discordant results were obtained for one fraction of elongate grains, which are inferred to have undergone minor Pb loss; it yielded a $^{207}\text{Pb}/^{206}\text{Pb}$ age of 254 ± 12 Ma, interpreted as the best estimate for the crystallization age of the rock.

Three fractions comprised of stubby grains yielded strongly discordant results and are interpreted to contain significant old inherited zircon (ca. 1.9 to 2.7 Ga).

Within the Cassiar Terrane volcanism of Late Permian age is relatively rare and its paleogeographic significance is not yet fully understood. Limited geochemical data suggest that felsic melts may have been produced in association with alkaline volcanism and are extensional in nature. However, many tectonic models of the Cordillera for this time period propose that the Slide Mountain ocean is collapsing immediately outboard of the Ancestral North American margin. One possibility is that the Permian felsic volcanics formed in an extensional arc or back-arc setting as a result of eastward subduction of Slide Mountain oceanic crust. However, the major problem with this hypothesis is that the polarity of subduction does not easily permit thrusting of Slide Mountain rocks onto Ancestral North America. Clearly, further investigation is required.

Middle and Late Paleozoic arc volcanics and pericratonic sediments of the Lay Range Assemblage lie structurally above the Big Creek Group. The Lay Range Assemblage is the local representative of the Harper Ranch Terrane which regionally forms basement to Quesnellia. It broadly consists of two packages: mid Mississippian to mid Pennsylvanian pericratonic sediments and volcanics of the Lower Sedimentary Division (LSD) and Permian mafic arc volcanics of the Upper Mafic Tuff Division (UMTD). The age of the Lay Range Assemblage, particularly the UMTD, is poorly constrained. Previous data from the UMTD consisted of two fossil localities; a single radiolarian Early Permian age (late Asselian to Sakmarian) near the base of the sequence and a collection of conodont fauna of undefined Permian age.

Recent mapping in the northern Lay Ranges indicated that the lower-most tuffs of the UMTD interfinger with mid-Pennsylvanian limestone. A felsic tuff, which is intercalated with mafic volcanics of this unit, was also collected for U-Pb dating. This sample yielded a small quantity of zircon which was split into four multigrain fractions. One of these yielded concordant results, with a $^{206}\text{Pb}/^{238}\text{U}$ age of 275 ± 2 Ma, interpreted as the best estimate for the crystallization age of the rock. The other three gave discordant results that indicate the presence of inherited zircon. Chords projected from the interpreted age of the rock (275 ± 2 Ma) through error ellipses associated with two of these fractions suggest average inheritance ages of ca. 500 Ma and 2 Ga. This is consistent with deposition of the Lay Range Assemblage on Precambrian (Archaen –Early Proterozoic) continental basement and overlying Late Proterozoic-Middle Paleozoic miogeoclinal sequences.

Age Constraints for Emplacement of the Northern Cache Creek Oceanic Crust and Implications of Youngest Blueschist Metamorphism

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Geochronologic and fossil age data presented here, together with a revision of the Jurassic time scale (Pálffy et al., 2000), permit unprecedented resolution of tectonic events that accompanied emplacement of the northern Cache Creek oceanic crust (Atlin complex). Muscovite and intergrown blue amphibole are developed in a coherent crustal block in which stratigraphy can be traced for several kilometres. Muscovite is dated by the stepwise ⁴⁰Ar/³⁹Ar technique at 174 Ma. The blueschist is unique in the Cache Creek terrane for three reasons: it is the youngest, it is the most extensive; and it is developed within a coherent crustal fragment.

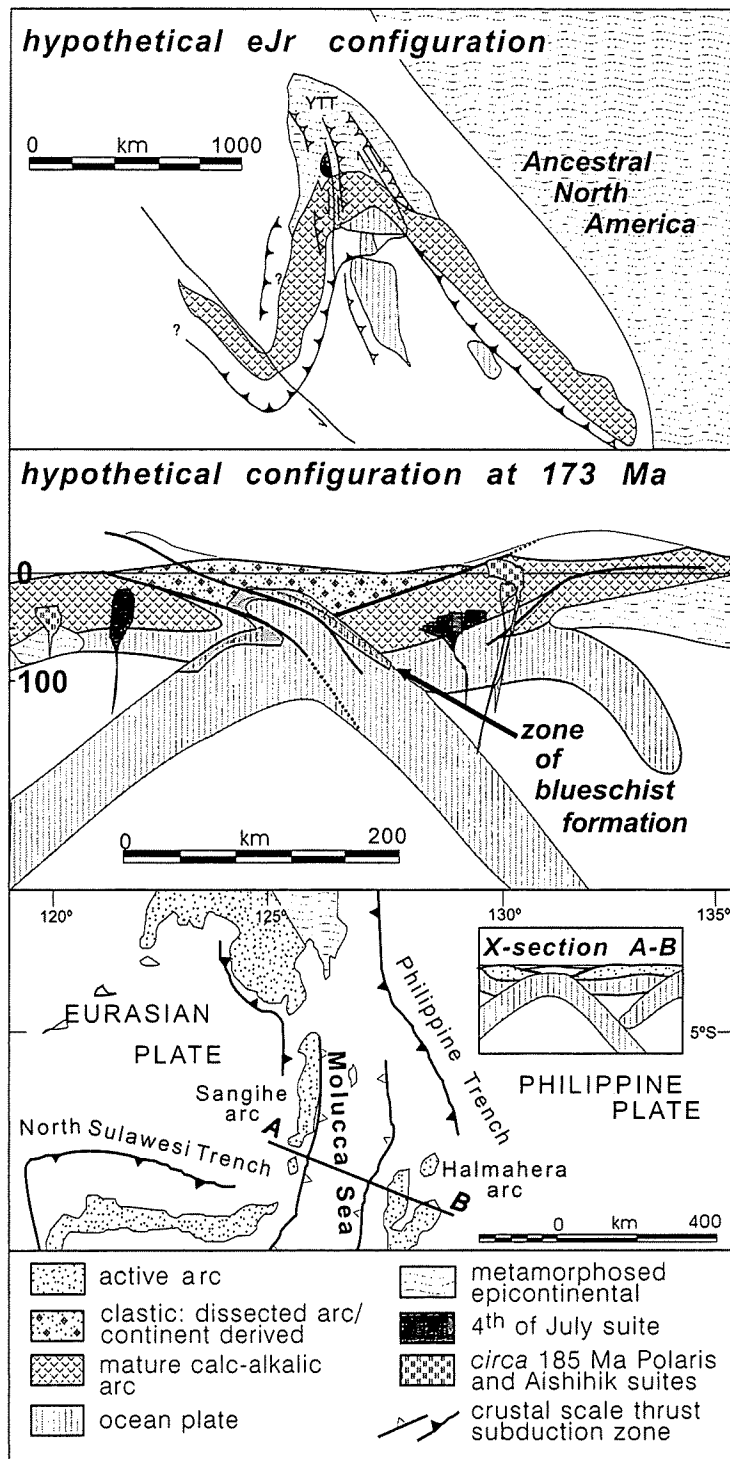
Emplacement-related fabrics in the Atlin complex are cut on its outboard edge by the *circa* 172 Ma Fourth of July batholith (Mihalynuk et al., 1992). Geological relationships indicate that its inboard edge is cut by a foliated quartz diorite that is dated by U-Pb methods at two localities to yield minimum ages of 167 and 171 Ma.

Sedimentological expressions of Atlin complex emplacement are recorded in the Early Jurassic Whitehorse Trough. Provenance and paleoflow indicators change in the Early Bajocian. Pre-Bajocian conglomerates record progressive dissection of the Stuhini arc to the west. In the Early Bajocian (*circa* 174-170 Ma), provenance and paleoflow switch and an eastern Cache Creek source and radiolarian-bearing chert granule conglomerates flood the basin. Taken together, the data suggest that blueschist metamorphism, tectonic emplacement and exhumation of the Atlin complex all took place between 174 and 170 Ma. Rapid exhumation may have occurred because of rupture of doubly subducting Cache Creek oceanic crust (Figure 1).

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Figure 1. Model for blueschist emplacement. A. Map view of hypothetical Early Jurassic arc configuration. Strata deposited in axial region between arcs included hemipelagite, wacke and chert sharpstone conglomerates near exposed fault scarps. B. Hypothetical cross section at about 173Ma shows rupture of doubly subducting Cache Creek oceanic crust. C. The Molucca Sea region and a modern day analogue of a double subduction zone with an exposed axial region.



The Bear River Dykes: A Westward Extension of the Mackenzie Dyke Swarm?

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The Bear River dykes are a suite of Middle Proterozoic mafic intrusions that crosscut Proterozoic dolostone and minor mudrock of the Wernecke Supergroup. They occur in at least five localities within the Wernecke Mountains of northeastern Yukon (Thorkelson and Wallace, 1993, 1994) (Fig. 1 and 2). The dykes are 5 to 15m wide, are up to 5 km long and are commonly flanked by a white-weathering metasomatic contact aureole. Dyke orientations are variable although the most common strike is approximately northwest, and dips are near vertical.

A dyke in the study area was examined in detail (Fig. 2). The petrology of this dyke is typical of other Bear River dykes and consists of fine to medium-grained gabbro, dominated by clinopyroxene and plagioclase. At this locale, two episodes of metasomatism are recorded: skarnification of the dolostone country rock related to intrusion of the dyke and post emplacement hematitic alteration of the dyke. This second zone of iron-enriched metasomatism is notable in that mineral features of the alteration are similar to those in a nearby zone of earlier-formed iron-oxide-enriched Wernecke Breccia. These breccia zones may also be characterized by localized enrichments of Cu, Co, U and Au (e.g. Laznicka and Edwards, 1979; Bell, 1986; Hitzman et al., 1992; Thorkelson, 2000). Evaluation of the geochemistry of the altered dyke using a Stonergram (Schwab and Thorkelson, in press) also reveals Cu and U enrichments and indicates that a later pulse of Wernecke Breccia style metasomatism was probably the source of this alteration.

Preliminary U-Pb baddeleyite dates on the Bear River dykes of ca. 1.27 Ga (J.K. Mortensen, pers. comm., 1995, Thorkelson, 2000) also have interesting regional implications for these intrusions. This age demonstrates that the dykes are time-equivalent with the largest dyke swarm in the world, the Mackenzie dyke swarm (Ernst et al. 1995, Fahrig and West, 1986). The Mackenzie dykes are part of the Mackenzie igneous event, which also consists of the Coppermine River flood basalts and the Muskox intrusion (Fig. 3). Mackenzie magmatism is thought to be the result of a mantle plume with a focal point on Victoria Island (LeCheminant and Heaman, 1989). A comparison of trace element patterns of the Bear River dykes to that of the Mackenzie dykes and Coppermine River basalts reveals similar geochemical signatures. In addition, there is significant overlap between incompatible immobile element ratios of the Bear River dykes and the Mackenzie dykes and Coppermine River basalts. If the similar ages and geochemistry are indeed indications that the Bear River dykes are part of the Mackenzie dyke swarm, the swarm would be 50% greater, and would extend 50° farther to the west, than previously recognized.

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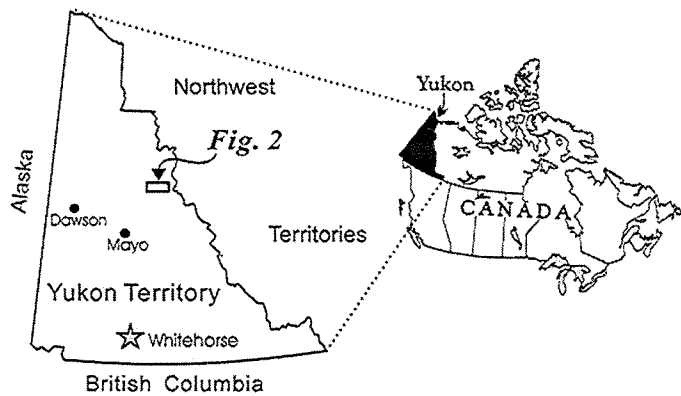


Figure 1 Location map showing region of Bear River dyke exposures in northeastern Yukon (area of Fig. 2).

Figure 2 Simplified geological map of Slat Creek, Fairchild Lake and "Dolores Creek" map areas (106C/13, 106C/14, 106D/16) showing schematic location of Bear River dykes and Wernecke Breccia zones (after Thorkelson, 2000). Study area indicated in northwestern part of map area.

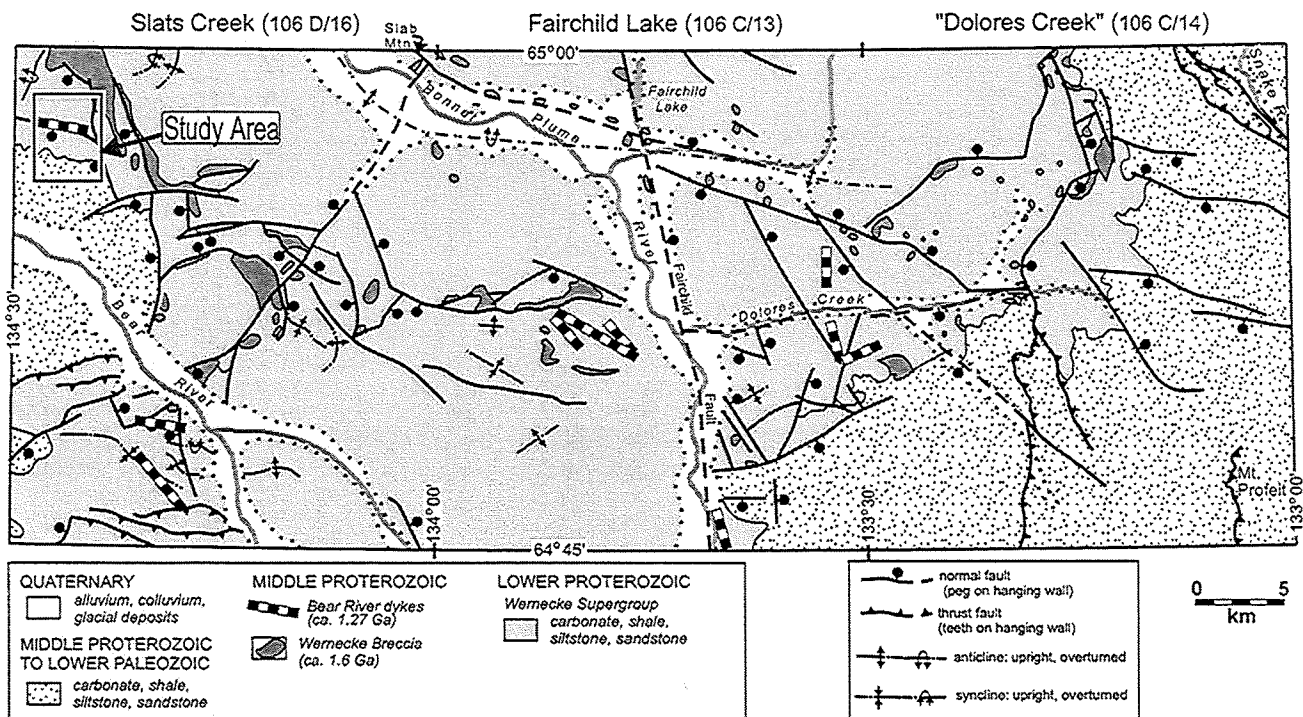
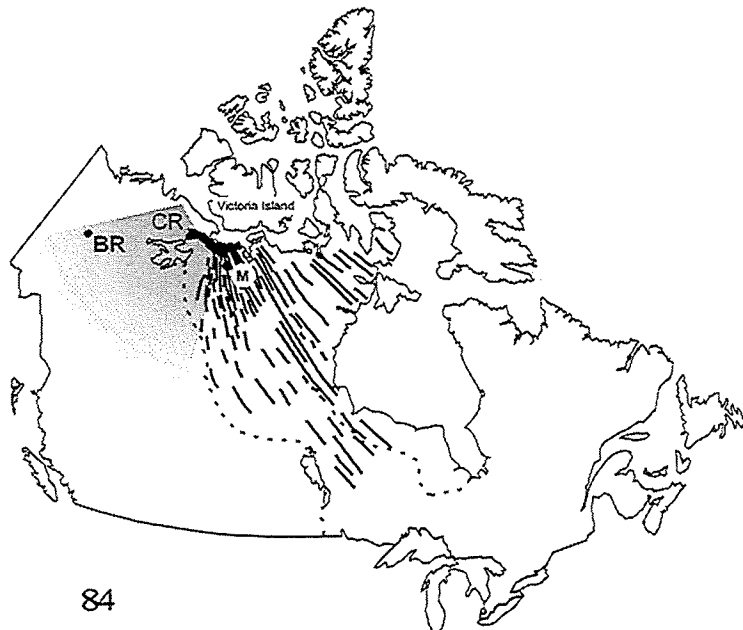


Figure 3 Principal features of the Mackenzie igneous event (modified from Hoffman, 1989). Mackenzie dykes are shown as thick lines. CR = Coppermine River basalts; M = Muskox intrusion; BR = location of Bear River dykes. Shaded area indicates possible westward continuation of dyke swarm.



Solitary Mountain, Yukon: Preliminary paleomagnetic results strengthen the correlation with the Carmacks Formation

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We report on the results of a multidisciplinary study of volcanic rocks that underlie Solitary Mountain, south central Yukon. The lower slopes of Solitary Mtn. are underlain by a variably oriented sequence of andesitic to basaltic andesite tuffs, flows, lahars, breccias and immature volcanoclastic rocks. These rocks pass up section into a series of relatively flat lying basaltic flows. This sequence is similar to that documented for the Carmacks Group in the Dawson Range, which is divisible into a lower fragmental sequence and an upper flood basalt. $\text{Ar}^{40}\text{-Ar}^{39}$ age determinations of the Solitary Mtn. volcanic sequence yields an age of 70.4 ± 0.7 Ma, consistent with correlation with the Carmacks Group. In addition, like the Carmacks Group, the Solitary Mtn. rocks are highly magnesium, potassic shoshonites, enriched in large ion lithophile and light rare earth elements, but depleted in high field strength elements.

The base of the volcanic sequence at Solitary Mtn. is not exposed. Foliated metamorphic rocks included in the Yukon-Tanana terrane crop out adjacent to Solitary Mtn. and rare clasts of metamorphic rock occur in the lower fragmental portion of the volcanic sequence. We interpret these relationships to suggest that the volcanic rocks rest disconformably atop the Yukon-Tanana terrane. Immediately west of Solitary Mtn. flood basalt flows of the upper portion of the Solitary Mtn. volcanic sequence immediately overlies schist of the Yukon-Tanana terrane – the lower fragmental sequence is absent. This relationship requires significant syn-depositional faulting.

Fourteen sites were collected for paleomagnetic study. Magnetizations are strong (0.1 to 100 A/m), but many of the stronger magnetizations were from rocks affected by lightning (low stability, random directions, high Koenigsberger ratio). None of the sites from the top of the mountain escaped the effects of lightning – a result discovered earlier in a preliminary study by Marquis and Erdmer. Despite the lightning, we observe that all sites have an underlying reverse polarity magnetization, consistent with the age (polarity chron 31R) and with the dominant polarity found in the previous paleomagnetic work on the Carmacks Formation (Marquis and Globberman, 1988; Wynne et al., 1998). Magnetization is carried by both titanomagnetite and pyrrhotite, and when the remanence is uncontaminated by lightning, the demagnetization characteristics are typical of ideal single-domain carriers. The 7 apparently uncontaminated sites give a mean direction which is insignificantly shallower ($5.4^\circ \pm 10.5^\circ$) than that found in the rest of the Carmacks Formation. In contrast, the Solitary Mountain remanence direction is $15.8^\circ \pm 9.6^\circ$ shallower than that expected from the mean Late Cretaceous pole of Cratonic North America.

This study extends the domain of rocks characterized by anomalously low Late Cretaceous paleomagnetic inclinations to encompass the Yukon-Tanana terrane extending from the Dawson Range in central west Yukon, east to where the terrane terminates against the D'Abadie and Tintina faults. If the anomalously low inclinations are interpreted as providing a record of paleo-latitude, it suggests that all of the Yukon-Tanana terrane lay some 2000 km south, relative to autochthonous North America, at 70 Ma. The D'Abadie fault, which places Yukon-Tanana terrane against the Cassiar platform, is plugged by mid-Cretaceous plutons and is interpreted as a Jurassic or older thrust fault, while the dextral Tintina strike-slip fault, which juxtaposes the Yukon-Tanana terrane against autochthonous North American strata, is characterized by 425 km of Eocene motion. Thus neither of these faults can accommodate the required northward displacement of Yukon-Tanana terrane. Resolution of the conflict between the geological and paleomagnetic data sets remains the focus of our research.

Preliminary paleomagnetism and geothermobarometry of the Granite Mountain batholith, Yukon: Minimal geotectonic motion of the Yukon-Tanana terrane relative to North America

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1. Introduction

Paleomagnetic research conducted in the Northern Cordillera through the support of LITHOPROBE-SNORCLE provides regional-scale constraints upon possible past terrane motions. The investigation of intrusive rocks in the Intermontane Belt terranes has refined estimates of the timing and magnitude of their poleward motion and clockwise rotation with respect to North America (Symons et al., *this volume*; Harris et al., 1999).

A complementary programme of study is underway for the pericratonic Yukon-Tanana terrane (YTT) to discern its geotectonic motion history and to test its coherency since Early Jurassic time. The YTT is a poorly-exposed assemblage of Paleozoic continental North American rocks which underlies much of central Yukon and eastern Alaska. Its size, cratonic provenance and 'inboard' position make the YTT a key tectonic element in the assembly of the Northern Cordillera (Figure 1), but its past relations to other far-travelled terranes and to North America itself are poorly understood (Mortensen, 1992; Lowe et al., 1994). The YTT may have acted as an *in situ* 'bumper' to halt the motions of other terranes (McClelland et al., 1992; Mihalynuk et al., 1994; Symons et al., 2000), or it may have participated, at least in part, with other accreted terrane motions relative to North America (Irving et al., 1996; Johnston et al., 1996).

Paleomagnetic results from the ~190 Ma Big Creek batholith of west-central Yukon imply that the YTT has experienced minimal net transport and counter-clockwise rotation with respect to North America since the Early Jurassic, markedly different than the significant poleward transport and clockwise rotations exhibited by units of the Intermontane Belt terranes (Symons et al., 2000). In this study, preliminary paleomagnetic and geothermobarometric results are presented from the nearby and similar-aged Granite Mountain batholith, as a further exploration of YTT history since the Early Jurassic.

2. Geological setting

The Early Jurassic Granite Mountain batholith is a melanocratic biotite-hornblende granodiorite intrusion of ~600 km² extent, emplaced in schists and gneisses of the YTT (Figure 2). It is intruded by dacite dykes related to the ~100 Ma Mt. Nansen Group volcanics, and is overlain, especially to the north, by the flat-lying basal andesite

flows of the ~70 Ma Carmacks Group (Tempelman-Kluit, 1984). The dextral Big Creek Fault forms the southwest limit of the Granite Mountain batholith, separating it and YTT host rocks from the ~190 Ma Big Creek batholith and the mid-Cretaceous Dawson Range batholith, a regionally-important large-volume intrusion. Across the southern boundary of the Granite Mountain batholith, the Carmacks Group volcanics of Millers Ridge lie on the downthrown side of the Miller Fault. To the northeast, Triassic volcanic breccia and tuff along the Yukon River is in contact with the batholith along the strike-slip Hoochekoo Fault. The batholith has given a K-Ar hornblende age of 142 $\pm$ 10 Ma (Stevens et al., 1982) and a U-Pb zircon age of ~191 Ma (Mortensen, pers. com., 1999).

3. Paleomagnetism

A total of 331 specimens from 24 sites were collected along mining roads in the southern and eastern parts of the batholith, and by helicopter along Big Creek in the northwest (Figure 2). The specimens underwent standard thermal and alternating field (AF) stepwise demagnetization at the University of Windsor Paleomagnetism Laboratory using a Sapphire Instruments SI-4 AF demagnetizer, a Magnetic Measurements MMTD1 thermal demagnetizer, and an automated CTF DRM-420 cryogenic magnetometer, all set inside a magnetically-shielded room which permits only 0.2% of the Earth's ambient magnetic field intensity. Additionally, 130 specimens underwent low-temperature demagnetization in field-free space prior to stepwise thermal demagnetization, to remove the multidomain magnetite contribution to remanence (Dunlop and Argyle, 1991).

Stepwise thermal demagnetization of 48 pilot specimens revealed distributed unblocking punctuated by an intensity drop in the 530°-580°C range, implying that the characteristic remanent magnetization (ChRM) is carried by low-Ti magnetite or magnetite. Specimens from many sites tended to break up in the oven at temperatures >470°C, failing along mica surfaces. Some surviving specimens exhibited anomalous strengthening of remanence intensity at temperatures of >530°C, likely due to the formation of new magnetic minerals. AF step demagnetization removed a viscous remanent magnetization (VRM) by ~15 mT, whereas thermal methods required ~260°C. ChRM directions were identified using the least-squares method (Kirschvink, 1980).

Paleomagnetic analyses yield three characteristic remanent magnetization (ChRM) directions that are carried by magnetite. In the north and west of the batholith, a ChRM **A** with mean direction $D=335$, $I=68^\circ$ (8 sites; $\alpha_{95}=7.1^\circ$) is present, although least-squares line fitting is practicable on only 45 specimens. A weakly-defined north-and-up ChRM **B** exists in sites 2,3 and 5 in the southeast. A Cretaceous andesite porphyry dyke (site 7) and its contact zone (site 6) give a westerly, steep down remanence direction, **C** (Figure 3). Specimens from other sites either did not carry a stable remanence, or had inconsistent directions from sample to sample.

Specimens selected for low-temperature demagnetization (LTD) at all sites except the andesite dyke (site 7) lost between 55% and 85% of their initial natural

remanent magnetization (NRM), suggesting that multidomain magnetite provides a significant contribution to remanence (Dunlop and Ozdemir, 1997; Warnock et al., 2000). This is of concern since multidomain magnetite can acquire a remanence at relatively low temperatures ($<200^{\circ}\text{C}$), but may require laboratory unblocking temperatures of up to 450°C to remove, significantly overlapping with the more resilient (and likely older) remanence carried by pseudosingle domain and single-domain magnetite (Dunlop and Ozdemir, 1997).

For most sites, LTD removed remanence directions that are steeply down to the northeast, in good agreement on a site-by-site basis with the direction and magnitude of the VRM *and* remanence that is removed by further demagnetization to ~ 20 mT and $\sim 450^{\circ}\text{C}$. Accordingly, ChRM directions are determined using specimens which underwent post-LTD thermal demagnetization, along with other specimens that exhibited stable remanence at unblocking temperatures of $>450^{\circ}\text{C}$ and coercivities of $>20\text{mT}$.

4. Paleodepth estimates

Emplacement depths are estimated (for possible tilt-correction of paleomagnetic data) using the Al-in-hornblende geobarometer (Anderson and Smith, 1995). After petrographic inspection of a thin section from each site, specimens from 9 sites were chosen to represent the sampled area of the Granite Mountain batholith. Electron microprobe analyses of hornblende and plagioclase pairs were performed using a JEOL JXA-8600 Superprobe at the University of Western Ontario (e.g., Symons et al. 2000).

Eight sites gave temperature-corrected pressure estimates for the batholith. Site 9 failed to provide useable plagioclase for temperature correction of the preliminary pressure estimate. Emplacement depths, calculated assuming a baric gradient of 0.036 km/MPa (density 2.8 g/cm^3), define two domains in the batholith: ~ 16 km in the northwest (3 sites), and ~ 18 km to the southeast (5 sites). The depth estimates imply either a west to northwest $\sim 6^{\circ}$ tilt of a coherent intrusive body, or a subdivision of the batholith into at least two blocks which experienced differential uplift.

5. Discussion

Given the estimated emplacement depths of 16-18 km, the ambient post-emplacement temperatures would have been $>480^{\circ}\text{C}$ ($\sim 30^{\circ}/\text{km}$ geothermal gradient), too high for an enduring remanence to be acquired by magnetite (Pullaiah et al., 1975). Any ChRM present was likely recorded during uplift through ~ 15 km depth. Simple time-linear uplift of the batholith from 16-18 km at 190 Ma to the emplacement of the overlying Carmacks Group flows at 70 Ma indicates that the uplift remanence would have been acquired at 180-170 Ma. This is tentatively suggested to be the age of the A remanence found in 8 sites. Alternatively, if the K-Ar system has not been reset in the hornblende, its 142 ± 10 Ma age may represent the passage of the northwest part of the Granite Mountain batholith through $\sim 500^{\circ}\text{C}$, thus implying that the A remanence was

acquired during the Late Jurassic.

Remanence B has no clear tectonic meaning, its direction plotting well away from the antipode to A, or any possible North American reference pole. It resides at high unblocking temperatures, but is poorly resolved.

The andesite dyke and contact zone carrying remanence C (sites 7 and 6, respectively) are of unknown age, but may be related to the emplacement of the overlying andesitic Carmacks flows. Their steeply down to the west ChRM compares well with the antipodal steeply up to the east ChRM recovered from 16 sites of the nearby 69 Ma Seymour Creek stock (McCausland et al., 2001).

Remanence A gives a paleopole of 72.4°N, 103.9°E ($dp=10^\circ$, $dm=12^\circ$), which has negligible rotation and is 440 km +/- 997 km near-sided compared with the 180 Ma North American reference pole of Besse and Courtillot (1991). It is ~700 km far-sided compared to mid- to Late Jurassic North American reference poles. If the batholith has coherently experienced a 6° tilt to the northwest, the tilt-corrected A paleopole would indicate only ~300 km of net polewards transport with respect to the 180 North American reference pole.

Both A and C remanence directions are similar to those expected for Early Jurassic and Late Cretaceous North American reference poles, respectively, and are most simply interpreted as representing minimal motion of the Granite Mountain batholith relative to North America since Early Jurassic time.

These preliminary results are similar to those obtained from the Early Jurassic Big Creek batholith which intruded at a slightly shallower depth and carries a similar uplift remanence direction in 18 sites (Symons et al., 2000). The Granite Mountain A paleopole is also similar to the normal-polarity Carmacks Group paleopole (Wynne et al., 1998), and may be interpreted to have been acquired during a thermal or hydrothermal event related to the Carmacks volcanism at 70 Ma (Figure 3). However, there is no evidence in thin sections from the northwest of the batholith to suggest a pervasive hydrothermal event. A thermoremanent overprint is also unlikely for the origin of the A remanence which unblocks at temperatures of > 450°C.

At this juncture, both A and C remanence directions of the Granite Mountain batholith are not determined well enough for a more rigorous comparison with other Cordilleran results. Remanence A is based on only 45 specimens from 8 sites and remanence C (dyke + contact) likely does not average out paleosecular variation of the Earth's magnetic field. These preliminary paleomagnetic results from the Granite Mountain batholith are most consistent with, but not rigorously supportive of, theories which postulate minimal tectonic motion for the YTT since the Early Jurassic.

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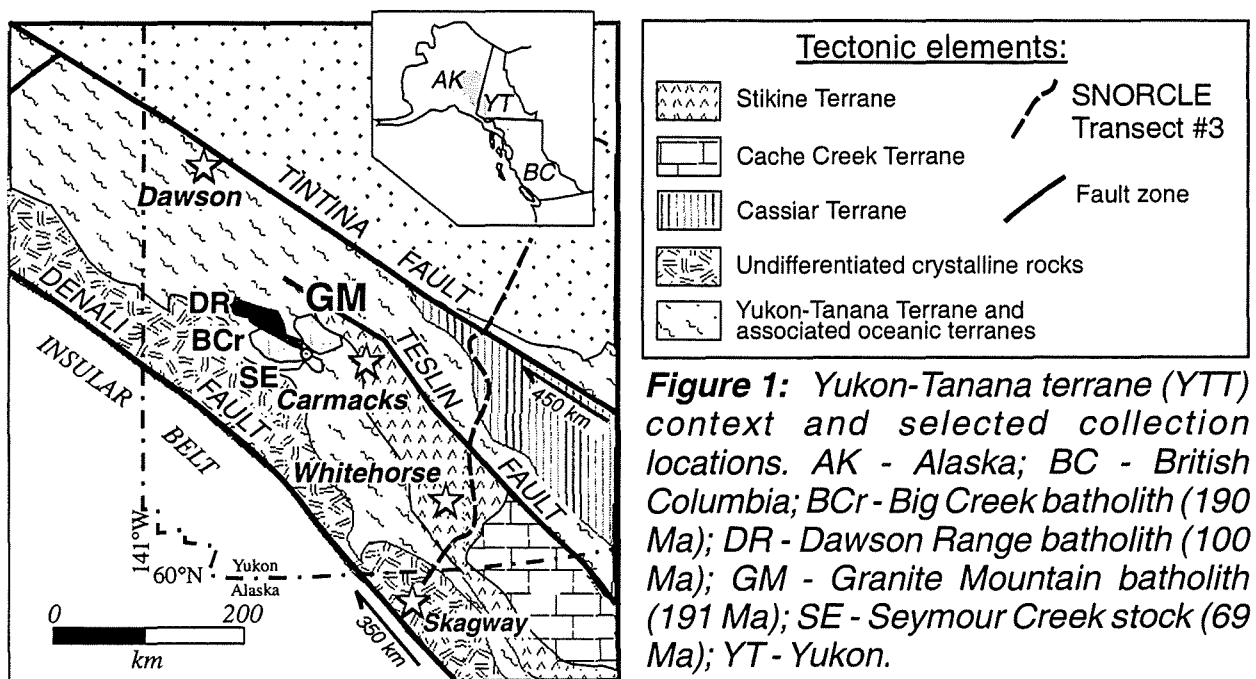
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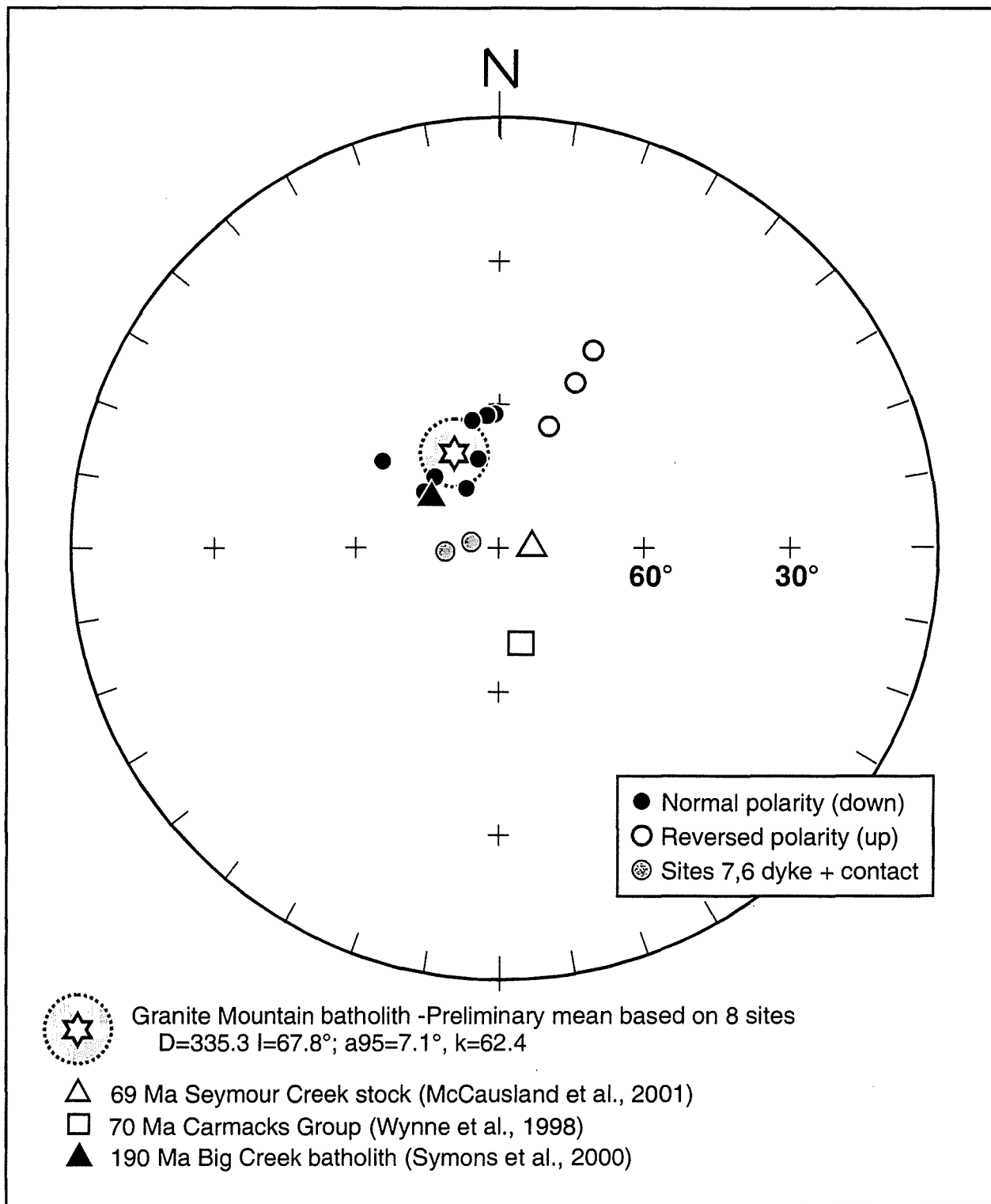


Figure 3: Granite Mountain batholith site and unit mean directions

A Preliminary Location for the Stikine Terrane at 53.6 Ma: Paleomagnetism of the Flat Creek Pluton in the Yukon

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Geology

The Flat Creek pluton is located ~50 km northwest of Whitehorse, YT (Fig. 1). It is a circular pluton with a diameter of ~10 km that belongs to the 57 to 54 Ma Nisling Range plutonic suite (Hart 1999; Fig. 2). The Flat Creek pluton is a recessive coarse-grained, smoky quartz-eye biotite granite with widely-spaced and horizontal joints. It has mariolitic zones that indicate a shallow emplacement depth and no evidence of post-emplacement deformation. The pluton has yielded a U-Pb age of 53.6 ± 0.2 Ma (Hart 1997).

The Flat Creek pluton is intruded into rocks of the Upper Triassic Povoas Formation, Lower to Middle Jurassic Tanglefoot Formation, and Late Cretaceous Carmacks Group that are part of the Stikine Terrane. The Carmacks Group is comprised of dark green massive basaltic lava flows with lithic tuff, ash flows and related volcanic rocks. These volcanics are flat-lying with dips of $<15^\circ$ that are randomly directed and likely primary in origin in most cases. They imply that the pluton has not been tilted since emplacement.

The Flat Creek pluton is cut by several narrow vertical Early Eocene andesite dikes with a northerly trend that give age dates of $\sim 52 \pm 2$ Ma. Their vertical attitude also implies no post-emplacement tilting of the pluton.

Paleomagnetism

The Flat Creek pluton was sampled at 25 widely-spaced sites by road and helicopter for paleomagnetic analysis. Of the 25, 21 were from massive granite, three from andesite dikes, and one from a dike's granitic baked zone. Each site was represented by six or more cores, yielding 12 or more specimens. First, for each site, two test specimens were thermally

demagnetized and two more were AF demagnetized in 12 or more steps. These specimens indicated that the former method was more effective and so the remaining specimens were thermally step demagnetized. The specimens from 20 of the 21 granitic sites yield a consistent population of high-temperature magnetite-borne reversely-polarized and steeply-inclined remanence components with a preliminary unit mean direction of Dec. = 162° , Inc. = -77° , $\alpha_{95} = 3^\circ$. Further, 11 of the 20 sites give an antiparallel normal low-temperature component with a preliminary unit mean direction of Dec. = 352° , Inc. = 77° , $\alpha_{95} = 4^\circ$. The apparent explanation is that the Earth's magnetic field went from reversed to normal polarity as this relatively large high-level pluton cooled.

The three dikes retain reversed characteristic remanence directions with a mean of Dec. = 158° , Inc. = -78° , $\alpha_{95} = 6^\circ$. Interpretation of the contact test is complex because the dikes and pluton are nearly coeval and because of the bipolar nature of the pluton's remanence.

Discussion

The unit mean direction for the reversed component of the Flat Creek pluton yields a pole position at $\sim 167^\circ\text{E}$, 81°N , $A_{95} = 5^\circ$. Using a reference pole of 169.5°E , 76.5°N on the apparent polar wander path of Besse and Courtillot (1991) for the North American craton at 54 Ma, the preliminary Flat Creek pole indicates a northward translation displacement estimate of $\sim 2 \pm 4^\circ\text{C}$ or $\sim 220 \pm 440$ km that is statistically nonsignificant plus a clockwise rotation estimate of $10 \pm 10^\circ$ that is significant.

Initially, the Flat Creek result can be compared with the White Pass dikes result (Symons et al. 2000). These mafic dikes are in the Coast Belt, inboard of the Coast Shear – Denali Fault system, just east of Skagway, AK. They are dated at $\sim 49 \pm 4$ Ma and are emplaced into four Early Eocene plutons with an “average” age of 53.5 ± 3.5 Ma. Using paleomagnetic data from specimens from 24 dike sites spread along an ~ 30 km long transect results in a translation estimate of $8 \pm 4^\circ$ southward and clockwise rotation of $40 \pm 9^\circ$. In this paper Symons et al. (2000) summarized data from ~ 200 sites in six collections of ~ 50 Ma volcanics, dikes and plutons throughout the Intermontane and Coast terranes inboard of the Coast Shear Zone. On average they yield a translation estimate of $3 \pm 5^\circ$ southward and $24 \pm 10^\circ$ clockwise. Two basic thoughts emerge.

First, the White Pass dikes and the Intermontane data overall indicate minor southward translation whereas the Flat Creek pluton result implies minor northward translation. Overall this modest difference could be either nonsignificant or indicate minor northward translation between ~54 and ~50 Ma. Regardless, it does cast further doubt on the results to date from the 70 Ma Carmacks Group volcanics that Wynne et al. (1998) interpret to record 1900 ± 700 km of northward displacement between 70 and 45 Ma. However, it does provide a good fit with the results from the 75 Ma Mount Lorne pluton, ~30 km southeast of Whitehorse YT, that yield a translation estimate of 1170 ± 390 km northward between 75 and 50 Ma (Harris et al. 1999).

Second, this initial estimate of $10 \pm 10^\circ$ of clockwise rotation for the Flat Creek pluton is consistent in clockwise sense and comparable in magnitude with the White Pass dikes' estimate of $40 \pm 9^\circ$ and with the overall ~50 Ma estimate for the "Intermontane" terranes of $24 \pm 10^\circ$. Clearly some post-50 Ma mechanism for clockwise rotation, such as distributed strain (Beck 1998) or sole fault rotation (Symons et al. 2000), must be incorporated into models for the tectonic evolution of the Intermontane terranes.

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Fig. 1. Location of the Flat Creek study area (after Symons et al. 2000).

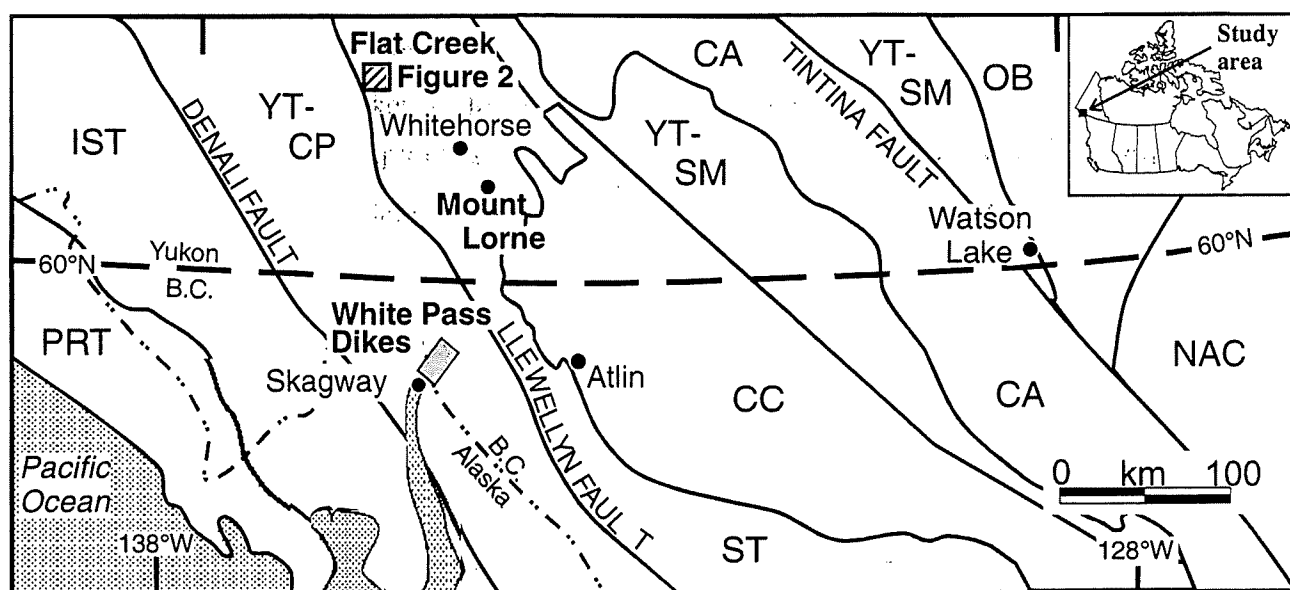
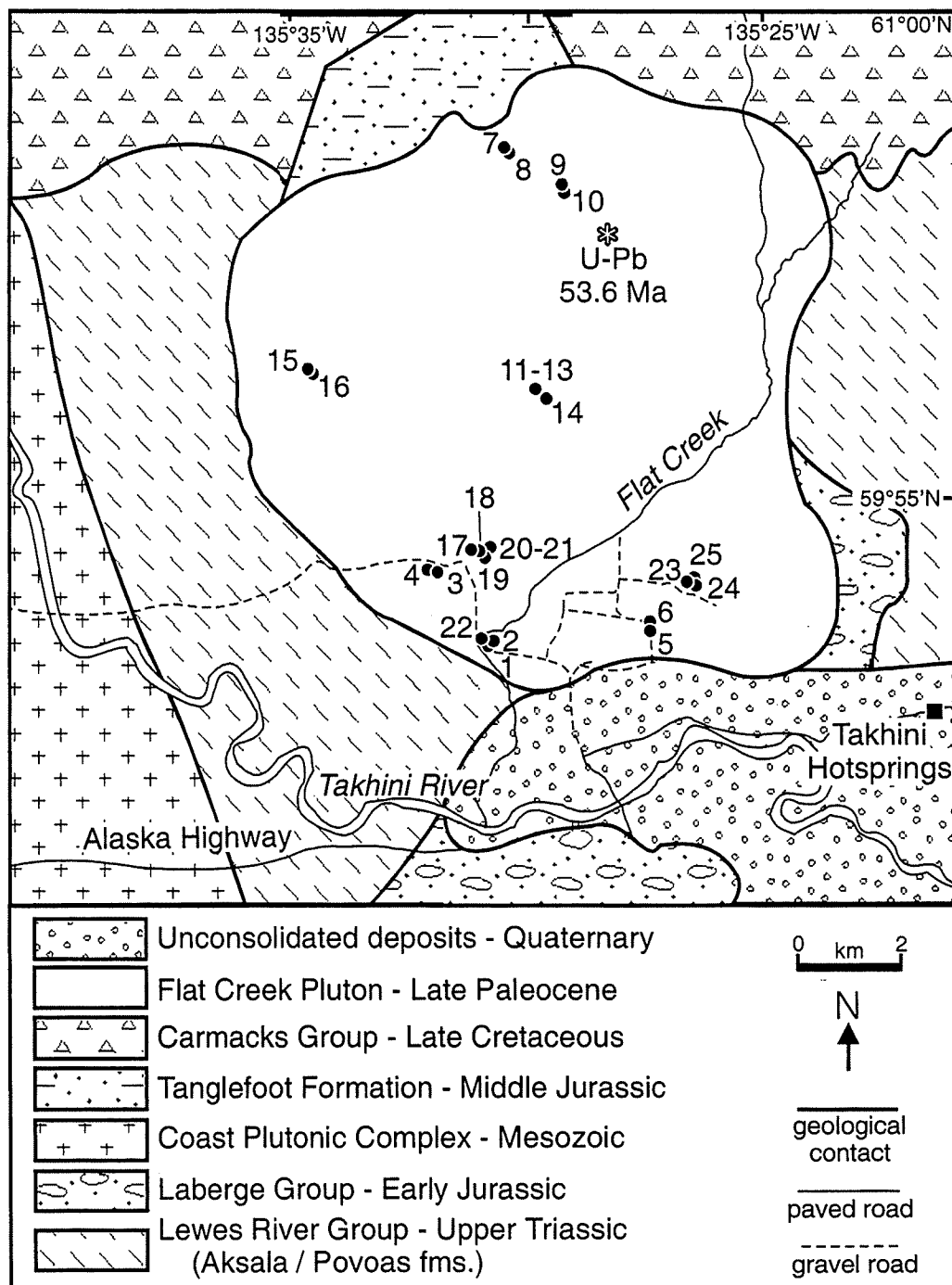


Fig. 2. Geology around the Flat Creek pluton after Hart (1999) with sampling sites.



Magmatic fabrics in batholiths as markers of regional strains and plate kinematics: example of the Cretaceous Mt. Stuart batholith

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The Cretaceous Mt. Stuart batholith was syntectonically emplaced within amphibolite grade metasedimentary rocks of the Cascades Crystalline Core, Washington State. The northern part of the batholith defines a NW–SE trending antiformal fold. We present fabric data from that part of the batholith, collected during field mapping and anisotropy of magnetic susceptibility (AMS) measurements. The significance of the data is discussed in terms of regional tectonic deformation and plate kinematics. The data were collected from rocks with well preserved igneous textures and the fabrics therefore formed during magmatic deformation. The AMS provides measurements of the preferred orientations of Fe-rich minerals (biotite $\pm$ hornblende $\pm$ traces of pyrrhotite and magnetite) which are consistent with field measurements of the mesoscopic fabrics defined by plagioclase, biotite and hornblende crystals. The magnetic fabrics are also consistent with the orientations of folds, mineral fabrics and boudinage structures that record high-temperature subsolidus deformation in the margin of the pluton and in its host rocks. The lineations are parallel to the stretching direction associated with small increments of strain that occurred during deformation of the magmatic arc, as the batholith was crystallizing and deforming in the tectonic stress field, *ca.* 93 Ma. The fabrics in the Mt. Stuart batholith are used to infer emplacement in a magmatic arc during either 1) plate displacement perpendicular to a NW-SE trending plate margin, or 2) wrench dominated transpression. In the second case the analysis suggests a nearly N-S plate vector along the western North American margin during plutonism. The results demonstrate the potential usefulness of magmatic fabrics in syntectonic plutons for plate tectonic analyses of orogenic belts.

Tertiary Crustal Scale Thinning across the Coast Mountains, near Prince Rupert, B.C.

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Post Cretaceous tectonics of the Coast Mountains near Prince Rupert, B.C. have been dominated by extensional strain fields, crustal thinning and exhumation of mid to deep crustal rocks. The multidisciplinary ACCRETE project identified an important extensional event that occurred within the Coast Mountains during the Early Eocene. ACCRETE geological studies and seismic profiling showed that a crustal scale gneiss dome extends from the Coast shear zone (CSZ) on the west to an east dipping normal detachment system on the east. The orientation of the dome and structures found within the detachment system indicate NE-SW stretching during the Eocene. West of the CSZ, ACCRETE also imaged structures in 3-D that are attributable to a previously recognized episode of Miocene extension; this extension was oriented EW. Although these structures are best displayed in seismic images west of the CSZ, they are also recognized east of the CSZ in the seismic data and as brittle high angle faults at the surface. The faults east of the CSZ have orientation and sense of displacement that are consistent with late Tertiary EW extension.

Petrologic, age dating, and structural studies show that there was about 20 km of exhumation during the Eocene. Taking the 5-10 km of previously recognized post Miocene exhumation plus the present, seismically determined 30 km of crust, the net thinning of the crust was from 55 km to 35 km during the Eocene (30-40%).

Our results show that most of the post Cretaceous extension in the Coast Mountains occurred in the Eocene and was most intense east of the CSZ. The terranes both west and east of the CSZ were then thinned during the Miocene and later extension. The Miocene extension is partially accommodated by arching in the Moho (1-2 km amplitude) and by grabens recognized at the surface. These features occur equally developed on both sides of the CSZ.

Yakutat Terrane Collision in the Gulf of Alaska, Northwestward Translation of the Northern Cordillera, and Present Thrusting and Seismicity at the Eastern Mountain Front

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The present active tectonic regime in the northern Canadian Cordillera provides some interesting lessons for understanding past Cordilleran deformation. The Yakutat oceanic plateau terrane is presently colliding with North America in the northeastern corner of the Gulf of Alaska, after having moved north along the Pacific coast. The Yakutat-North America relative motion is partitioned into northwesterly strike-slip motion and northeasterly convergence. The strike-slip motion is mainly on the Fairweather-Border Ranges faults near the coast, with small present motion on the Denali fault. The convergence is partly taken up by crustal shortening and thickening near the coast and partly by translation of the whole northern Cordillera to the northeast relative to the craton. The crustal shortening at the collision boundary near the coast has resulted in rapid uplift and strong present seismicity in the St. Elias-Mt. Logan region. This region is our mini-Himalayas. Surprisingly, however, deformation is not limited to this coastal region. Part of the convergence is accommodated at the eastern Cordillera mountain front as thrust shortening and strong seismicity. GPS geodetic data at Whitehorse indicate motion to the NNW at 2-4 mm/yr (km/Ma) relative to the craton (Yellowknife). The seismicity in the McKenzie Mountains near the eastern mountain front has an average seismic moment rate that also requires several mm/yr of NW motion. This process of oceanic terrane collision and transfer of motion completely across the Cordillera to produce overthrusting of the craton may be a good analogue to what happened in a previous collision on the coast of the southern Cordillera. The Crescent-Siletz terrane collided against Vancouver Island in the late Cretaceous-Eocene. This collision appears to have resulted in strong uplift of the west coast of Vancouver Island and, across the southern Cordillera, in a phase of thrust shortening at the Rocky Mountain front.

Retrograde mantle flow beneath the southern Canadian Cordillera: An explanation for both the enigmatic late Neogene uplift of the Canadian Rockies and the enigmatic Early Jurassic obduction of Slide Mountain terrane and disappearance of Slide Mountain basin?

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Abstract

The contemporary high topographic relief of the southern Canadian Rockies is enigmatic because of the time (~60 Ma) that has elapsed since the compressional deformation that produced the foreland thrust and fold belt and related crustal thickening beneath the southern Canadian Rockies. The Early Jurassic disappearance of the crust and the lower mantle lithosphere of the Slide Mountain basin is enigmatic because it accompanied the northeast-verging obduction of Slide Mountain terrane over the Cordilleran miogeocline, but occurred behind the Late Triassic-Early Jurassic Nicola magmatic arc and above the related southwest-verging subduction zone that consumed the Cache Creek Ocean. These two puzzling but disparate aspects of the tectonic evolution of the southern Canadian Cordillera may both be explicable in terms of the same basic pattern of retrograde southwestward flow which has involved the mantle wedge separating the subducting slab of oceanic lithosphere from the overlying Cordilleran lithosphere, and which has been entrained above the northeastward flow of the oceanic lithosphere that has been subducting beneath the western margin of North America.

The high topographic relief of the southern Canadian Rockies and adjacent parts of the southeastern Canadian Cordillera is associated with anomalously high upper mantle electrical conductivity and heat flow and with anomalously shallow mantle asthenosphere beneath the south-central Canadian Cordillera. The shallow asthenosphere is anomalous because of the time (> 40 Ma) that has elapsed since the east-west crustal stretching that occurred in this region. Gough (1986) attributed the shallow asthenosphere to symmetrical mantle upwelling beneath the south-central Canadian Cordillera; however, Clowes et al. (1995), in a recent lithospheric-scale section synthesizing seismic refraction and reflection data from the southern Canadian Cordillera, provided a different kinematic interpretation. They ascribed it to a pattern of asymmetric mantle flow that rises under the eastern Cordillera and descends under the western Cordillera where it is entrained above the subducting Juan de Fuca plate. Their revised kinematic interpretation, which involves retrograde mantle flow beneath the southern Canadian Cordillera, may offer a simple explanation for both the enigmatic contemporary topographic relief of the southeastern Canadian Cordillera and the enigmatic disappearance of the Slide Mountain basin. The upward mantle flow may provide dynamic support for the uplift of the southern Canadian Rockies. The horizontal flow sustains the shallow asthenosphere and high heat flow beneath the south-central Canadian Cordillera. The retrograde southwestward downward flow may provide an explanation for the Early Jurassic disappearance of the crust and lower mantle of the Slide Mountain basin, and for the northeastward obduction of Slide Mountain terrane over the Cordilleran miogeocline.

The collapse of the Slide Mountain basin and obduction of Slide Mountain terrane occurred as North America began rifting away from the Africa. It involved convergence between the North American miogeocline and the northeast dipping subduction zone along which the Cache Creek ocean was consumed beneath the Nicola magmatic arc (Monger and Price, 2000).

Slide Mountain terrane consists mainly of oceanic supracrustal rocks. The underlying oceanic crust and mantle lithosphere that formed the basin in which Slide Mountain terrane formed have disappeared. The convergence between North America and the northeast dipping subduction zone above the Nicola magmatic arc was expressed as the northeastward obduction of Slide Mountain terrane over the Cordilleran miogeocline. It was accompanied by the disappearance of the oceanic crust and mantle lithosphere of the Slide Mountain basin. This evidently was accomplished by southwestward retrograde downward flow that was entrained above the subducting oceanic lithosphere of the Cache Creek ocean as it sank into the mantle under the Nicola magmatic arc (Price and Monger, 2000). Following the disappearance of the Slide Mountain basin and the continuing convergence between North America and the Late Triassic Nicola magmatic arc, the Cache Creek subduction zone flattened and the Early Jurassic magmatic arc migrated northeastward into the rocks of the Cordilleran miogeocline.

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Upper Paleozoic stratigraphic linkage between Kootenay Arc and “Vernon Antiform”: New constraints on southern Canadian Cordilleran evolution

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A distinctive diopside-bearing calcareous quartzite-calcsilicate marker unit (CQM) of late Paleozoic age occurs in the Poplar Creek area of the Kootenay Arc, within the northern Valhalla Complex, along the eastern margin of Upper Arrow Lake, in both hangingwall and footwall successions of the Columbia River Fault, and as a nearly continuous belt from Upper Arrow Lake to the Pinnacles; from there, CQM can be mapped through the Sugar Lake-Mabel Lake-Trinity Valley regions to Vernon, north from Vernon to Mara Lake and south from Vernon to near Kelowna. The Chase quartzite, which can be mapped north and west of Vernon to the town of Chase, is a diopside-bearing calcareous quartzite like CQM.

In the Poplar Creek area, CQM is part of the Upper Paleozoic Milford Group which unconformably overlies the Lower Paleozoic (?) Lardeau Group, specifically the Broadview Formation.

At East Arrow Park, in the “Narrows” separating Upper from Lower Arrow lakes, CQM overlies paragneiss that bears a close resemblance to both the Vaseaux gneiss of the Penticton area and the “layered biotite-quartz-feldspar” gneiss of the southern Monashee Complex; and is overlain by metapelitic rocks, marble, amphibolitic schist, minor amphibolite, chlorite-biotite schist, and metaconglomerate,

all of which correlate with parts of the Milford and/or Kaslo groups. Here, it appears that Upper Paleozoic metamorphosed strata overlie much older (~ 2 Ga) “Monashee basement” with no Upper Proterozoic or Lower Paleozoic section preserved.

The presence of CQM immediately beneath the trace of the Columbia River Fault adjacent Arrow Park Lake, and immediately above the fault on the eastern shore of Upper Arrow Lake near Halcyon Hot Springs, constrains both the geometry and magnitude of displacement on the fault – if a fault exists.

In the Pinnacles, Sugar Lake and Mabel Lake areas, CQM overlies a succession of biotite-hornblende schist, pelitic (often calcareous) quartzite, marble and occasional amphibolite; these rock types are typical of the Milford Group. There is no compelling reason to suggest they are part of an older succession. CQM is overlain by muscovite-biotite schist, pelitic quartzite, calcareous schist, marble and, in places, amphibolitic schist belonging to the Milford and Kaslo groups.

The Upper Paleozoic succession is truncated, at a low angle, by a distinctive, carbonaceous succession of pyritic, black, biotite-quartzite, siliceous “argillite”, chlorite-biotite schist, and sooty phyllite containing thin, discontinuous units of volcanic and

volcaniclastic rocks; scant microfossil collections support a Late Triassic age. This carbonaceous succession correlates with the Slocan and Nicola groups

South of Vernon, within the “Vernon Antiform”, CQM overlies a thin (< 1 km) biotite-hornblende schist and calcsilicate-marble succession that in turn overlies orthogneiss typical of the “veined hornblende-biotite granodiorite gneiss” of the Monashee Complex. We suggest that Upper Paleozoic rocks are either in direct contact with Monashee orthogneiss or within at most a “stratigraphic kilometre” of it. Either way, little or no Lower Paleozoic or Upper Proterozoic section is present and the remaining few kilometres of Upper Paleozoic and Mesozoic section constitute a thin metasedimentary veneer atop a much older (~ 2 Ga) Monashee basement that cores the “Vernon Antiform”.

CQM varies in thickness from a few meters to about 500 m and maintains a remarkably consistent lithological character throughout the region. On weathered surfaces, its rough, sugary texture and etched banding brought about by the differential weathering of calcite surrounding quartz “grains” is distinctive; on fresh surfaces, it is uniformly pale grey to white and homogeneous. In most exposures, feldspar (dominantly orthoclase) is an important constituent and diopside a common accessory.

CQM is a time-stratigraphic marker that permits other, important stratigraphic markers to be placed in context. For example, the “Spa Creek assemblage” of amphibolitic schist, metaconglomerate and marble, which can be mapped on either side of the Okanagan Valley, overlies CQM, making it part of the Milford/Kaslo succession and correlative with quartzite-pebble metaconglomerate (including

granite clasts in some exposures) mapped between Vernon and Upper Arrow Lake. The age of the conglomerate at “Bob Lake” another 40 km to the west, is late Paleozoic, assuming correlation between it and the Spa Creek assemblage. The likelihood that CQM is correlative with the “Chase quartzite” helps constrain the age of the Tsalkom, Sicamous and Eagle Bay formations in the vicinity of Sicamous and Shuswap Lakes to late Paleozoic. As well, much of the muscovite-biotite-garnet schist correlated with the Silver Creek Formation may belong to the Upper Paleozoic Milford and Kaslo groups.

The presence of a stratigraphic thread spanning more than 150 Km of the southern Cordillera, from the Kootenay Arc across the Shuswap Metamorphic Complex and into the Intermontane belt provides an important new constraint for paleogeographic reconstruction. For example, CQM makes it possible to tie, stratigraphically, the Eocambrian Hamill Group to Upper Paleozoic rocks interpreted to be part of the “Quesnel terrane”, lending support to the notion that the terrane is stratigraphically tied to the ancient Pacific margin instead of having been emplaced as a tectonic flake sometime during the Mesozoic.

Early Eocene extension-related fluids in the Monashee Complex, southeastern BC.

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Oxygen and hydrogen isotope studies in the Monashee Complex indicate that meteoric fluids have caused heavy isotope depletion, and amphibolites from Thor-Odin (the southern dome of the Monashee Complex) show that there is a correlation between depletion and the $^{40}\text{Ar}/^{39}\text{Ar}$ ages of hornblende. The amphibolites have $^{40}\text{Ar}/^{39}\text{Ar}$ ages varying between ~54 and ~87 Ma. Amphibolites with ~54 Ma $^{40}\text{Ar}/^{39}\text{Ar}$ plateau ages suffered the most ^{18}O and D depletion. Older amphibolites, with near-plateau ages or disturbed spectra are less depleted in ^{18}O and D. There are two possibilities: 1) the presence of meteoric water at ~54 Ma $^{40}\text{Ar}/^{39}\text{Ar}$ ages in Thor-Odin resulted in complete resetting of the $^{40}\text{Ar}/^{39}\text{Ar}$ ages, and 2) the ~54 Ma ages could represent normal cooling ages, and older age spectra and disturbed spectra would be the result of excess ^{40}Ar ; the correlation between the heavy isotope depletions and young argon ages would then be fortuitous rather than causal. The possibility that meteoric waters could cause Ar-loss and resetting of $^{40}\text{Ar}/^{39}\text{Ar}$ ages in hornblende in amphibolites will be discussed.

Many monazites in the Monashee Complex give ages younger than ~54 Ma, the time of extension and retrograde metamorphism. Irrespective of which of the above interpretations of the ~54 Ma age we accept, the monazites must have grown below the closure temperature of argon in hornblende (530°C). Monazites are known to grow in the presence of fluids at low temperatures and it is suggested therefore that the ~54-50 Ma ages of monazites in Thor-Odin, do not represent peak metamorphic conditions, but represent growth during extension and retrograde metamorphism, in the presence of meteoric water. The argon and monazite work is ongoing, but so far it seems to have major implications for interpretation of ages of deformation in the Monashee Complex, and for interpretation of $^{40}\text{Ar}/^{39}\text{Ar}$ -hornblende and U-Pb monazite ages in general.

Structural data and interpretations in the northeastern Thor-Odin culmination, Monashee complex, BC.

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Detailed, 1:10000 scale mapping in the Mt. Begbie region revealed new data which must be integrated into any tectonic model of the Monashee complex. Two observations are presented. These are an episode of north-south directed extension, and a complex history of lineation over-printing. In addition, large scale folding is reinterpreted as an interference structure.

In the field area the presence of south dipping normal faults, boudinage, and local south plunging lineation suggests that the complex underwent extension in a north-south direction. The extension direction recorded in this area is perpendicular to the well documented, regional east-west extensional event.

Lineation cannot be attributed specifically to one generation of deformation in most outcrops, unless there is a clear relationship to other identifiable structures. For example, in some F_3 folds, the early L_1/L_2 lineation is folded around the hinge. In other F_3 folds the L_1/L_2 lineation is obliterated and replaced by an L_3 , axis parallel, crenulation lineation. Some F_3 folds exhibit both lineations. Without the benefit of folds or other structures in the outcrop, there is often no way to distinguish which lineation is present. Post F_3 extension subsequently produces a south trending lineation in some areas. This complex history of lineation formation makes map-scale kinematic interpretation, based on lineation, difficult.

The geometry of large-scale folding in the northeastern Thor-Odin culmination is compatible with multiple episodes of refolding. The Begbie anticline is a large F_2 reclined fold. This F_2 structure is refolded by west-plunging F_3 's (Mulvehill syncline, Tilley anticline). Finally a north-south trending, F_4 open warp affected the culmination.

Characteristics of Proterozoic basement orthogneiss and (late?) mafic intrusions: clues to an early history and the role of basement in Cordilleran orogenesis.

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The basement gneiss of the Monashee complex is an extensive unit of Proterozoic migmatitic ortho- and paragneiss. Though many studies in the Thor-Odin area of the complex have included the basement gneiss, relatively little work has been done to investigate the structural geometry of the basement and document its role in Cordilleran orogenesis. Recent field work may provide possible constraints.

Large areas of Proterozoic orthogneiss previously mapped as hornblende granodiorite, outcrop in the deeper structural levels of the field area. At the outcrop scale there exist several rock types including hornblende granodiorite gneiss, a biotite granitic gneiss, a biotite-augen gneiss and a hornblende leucogranite among other minor rock types. All these are deformed by several generations of structures including isoclinal folding which appears to deform a least two generations of migmatite. Observations of the basement orthogneiss suggest that they have a unique history not recorded in other basement rocks and may represent a separate Proterozoic igneous complex.

At the area around Frigg glacier dominant fabrics of the orthogneiss are cut by late, north striking mafic dykes. The dykes are boudinaged and host a foliation which is kinked. The dykes have a unique character in the basement of Thor-Odin but are possibly correlated with late lamprophyre dykes elsewhere in the complex or represent an earlier intrusion possibly of pre-Cordilleran age. Either of these possibilities offers valuable insight into the role of Proterozoic basement in Cordilleran orogenesis.

The Kinbasket High, Lower Cambrian syn-rift arkose and quartzite in the western Rocky Mountains and eastern Purcell Mountains, and Lower Paleozoic extension in the Southern Cordillera

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In the northern termination of the Purcell Mountains (Blackwater Ridge), Lower Cambrian Hamill Group strata comprise shallow marine subarkose and quartzite, deposited above turbiditic fans of the Neoproterozoic Windermere Supergroup (Horsethief Creek Group). The uppermost member of the Upper Hamill Group comprises bioturbated, cross bedded shallow marine quartzite, and is lithostratigraphically equivalent to the Poker member of the Upper McNaughton Formation of the Gog Group, in the Solitude Range of the western Rocky Mountains. The Kinbasket High comprises a Late Proterozoic-Early Cambrian topographic high, where the Poker member rests unconformably on the grit division of the Horsethief Creek Group (Middle Miette Group equivalent). Directly to the south, the Hamill Group thickens abruptly, and in one location this thickening is related to a mapped fault interpreted to be of Early Cambrian age.

On the southwest flank of the Kinbasket High, Upper Hamill grey weathering wacke and Lower Hamill Group arkose pass directly beneath the Poker member, in a wedge shape, and thicken dramatically. Southward, new units of pebbly quartzite and feldspathic quartzite appear in the Upper Hamill Group. As the Hamill Group thickens to the southwest of the Kinbasket High, it lies on progressively younger, higher units in the Horsethief Creek Group, suggesting significant uplift and beveling of these Proterozoic strata toward the Kinbasket High. Where the Hamill Group is thickest, between the Kinbasket High and Dogtooth High, it rests with apparent conformity on a 'transitional' upper clastic division of the Horsethief Creek Group. From the drastic thickening of the Hamill Group, and beveling of the Horsethief Creek Group, it is interpreted that the Hamill Group was deposited during extensional faulting. While Upper Hamill units thicken across the mapped 'growth' fault, Lower Hamill Group units do not thicken appreciably, suggesting that extension in this basin was episodic, and protracted (Figure 1).

The Lower Hamill Group on Blackwater Ridge is lithostratigraphically equivalent to the Jasper member of the McNaughton Formation (Gog Group), in the Solitude Range. This lithostratigraphic package occurs at the base of the Gog Group and Hamill Group, at the latitude

of the study area and further north. The Jasper member comprises shallow marine sub-arkose and quartzite, commonly occurring in planar tabular cross-bedded sets. The Kinbasket High, although more subdued, is found in the Blackwater Range and Solitude Range of the Rocky Mountains, suggesting that it was a rather broad, possibly north-northwest trending, topographic high.

The Kinbasket High occurs in a zone of interconnected, discrete early Cambrian Highs occurring in the Western Rocky Mountains and Eastern Omineca Belt. The sub-arkose of the Jasper member suggests it was derived from a continental, 'basement' source, and was shed from these uplifted blocks, and deposited in fault controlled basins. Upper Hamill units are more mature, pure quartzites. These early Cambrian Highs record part of a protracted extensional history, which began in Late Proterozoic, and probably continued until Devonian time. Difficulties in precise lithostratigraphic correlation of Lower Paleozoic strata from the western Foreland Belt into the Omineca Belt are exacerbated by these topographic highs, since strata were deposited in separate, but interconnected, fault bounded basins.

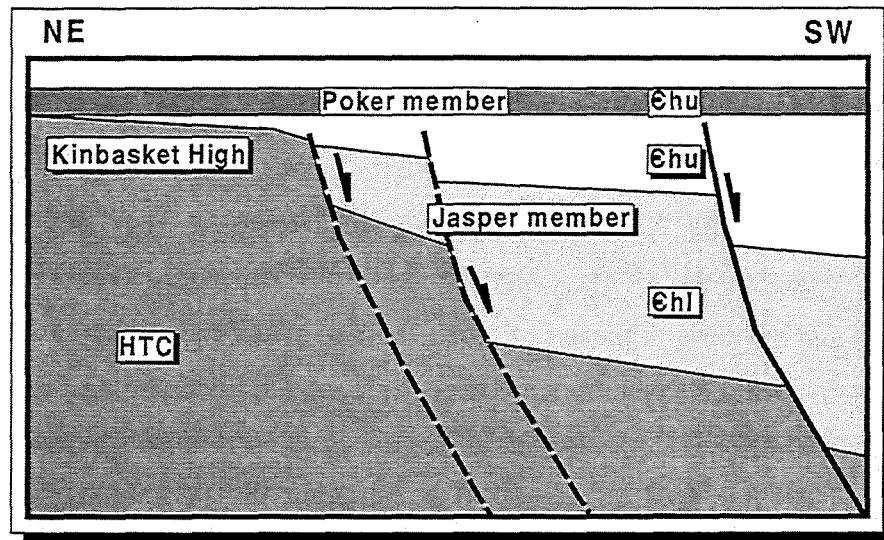


Figure 1. Interpreted geometry of the Hamill Group coming off of the southwest flank of the Kinbasket High.

Re-evaluation of Kinematic Models for the Selkirk Fan Structure in Light of New U-Pb Timing Constraints, Northern Selkirk Mountains, Southeastern British Columbia

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In the northern Selkirk Mountains situated within the hinterland of the Canadian Cordillera there is a km-scale structural fan, that is part of an en echelon series of structural fans that trend NW-SE parallel to the strike of the Cordilleran orogen from Idaho to Alaska. Southwest verging second generation folds (F2) with shallow dipping axial surfaces (S2) dominate the west flank of the fan, and become progressively steeper toward the axis. To the east are northeast verging, third generation folds (F3) with axial surfaces that become progressively overturned eastward toward the Rocky Mountain Trench. This divergence defines the geometry of the present day fan. However, the fan may have been initiated in F2 time; F2 structures east of the current fan axis may have been overturned toward the east prior to superposition of F3. Peak sillimanite-grade rocks core the fan structure, and are flanked on either side by progressively lower grade rocks. A more complete understanding of the kinematic evolution of the fan structure is fundamental to elucidating hinterland tectonics in the Cordillera.

Geochronologic data constrain timing of metamorphism and deformation, and demonstrate that the thermotectonic history of the region was protracted and complex. Data provided by monazite and zircon from pelitic schists, pre- to syn- and post-tectonic pegmatite, leucogranitic and granodioritic dykes, and plutons indicate that the structural and metamorphic development of the west flank of the fan occurred primarily in the Middle Jurassic. East of the fan axis deformation and metamorphism is more complex and there is evidence of both Jurassic and Cretaceous events.

On the west flank, development of the dominant F2 folds and S2 transposition foliation has been constrained to ca. 161-170 Ma. Based on structural analyses at hand sample and thin section scales, the peak metamorphic assemblage is interpreted to have developed primarily pre- to syn-F2, and is therefore also Middle Jurassic. This is corroborated by previous research that has shown that this part of the orogen was exhumed to upper crustal levels by the end of the Jurassic.

On the east flank, the development of F2 and the accompanying S2 transposition foliation is interpreted to be Middle Jurassic, whereas overprinting F3 folds which control outcrop and map-scale geometries have been constrained to no older than ca. 100 Ma and as young as 81 Ma. Within this Late Cretaceous timeframe, older F3 folds appear to be situated near the fan axis and younger F3 folds occur to the east. There were at least two metamorphic events that affected the east flank: a Middle Jurassic event which was primarily pre- to syn-D2, and a sillimanite-bearing syn- to post D3 Cretaceous thermal pulse.

These new data, together with results from previous studies in adjoining regions, support the following scenarios: 1) The Selkirk Fan Structure began in the Middle Jurassic when southwest verging folds and thrusts were developed in conjunction with peak metamorphism. 2) The superposition of F3 occurred in the Late Cretaceous coeval with a metamorphic event caused by either an influx of hydrothermal fluids related to Cretaceous plutonism or tectonic burial. We would argue that the available data are compatible with a diachronous structural fan where southwest verging F2 development in a prowedge during the Middle Jurassic was followed by superposition of F3 in the retrowedge. It is possible that a proto-F2 fan developed in the Middle Jurassic whose retrowedge was significantly overprinted by Cretaceous northeast verging F3 folds related to the decoupling of the fan above a singularity (i.e. subduction zone).

MID-CRUSTAL CHANNEL FLOW IN LARGE HOT OROGENS: RESULTS FROM COUPLED THERMAL-MECHANICAL MODELS

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ABSTRACT: Coupled thermal-mechanical models are used to investigate interactions among metamorphism, deformation, and exhumation in large convergent orogens, and the implications of coupling and feedbacks between these processes for observed structural and metamorphic styles. The models involve subduction of sub-orogenic mantle lithosphere, large amounts of convergence (up to 2400 km) at 2 and 5 cm/y, and a slope-dependent denudation rate. The model crust is layered with respect to thermal and rheological properties - the upper crust (0-25 km) has a wet quartzite flow law, with heat production of $2.0 \mu\text{W}/\text{m}^3$ from 0-20 km, and the lower crust (25-35 km) has a dry diabase flow law, with heat production of $0.75 \mu\text{W}/\text{m}^3$ from 20-35 km. After ca. 45 My of convergence at 2 cm/y, the model orogens develop a plateau underlain by crust that is about 75 km thick, with middle to lower crustal temperatures above 700°C. In most models, an extra increment of temperature-dependent "melt weakening" is introduced so that the effective viscosity decreases to a minimum of 10^{19} Pa.s at 750°C in material derived from the weak upper crustal layer. As a result, by ca. 30 My, a thin zone of outward channel flow develops at the base of the weak upper crustal layer where $T \geq 700^\circ\text{C}$. The channel flow zone is characterised by a reversal in velocity direction on the pro-side of the system, and is driven by a combination of a horizontal pressure gradient and the development of the temperature-dependent low viscosity horizon in the mid-crust. Exhumation resulting from surface erosion that is focused on the pro-side flank of the plateau leads to "ductile extrusion" of the channel flow zone. The exhumed channel displays normal-sense offset at its upper boundary, reverse-sense offset at its lower boundary, and an "inverted" metamorphic sequence across the ductile extrusion zone. Possible natural examples of the channel flow zones produced in these models may be found in the Himalayan-Tibetan system, and parts of the Grenville orogen and Canadian Cordillera.

INTRODUCTION

Consideration of observations and recent model results has led us to a conceptual prototype of the way that orogens may evolve in response to thermal-mechanical coupling (Fig. 1). Most of our earlier model experiments focused on the Type 1 and Type 2 stages. In particular, in Type 2 orogens the central core region becomes progressively hotter and weaker and is constrained by the 'vise jaws' of the colder and stronger exterior crust. Ellis et al. (1998) investigated aspects of the behaviour of 'vise' models using purely mechanical models in which a pre-defined weak zone was deformed between two converging stronger blocks. This is a progress report on fully coupled thermal-mechanical models in which the weak internal zone develops during the evolution of the model as a result of crustal thickening and heating (Type 3, Fig. 1; e.g., Vanderhaeghe et al. 1999; Jamieson et al. in revision). To the extent that the exterior regions remain strong, these models have a 'vise-like' character; however, the thermal and mechanical interaction between the orogen core and margins is more complex than in the simple mechanical models investigated previously. In particular, these models develop orogenic plateaus (e.g., Vanderhaeghe et al. 1999; Medvedev et al. 2000) and mid-crustal channel flows (this report; Jamieson et al. in revision), which we consider to be characteristic of Type 3 orogens.

Models of flow in low viscosity zones bounded by stronger material indicate that channel flow may be an important process in subduction zones (Shreve & Cloos 1986; Mancktelow 1995), crustal-scale shear zones (Grujic et al. 1996), and beneath plateau regions (eg. Bird 1991; Royden 1996; Royden et al. 1997; Clark & Royden 2000). Jamieson et al. (in revision) recently showed that a temperature-dependent decrease in mid-crustal viscosity, for example as the result of incipient partial melting, can produce narrow zones of outward-directed channel flow. If the channel flow zone is exhumed by focused erosion the resulting structure resembles a crustal-scale shear zone with reverse-sense kinematics at the base, normal-sense kinematics at the top, and an intervening zone of metamorphic rocks with an "inverted" profile. This report summarises our more recent investigations of the controls on, and consequences of, mid-crustal channel flow, using coupled thermal-mechanical models with a high resolution crustal grid. The results are compatible with structural and metamorphic features of the Greater Himalayan Series and Main Central Thrust zone of the Himalayas and the eastward migration of the Tibetan plateau (e.g. Royden et al. 1997; Clark & Royden 2000; Kirby et al. 2000), and may have some applications to parts of the Grenville and Canadian Cordillera orogens.

THERMAL-MECHANICAL MODEL

In the numerical models, the pro-mantle lithosphere detaches and subducts beneath the stationary retro-mantle (Fig. 2; Willett et al. 1993). The imposed basal velocity distribution therefore varies from the convergence velocity (V_p) to zero across the detachment point, S. A variation on this basal boundary condition involving advance of the S-point is presented below (HTRL-74). The mechanical model calculates the resulting crustal deformation assuming 2D vertical plane strain. The numerical technique uses the Arbitrary Lagrangian-Eulerian (ALE) formulation in which flows with free surfaces and large deformation are computed on an evolving Eulerian finite element grid that conforms to the material domain (Fullsack 1995). An associated Lagrangian grid, which moves according to the Eulerian velocities, and a tracking algorithm allow the associated Lagrangian motion and fields to be followed and transferred to the Eulerian finite element grid (Fullsack 1995). The imposed kinematic condition on the mantle lithosphere limits the coupled dynamic thermal-mechanical investigation to the crust, but thermal-kinematic calculations for the mantle are necessary to create the correct subduction-related thermal conditions at the base of the crust. The flexural isostatic compensation resulting from changes in crustal thickness is calculated from the elastic flexure of a beam embedded in the model at the base of the crust.

In the models considered here, denudation rate is calculated from the product of the local surface slope of the model, a time function, $f(t)$, and a spatial climate function, $g(x)$ (Fig. 2). This approach assumes that denudation is primarily controlled by bedrock incision by rivers and that the incision rate is proportional to the slope-discharge product. The time function in the present models involves an increase from zero to a constant value when denudation is turned on. In other examples (not described here) it serves to vary the discharge intensity on geological timescales. The climate function is a simple proxy for discharge in which precipitation is assumed to be orographic and focused on the pro-side of the orogen. On the retro-side of the orogen and on the plateau, $g(x) = 0$, representing an extremely arid climate. Across the pro-side of the orogen, $g(x)$ increases linearly to 1 and maintains this value over the undeformed pro-crust. This erosion function gives high denudation rates on the steep slopes on the pro-flank of the plateau and low to negligible rates elsewhere. In the present models, the model orogen first grows without erosion, in order to develop thick crust and an associated plateau relatively quickly. Denudation is then turned on (at $\Delta x = 750$ km), and the relative and combined effects of erosion at the

plateau edge and channel flow beneath the plateau are investigated. Results from models in which surface erosion operates throughout the development of the orogen are presented elsewhere (Jamieson et al. in revision).

The finite element calculations (Fullsack 1995) use a visco-plastic formulation of the relationship between strain rate ($d\epsilon/dt$) and stress (σ). The model crust consists of two layers with contrasting rheological properties (Fig. 2, 3). In the frictional field (at Coulomb yield at low temperatures and/or high stresses), both the upper and lower crust deform by incompressible frictional plastic flow with an internal angle of friction $\phi = 15^\circ$. In the ductile field (viscous flow at stresses below plastic yield), the rheology of the upper crust (0-25 km) follows the "wet Black Hills quartzite" flow law of Gleason & Tullis (1995), with the "A" value reported in that paper converted to the corresponding B^* value ($B^*(WQz)$, Table 1) using conditions appropriate for cylindrical creep tests. Sensitivity tests indicate that models in which the rheology of the upper crust is defined by $B^*(WQz) \times 5$ produce crustal thicknesses of 70-75 km. This scaling was adopted because it is also within the error estimates of the laboratory measurements (Gleason & Tullis 1995) and the uncertainties associated with crustal composition. For the lower crust (25-35 km), the "dry Maryland diabase" flow law of Mackwell et al. (1998), with the corresponding conversion from "A" to $B^*(DMD)$ (Table 1) is used. This rheological structure can be considered to represent a (meta)sedimentary and/or granitic upper crust underlain by a mafic granulite lower crust.

In most of the models presented here, an additional component of temperature-dependent weakening is introduced in upper crustal material. As shown diagrammatically in Fig. 3 (bold box and lower panel), when the ambient temperature exceeds a threshold value of 700°C , and below an upper transition temperature of 750°C , the value of the model effective viscosity varies linearly between that defined by the flow law ($WQz \times 5$) at $T = 700^\circ\text{C}$ (η_{700}) and 10^{19} Pa.s at $T \geq 750^\circ\text{C}$ (η_{750}). Models with no viscosity reduction (HTRL-29) and with a lower minimum viscosity (HTRL-26) are described below. This temperature-dependent weakening represents a simple serial mixing model between two end-member components - the reference flow law ($WQz \times 5$) and a weak, linear viscous component with viscosity η_{750} . A simple interpretation of this viscosity model is that *in situ* partial melting of the crust reduces the bulk effective viscosity to η_{750} at $T \geq 750^\circ\text{C}$. Although η_{750} is much larger than the viscosity of real crustal melts, it could correspond to the effective viscosity of crust containing a small fraction of *in situ* melt. The

natural value of η_{750} is not known; we therefore determined the sensitivity of model results to a range of values of η_{750} and present results with the maximum value ($\eta_{750} = 10^{19}$ Pa.s) that produces significant channel flows in the models. Lower values of η_{750} may result in more focused, vigorous flows. For convenience, we refer to this temperature-dependent viscosity reduction as "melt weakening"; however, the model would apply to any corresponding process that reduces effective viscosity at high temperature. We emphasize that the present models are not designed to investigate the effects of segregation, migration, and emplacement of magma.

The thermal calculations include heat production, conduction, and advection (Fig. 2). Radioactive heat-producing material is present throughout the model crust, with higher heat production in the upper crust ($A_1 = 2.0 \mu\text{W}/\text{m}^3$ from 0-20 km) than in the lower crust ($A_2 = 0.75 \mu\text{W}/\text{m}^3$ from 20-35 km) (Fig. 3). This differs from our previous work (e.g., Jamieson et al. 1998; Vanderhaeghe et al. 1999) in which there was no lower crustal heat production. However, although lower crustal heat production is probably low, it is unlikely to be zero, especially in tectonically active regions (e.g., Rudnick & Fountain 1995; Russell & Kopylova 1999). The crustal heat production values used in these models can be taken to represent crust that is not strongly fractionated. We recognize that this probably represents the end-member case for orogens in which heat is generated almost entirely from thickened radioactive crust, with no additional heat input from increased basal heat flux, magmatic underplating, lithospheric thinning, or strain heating. Addition of heat from any of these sources would allow corresponding reductions in A_1 , A_2 , or an increase in K , compared with the values used here (Table 1).

For each model run, an initial temperature field is calculated at the scale of the lithosphere, assuming an initial one-dimensional, vertical, conductive steady-state ($V_p = 0$) that reflects conduction of heat from the mantle and heat production within the radioactive layers. The lithosphere-asthenosphere boundary is defined at a depth such that the basal temperature, T_a , is 1350°C , the basal mantle heat flux (q_m) is $20 \text{ mW}/\text{m}^2$, and the surface temperature is 0°C . For these boundary conditions, the surface heat flux (q_s) is $70 \text{ mW}/\text{m}^2$ before crustal thickening. Because there is no lateral variation of thermal properties, isotherms are initially horizontal, and for $K = 2.0 \text{ W}/\text{m}^\circ\text{K}$ the Moho temperature is everywhere 704°C (Fig. 2). The present models contrast with some of our previous work (e.g., Jamieson et al. 1998), which assumed a precursor phase of subduction in order to investigate the early stages of orogenic evolution, and thus

involved laterally variable initial thermal conditions across the model crust and mantle. However, for the large amounts of convergence considered here ($t = 45\text{-}75$ My, $\Delta x = 1200\text{-}2400$ km), the initial thermal conditions have little effect on the final crustal thermal structure (Vanderhaeghe et al. 1999); the effect of precursor subduction has therefore been ignored in the present study.

REFERENCE MODEL: HTRL-25

In order to facilitate comparison among models, the configuration and results of a reference model with "standard" properties are described first (HTRL-25; Figs. 2,4,5,6,7; Table 1). Subsequent models are developed from this reference model by systematically varying a few important parameters (Fig. 8). Model layering and corresponding rheological and thermal parameters are as described above and in Figures 2 and 3. For this model $V_p = 2$ cm/y and the S-point is fixed. As described above, denudation rate is proportional to surface slope and varies with both distance and time (Fig. 2).

Figure 4 shows how model HTRL-25 evolves between 7.5 and 22.5 My; the early evolution of other models is similar and is not presented. Figure 4a shows the total velocity field and deformed passive marker grid for the crust; Figure 4b shows the distribution of crustal radioactivity and corresponding thermal structure. In the region of thickened crust, contrasts in the relative magnitudes of upper and lower crustal velocities across the S-point reflect the degree of coupling between upper crust, lower crust, and upper mantle (upper panels of each pair, Fig. 4a). On the pro-side, high lower crustal velocities reflect nearly total coupling to the subducting pro-mantle lithosphere; on the retro-side, lower crustal velocities are close to zero, reflecting similar coupling to the stationary retro-mantle lithosphere. By 22.5 My, a mid-crustal velocity minimum is evident on the pro-side of the S-point. The deformed marker grid (lower panels, Fig. 4a) shows an asymmetric antiform developed from lower crust on the retro-side of the S-point. As the model orogen grows, the antiformal structure develops into a bivergent tectonic wedge where lower crust is detached at S. The retro-limb of this structure becomes increasingly recumbent and the structure as a whole becomes increasingly asymmetric with time. However, this wedge is nested within the larger-scale model orogen, and lower crustal deformation is generally not evident near the surface of the model. The geological significance of this feature remains unclear and it is not discussed in any detail below.

The thermal structure of the model orogen (lower panels, Fig. 4b) is closely linked to the distribution of crustal radioactivity (upper panels, Fig. 4b). Isotherms are initially elevated in the region of thickened upper crust and depressed in the vicinity of the subducting mantle lithosphere. By 22.5 My radioactive heating dominates, so that much of the middle and lower crust and Moho temperature are significantly above 700°C. The velocity minimum seen at this stage (Fig. 4a) occurs where the temperature of the weak upper crust is 600-700°C, and therefore reflects weakening according to the wet quartzite flow law (Fig. 3) rather than the influence of additional "melt weakening".

Figure 5 shows the mechanical and thermal evolution of model HTRL-25 from 30-60 My. At 30 My, a velocity reversal is evident at the base of the upper crustal mechanical layer (Fig. 5a). In this region, the mechanically weak layer has reached 700-750°C, corresponding to the temperature range where the viscosity is reduced by "melt weakening" (Fig. 3). Although crustal thickness approaches 80 km and Moho temperature exceeds 800°C on the retro-side of the system, a true plateau (e.g., Medvedev et al. 2000) has not yet developed. At 37.5 My ($\Delta x = 750$ km), slope-dependent erosion begins at the pro-ward edge of the thickened region according to the model described above (Fig. 2). By 45 My, the mid-crustal velocity reversal has developed into a well-defined zone of outward flow, about 15 km thick, lying close to the base of the weak upper crustal layer. The erosion front and the tip of the channel flow zone are well separated (decoupled) at this stage. A plateau has developed, and much of the mid- to lower crust is above 800°C. By 52.5 My, the channel flow zone has nearly reached the edge of the plateau region, and coupling has been initiated with the exhumed crust beneath the erosion front. Mid- to lower crustal isotherms are displaced pro-ward in response to the combined effects of advection and radioactive self-heating of hot material within the channel (Fig. 5b). The strong positive feedback between heating above 700°C, "melt-weakening", and the onset of channel flow means that the channel tip is near the 700°C isotherm; the 700°C and higher isotherms are inverted by the reversal in flow sense between the channel and the lower crust. The strong curvature of isotherms produces a near-isobaric temperature increase of 600-900°C from the edge of the plateau towards the S-point at the level of the channel flow zone.

By 60 My (Fig. 5, 6), focused erosion at the plateau edge has transported channel material to the surface within the exhumation zone. Convergent pro-crust is exposed in the footwall beneath the channel material and a cross-section of plateau-edge upper crust is exposed in the

hangingwall. The stippled regions in Figure 6 show the approximate channel position and the arrows (Fig. 6a) show the sense of distributed shear across the channel boundaries. The lower boundary is a reverse-sense ductile shear zone and the upper boundary is normal-sense. Isotherms (Fig. 5b,6b,6c) are advected outward and upward by the flow beneath the erosion front as the channel is exhumed (Fig. 6b), forming a thermal boundary layer in which 700°C crust lies only about 15 km beneath the surface. Above this zone, near-surface isotherms are strongly condensed but not inverted. The maximum exhumation rate in this region is ca. 1 cm/y. Throughout the evolution of this model, the "suture" between pro- and retro-crust (vertical marker 0, corresponding to the crustal column immediately above the S-point at the onset of convergence; Fig. 2, 5a) is transported progressively retro-ward; by 60 My its surface position lies more than 300 km retro-ward of the S-point.

Figure 7 shows the contributions to the total velocity field at 60 My by the subduction boundary condition (V_p) and by gravitational forcing. The velocity field attributable to the basal boundary condition is consistently retro-ward, although it is effectively zero in the hottest part of the channel flow zone. On the pro-side of the system, gravitational flow is consistently pro-ward and is highest in the channel flow zone, reaching a maximum at the outer edge of the channel beneath the denudation front. The decomposition of velocities provides insight into the relative nature of the tectonic and gravitational flows and demonstrates that gravitational forces drive the channel flow.

In the following sections, we explore the sensitivity of the model to the effects of denudation, "melt weakening", and convergence velocity. Specific changes in standard model parameters are summarized in Figure 8.

EFFECT OF NO EROSION: HTRL-30

Model HTRL-30 is identical to HTRL-25 except that there is no surface erosion at any stage. Reverse channel flow initiates at about 30 My. The tip of the channel flow zone migrates pro-ward between 45 and 75 My (Fig. 9a), but the rate of outward migration of the tip and velocities within the channel are both less than observed for the equivalent times in HTRL-25. Crustal thickness reaches a maximum of about 75 km by 45 My and a plateau has developed by 52.5 My. After this time, the model orogen grows outward in both directions (pro-side only shown in Fig. 9) at approximately constant thickness in response to a combination of influx of

material at the edge of the orogen and outward flow within the channels. Temperatures within the orogen are similar to those seen in HTRL-25 (Fig. 9b), although isotherms are not displaced pro-ward to the same extent. Much of the mid- to lower crust has reached temperatures of 700-950°C by 60 My, and by 75 My lower crustal temperature exceeds 1000°C in the vicinity of the S-point, where some radioactive upper crustal material has been incorporated into the lower crust. By 75 My, the suture has been transported more than 350 km retro-ward of the S-point. Model HTRL-30 demonstrates the form of channel flows that are injected laterally into the crust, as opposed to those that are extruded to the surface, as in HTRL-25.

EFFECT OF NO "MELT WEAKENING": HTRL-29

Model HTRL-29 (Fig. 10) is identical to HTRL-25 except that there is no additional "melt weakening"; the rheology of the upper crust is controlled entirely by the wet quartzite flow law ($B^*(WQz)^{x5}$; Table 1; Fig. 3). A reversal in the mid-crustal velocity field is not observed until ca. 52.5 My (Fig. 10a). By comparison with HTRL-25, reverse velocities are small, the channel is relatively thick, and the channel tip migrates pro-ward very slowly between 52.5 and 67.5 My. After 67.5 My, a significant change in velocity field results in reverse flow of the entire upper crust on the pro-side of the S-point. The upper crust is no longer driven retro-ward by the basal forces (cf. Fig. 7). Instead, removal of strong upper crust at the erosion front has allowed the upper crust to detach and flow outward under gravitational forces. The outward flow extends and thins the crust above the lower crustal antiform, which grows upward in response to the reduced upper crustal load. The exhumation zone at the plateau edge is broader than for HTRL 25. The distribution of radioactive upper crust and corresponding crustal thermal structure (Fig. 10b) are similar to HTRL-25, although mid-crustal isotherms are not displaced pro-ward, nor condensed beneath the erosion front, to the same extent. In this model, the suture lies more than 500 km retro-ward of the S-point by 75 My.

EFFECT OF LOWER "MELT WEAKENED" VISCOSITY: HTRL-26

Model HTRL-26 is identical to HTRL-25, except that the minimum "melt weakened" viscosity, η_{750} , is reduced to 2×10^{18} Pa.s at 750°C. Crustal velocities, deformation, and thermal fields evolve in a very similar manner to those for HTRL-25, and are not shown. Figure 11 shows the effect of the five-fold reduction in η_{750} on the various components of the crustal velocity field at 60 My. Despite the lower viscosities, the gravitationally-driven pro-ward

velocity field in the channel is very similar to that observed in HTRL-25 (Fig. 7, Fig. 11). This result indicates that the velocity in the channel does not increase directly with decreasing viscosity, as predicted by one-dimensional analyses of channel flow (e.g., Turcotte & Schubert 1982). Instead, the rate of outward flow is strongly controlled by the strength of the upper crust at the erosion front, which partially blocks the flow.

EFFECT OF POST-CONVERGENT GRAVITATIONAL FLOW: HTRL-32

Model HTRL-32 is identical to model HTRL-30 (no erosion) except that convergence stops at 52.5 My ($V_p = 0$; Fig. 12). At this time, the model crust is ca. 75 km thick (Fig. 12a), and the suture lies about 300 km retro-ward of the S-point. Much of the middle and lower crust beneath the plateau has reached 700-900°C (Fig. 12b), and Moho temperature beneath the central part of the plateau is 800-900°C. Immediately following the cessation of convergence at 52.5 My, the model orogen flows outward in both directions in response to the gravitational potential energy gradient, resulting in crustal thinning in the centre of the model orogen (<60 km by 67.5 My) that is balanced by thickening at the edges. The suture is transported further retro-ward during outward flow of the retro-crust. The rate of outward flow, which is highest within the mid-crustal channel on the pro-side of the system, declines rapidly between 52.5 and 60 My. Outward flow within the pro-crust is much more significant than in the retro-crust, where flow velocity decreases after 52.5 My (Fig. 12a). This has potentially important implications because it predicts that post-orogenic "gravitational collapse" in orogens driven by mantle subduction will be less significant in the retro- and transverse directions by comparison with gravity-driven flow in the pro-crust, where basal traction during convergence is strongest.

Although crustal thickness decreases, the lower crust continues to heat up because the cooling effects of subduction and influx of cool material from the pro-side have been removed. Crustal isotherms are displaced pro-ward by the outward flow, and temperatures also increase owing to radioactive self-heating at the thickened outer edges of the plateau. At 67.5 My, most of the middle and lower crust is still above 700°C, locally exceeding 1000°C; Moho temperature exceeds 900°C beneath much of the model orogen.

EFFECT OF ADVANCING SUBDUCTION: HTRL-74

All of the thermal-mechanical model results presented above and in our previous work (e.g., Jamieson et al. 1996, 1998, in revision) have assumed that the position of S-point is fixed.

However, in real orogens, the point at which sub-orogenic lithosphere is detached may migrate either away from or towards the converging pro-lithosphere, leading to subduction zone advance or subduction zone retreat respectively (e.g., Royden 1993; Willett & Beaumont 1994; Waschbusch & Beaumont 1996). Advance of the S-point (V_s same sign as V_p) implies that convergence of the pro-lithosphere is partitioned between subduction zone advance (at V_s) and subduction (at $V_p - V_s$). In the limit where $V_s = V_p$, pro-lithosphere indents the retro-lithosphere at V_p and no lithosphere is subducted. Subduction zone retreat (V_s opposite sign to V_p) has the opposite effect and increases the subduction velocity ($V_p - (-V_s)$).

Model HTRL-74 was designed to investigate the effect of subduction zone advance on the thermal and mechanical evolution and channel flow within a large orogen, using velocities appropriate for Cenozoic convergence of India relative to Asia. In this model $V_p = 5$ cm/y, $V_s = 2.5$ cm/y, and subduction velocity = 2.5 cm/y, relative to the fixed retro-limit of the model where the retro-crust is stationary ($V_r = 0$; Fig. 13a). The results are shown in a reference frame in which the S-point is located at 0 km at 48 My (Fig. 13). During the evolution of the model, S moves 1200 km retro-ward at 2.5 cm/y, while the pro-mantle lithosphere subducts beneath S at 2.5 cm/y and converges 2400 km on the undeformed exterior retro-lithosphere. Because of the high overall convergence velocity ($V_p = 5$ cm/y), equivalent amounts of convergence (Δx) are achieved in shorter times and the orogen grows faster. In this case, erosion begins at 15 My ($\Delta x = 750$ km), when the orogen has achieved the same amount of convergence as in models with $V_p = 2$ cm/y.

In this model, thickening of the retro-crust largely results from convergence driven by the motion of S at $V_s = 2.5$ cm/y. At 24 My, a mid-crustal velocity minimum has developed on the pro-side of the system, where crustal temperatures are 700-800°C (Fig. 13); there is also a corresponding outward flow in the retro-crust. By 36 My, the velocity minimum has developed into a narrow zone of reverse channel flow, localized where mid-crustal temperatures exceed 800°C. Although the channel flow may appear less significant than in HTRL-25 (Fig. 5a), the velocity scales are different and the relative velocities are large; parts of the channel are flowing pro-ward at rates that exceed the 5 cm/y convergence velocity. A plateau has formed, the crust is 70-75 km thick, and the middle crust is being exhumed in a narrow zone at the erosion front. As in the other models, the lower crustal antiform grows upward to replace material lost by outward channel flow, but unlike the other models, the pro-side of this structure becomes increasingly

recumbent and appears to be displaced pro-ward because the S-point and lower crust are transported retro-ward beneath it.

In marked contrast to all the other models, the suture initially remains close to the advancing S-point (0-36 My). Its surface expression then moves relatively pro-ward so that by 48 My it is 200 km pro-ward of S. These relative motions can be explained by the evolving velocity distributions (Fig. 13a). Early in the model evolution, the velocity of the upper crust near S is similar to V_s , so that the suture position (marker 0, Fig. 13a) is still near S at 24 My. Following the onset of channel flow and erosion, the velocity of the upper crust in the pro-plateau progressively decreases to ca. $V_s/2$ at 36 My. Under these conditions, S must advance retro-ward with respect to the suture at the same rate (ca. 12.5 km/My). This relative velocity continues to increase as the velocity of the upper crust in the plateau decreases (Fig. 13a). This produces ca. 200 km separation by 48 My, and this distance continues to increase with time. In effect, the advancing S-point underthrusts the suture and the lower crustal antiform. These motions and their implications for the relative positions of S, the suture, and the erosion front are important in the context of the Himalayan-Tibetan system discussed below.

In HTRL-74, the distance between the suture and the erosion front remains nearly constant (Fig. 13a, 24-48 My). There is little deformation of the pro-plateau crust during this interval, and the plateau edge above the exhumed channel undergoes little denudational retreat (marker 3, Fig. 13a). In contrast, a large section of upper pro-crust is first deformed and then removed by denudation at the erosion front (markers 5-9, Fig. 13a). Although its upper crustal geometry is nearly symmetric (Fig. 13a, 36-48 My), the erosion front preferentially removes pro-crust as material moves through the front with velocities shown in Fig. 13a. Pro-crust that enters the orogen early in its evolution is preserved in the plateau pro-ward of the suture, whereas upper pro-crust that enters the orogen later is removed by erosion. As a result, crustal markers that were originally separated by 1200 km (3 and 9, Fig. 13a) are juxtaposed across the erosion front at 48 My.

In terms of the thermal evolution, high lower crustal temperatures are reached somewhat earlier than for model HTRL-25 because the upper radioactive crust thickens faster. However, for a given amount of convergence, HTRL-74 is cooler than HTRL-25 (Fig. 5b, 13b), because the rate of influx of cool material is higher relative to the rates of heat production and diffusion. By 24 My the middle and lower crust have reached 700-850°C, with Moho temperature above

800°C on the retro-side of the advancing S-point (Fig. 13b). The onset of channel flow, marked by the velocity minimum at 24 My (Fig. 13a), is accompanied by the pro-ward advection of mid-crustal isotherms. By 36 My temperature exceeds 900°C in the lower crust near the S-point, most of the lower pro-crust has reached 700-900°C, and the 700°C isotherm lies less than 15 km beneath the erosion front. Except for pro-ward entrainment of the 900°C isotherm, thermal conditions remain similar between 36 and 48 My.

Model HTRL-74 demonstrates that even when the overall convergence rate is high and the tectonic conditions include subduction zone advance, the tendency for hot, weak pro-crust to flow outward in mid-crustal channels persists. In fact, the conditions at the erosion front at 36-48 My (Fig. 13) are quite similar to those in HTRL-25 at 60 My (Fig. 5; note difference in velocity scales). This result implies that coupling between channel flows and erosion focused on the pro-side of an orogenic plateau may be relatively insensitive to the tectonic conditions elsewhere within and beneath the orogenic plateau. In addition, the model demonstrates that there is a significant difference in the final large-scale geometries of models with and without subduction zone advance; these differences may be present, and possibly detectable, in natural orogens.

DISCUSSION

These results offer some insight and raise some questions concerning the controls and consequences of mid-crustal channel flows in large, hot orogens. In the following discussion, we highlight some problems that require further investigation. In particular, we note the role of positive feedbacks between the thermal and mechanical evolution of the models, surface denudation, and the styles of channel flow. Comparisons with work on channel flows by others (e.g., Bird 1991; Royden 1996; Royden et al. 1997) and with our previous "large hot orogen" models are presented elsewhere (Jamieson et al. in revision). No in-depth discussion of the physics of channel flow is attempted here.

Feedbacks between Thermal-Mechanical Evolution and Channel Flow

In the present models, channel flow can develop in the absence of surface denudation (HTRL-30, Fig. 9a). In this case, the channel propagates by horizontal injection beneath overlying cooler crust, with the outward migration of the channel tip closely linked to viscosity reduction at the 700-750°C melt-weakening front (Fig. 9). This implies that channel flow will propagate inefficiently in the absence of "melt weakening" - i.e., where $T < 700^{\circ}\text{C}$ or at higher

temperature where crustal composition inhibits melting. This is shown by model HTRL-29 (Fig.10, 60 My) in which channel flow is sluggish even at $T > 750^{\circ}\text{C}$. It follows that in the models described here, outward propagation of the channel tip is sensitively coupled to mechanical thickening of the crust, and the ensuing thermal relaxation and radioactive self-heating responsible for heating the mid-crust to $T > 700^{\circ}\text{C}$. Once this threshold temperature has been reached, a strong positive feedback is established whereby injection of channel material advects hot material, thickens the crust, and increases the temperature by radioactive self-heating of the newly thickened crust. It is likely that changes of surface topography and accompanying outward growth of the plateau are more closely related to this thermal-mechanical coupling at the tip of the channel near its threshold temperature than to the properties of the channel material in its hottest, lowest viscosity region.

If correct, this interpretation implies that these models are not equivalent to ones in which there is a pre-existing channel. For example, the behaviour contrasts with the simplified model of ductile flow in the lower crust (Clark & Royden 2000), which assumes a pre-existing uniform channel thickness (h) and a uniform viscosity material (η) flowing through the channel under pressure gradients derived from the topographic gradient. This model predicts that topography diffuses in response to channel flow with a diffusivity that scales with h^3/η . However, in the models discussed here, the variations in channel thickness and viscosity near the channel tip are large and no channel exists beyond the periphery of the plateau (e.g., HTRL-30, Fig. 9a). We suggest that an analytical model approximately corresponding to our numerical results could be developed using a topographic and crustal thickening diffusion model (e.g., Clark & Royden 2000) in which h^3/η is a function of temperature calculated from thermal relaxation within the thickening crust.

Feedbacks between Surface Denudation and Channel Flow

Although channel flows may develop in the absence of surface denudation (HTRL-30), the limited range of model experiments reported here indicates that channel flows are strongly enhanced by surface denudation when flow and exhumation become coupled (HTRL-25, Fig. 5a). Surface denudation modulates the "openness" of the channel conduit by removing upper crustal material; our model results suggest that this exerts a fundamental control on the style and velocity of channel flow. In this context, comparison with "salt fountains" is instructive. Open-conduit salt fountains are known to extrude salt at rates of 2-3 cm/y (Talbot et al. 2000) if the salt

is removed efficiently by dissolution or flow over the surrounding landscape. Salt intrudes much less efficiently (<1 mm/y) where an overburden "plug" partially blocks the conduit and impedes the flow. In the present models the "plug" is the cool upper crust, and the efficiency of surface denudation should affect the rate of extrusion of the hot material in the channel. By analogy with salt fountains, the potential for gneiss or migmatite "fountains" or protrusions in orogenic regions with unusually efficient surface denudation should be considered. We speculate that the Nanga Parbat massif may represent a present-day gneiss protrusion, the surface expression of a low viscosity channel that is being rapidly extruded in response to very high rates of erosion (e.g. Zeitler et al. 2001).

A closely associated problem concerns the determinants of channel velocity. As demonstrated by comparison of models HTRL-25 (Fig. 7) and HTRL-26 (Fig. 11), a decrease in η_{750} does not cause a proportional increase in the channel flow velocity, as would be predicted from analysis of one-dimensional, constant-viscosity channels (e.g., Turcotte & Schubert 1982, sections 6.2, 6.3). Instead, the efficiency of flow is determined by the openness of the conduit. In this case, it is limited by the thermal and mechanical properties of the cool, strong surface boundary layer - channel material cannot spurt out as it would from an open, efficient channel. For comparison, we define a uniform, open-ended channel, with the same geometry as the model channel, and a uniform viscosity equal to that in the highest temperature part of the channel ($T \geq 750^\circ\text{C}$, $\eta_{750} = 10^{19}$ Pa.s). The efficiency of channel flow, which is the ratio of the measured flux to the flux through the equivalent uniform channel, is determined by the viscous/plastic properties of any material that tends to block the outlet. In the absence of denudation the conduit is closed, and the channel propagates by horizontal injection as described above.

Denudation ensures that channel flow is efficient when the thermal-mechanical coupling allows rapid intrusion of the channel material into the near-surface thermal-mechanical boundary layer. Eventually, the channel material is extruded where the surface thermal boundary layer is strongly thinned by advection and surface denudation, as in HTRL-25 (Fig. 5, 60 My). The thinned boundary layer in this case has no significant frictional-plastic upper region, and the ductile, power-law viscous region above the melt-weakening front is only 15 km thick. A near steady-state condition is achieved among uplift of the surface, slope-dependent denudation, and upward flow of channel material into the thinned boundary layer. Sensitivity experiments in which erosion rate is varied with time demonstrate that, below a threshold denudation rate, the

channel propagation reverts to the horizontal injection mode and the channel thickens to form a dome below the region of reduced erosion (e.g., Jamieson et al. in revision; Fig. 6, 45 My). The crustal geometry and structural style of the exhumed channel at the denudation front can range from the gneiss dome style, through a near-symmetric "decapitated" protrusion (HTRL-25, Fig. 5a, 60 My), to an asymmetric spout aligned with thrust-sense structures (not shown here). In the absence of denudation, there is no tendency for upward, dome-like intrusion of the channel material in any of the models shown here, even though the vertical pressure gradient is much larger than the horizontal equivalent.

Implications for Topography, Plateau Growth, and Denudation Fronts

Figure 14 illustrates conceptually how the outward propagation of a channel containing low viscosity material leads to outward growth of a plateau with a steep topographic front. Given the right climatic conditions, this can become a denudation front. In the early stages of convergence when the crust is cold and basal traction is high (Fig. 14a), the model orogens develop back-to-back tectonic wedges with low but relatively uniform surface slopes across the pro- and retro-wedges (e.g., Fig. 4a). As the base heats up and decouples, the wedge taper decreases and the interior of the model orogen becomes a plateau. Development of the weak channel further reduces the internal strength of the orogen. In the present models, a steep topographic front forms between the edge of plateau and the foreland owing to the strong contrast in crustal strength between the hot, weak channel tip and the cool, strong foreland. When there is no denudation, the topographic front will propagate outward at the intrinsic horizontal injection rate of the channel tip (e.g., HTRL-30, Fig. 9). When the topographic front is denuded, the associated coupled rapid propagation of the channel accelerates the outward migration of the plateau edge and steepens the topographic front, which in turn produces increased denudation rates. This type of feedback loop could lead to a steady-state plateau margin that is denuded at the rate of outward growth. An intrinsic part of the denudation front would be the structurally lower extruded and eroded channel spout.

Flow Modes

The styles of flow observed in the model experiments, simplified into one dimension, are summarized in Figure 15. These styles can be characterized by upper and lower crustal velocities and the channel velocity. Modes 1 and 2, which occur in our earlier models (e.g. Beaumont et al. 1994), represent cases where the upper and lower crust are strongly or

weakly coupled. Mode 3 is the reverse channel flow mode seen in models like HTRL-25 (Fig. 5, 60 My); the upper and lower crust move in the same direction in response to the basal boundary conditions, separated by the channel in which the sense of flow is reversed or greatly reduced. Mode 4, "gravity sliding" of the upper crust, occurs when the crust above the channel becomes decoupled from the basal boundary condition. This can result from erosion of the plateau edge (e.g., HTRL-29), and leads to outward translation of the upper crust at the same rate as the channel flow. Mode 5 represents an intermediate style between modes 3 and 4.

The relative shear sense at the upper and lower boundaries of the channel in modes 3, 4, and 5 is relevant to comparisons with natural orogens (below). In both modes 3 and 5, the shear sense reverses between the upper and lower channel boundaries, and the shear sense across a given boundary is the same in the two modes. However, the direction of motion of the upper crust between the two cases is opposite, illustrating that kinematic shear sense alone is not diagnostic of the direction of motion. In Mode 4 there is no reversal of shear sense because the upper crustal velocity matches the maximum channel velocity. Presumably there is another flow mode in which the upper crust moves outward even faster than the middle crust, but this has not been observed for parameter range used in these models.

Implications for Large-Scale Orogenic Thermal and Transport Regimes

Although it is well recognized that motion on faults and shear zones at high thermal Peclet number leads to significant advection of temperature fields and thermal disequilibrium within orogens (e.g., Molnar & England 1990; Ruppel & Hodges 1994; Jamieson et al. 1996; Huerta et al. 1999), the same processes acting at the lithospheric scale are under-appreciated. For example, inferences are quite commonly made concerning thermal conditions at depth based on one-dimensional, steady-state vertical thermal extrapolations of upper crustal temperature fields and heat fluxes. However, orogens are by their nature places where deformation and transport of material at high thermal Peclet number can be expected. Moreover, if the results of the present experiments apply even approximately to large natural orogens, opposite-sense sub-horizontal flows may be widespread, leading to further thermal disequilibrium in the orogenic crust and sub-orogenic upper mantle. We illustrate the pitfalls of simple one-dimensional extrapolations using examples from HTRL-25.

At the largest of scales, the subducting pro-lithosphere can remain cool, as in oceanic subduction zones, even when it is advected beneath hot crust. For example, there is little change

in the temperature of the pro-mantle as it is carried beneath the plateau in HTRL-25 (Fig. 5b, 60 My; Fig. 6b). Thermal gradients in the lower crust and upper mantle can be low or inverted where advection of cool material into or beneath an orogen limits lower crustal and upper mantle temperatures. At the crustal scale, temperatures may be significantly increased by advection in channel flow zones (Fig. 5b, 60 My; Fig. 6b). Simple extrapolation of upper crustal temperatures (700°C at ca. 25 km depth beneath the plateau of HTRL-25) would imply near-melting conditions within the uppermost mantle, in contrast to the 800-900°C seen in the model (Fig. 5b). Equivalently, evidence of high temperature metamorphism at low pressures should be used with care when inferring contemporary conditions deeper in the lithosphere. The models demonstrate that high surface heat flux and high upper to middle crustal temperatures do not always imply mantle melting, and show that simple vertical extrapolations of upper crustal thermal conditions could produce meaningless estimates of deep thermal structure.

A related problem at a smaller scale concerns estimation of exhumation rates using cooling histories derived from thermochronological data. In the absence of independent depth control, cooling rates are commonly converted to exhumation rates by assuming a "typical" geothermal gradient. However, crustal isotherms are likely to be strongly condensed in the vicinity of an active erosion front (e.g., Batt & Braun 1997; Jamieson et al. 1998; HTRL-25, Fig. 5b, Fig. 6). The resulting near-surface geotherm may be highly "atypical" and is also likely to vary with time, rendering exhumation rates calculated from an assumed constant geotherm meaningless. Finally, it is clear that lateral tectonic transport is far more important than vertical transport in these models (see also Jamieson et al. 1996, 1998; Batt & Braun 1997; Huerta et al. 1999). A simple one-dimensional view of material paths in orogens (burial followed by exhumation) is therefore likely to be inaccurate, and may lead to incorrect interpretations of orogenic transport and provenance. The same one-dimensional approach, which does not account for transport of material through lateral thermal gradients, can lead to problems in interpreting metamorphic P-T-t paths. A better understanding of transport trajectories during orogenesis can lead to a better appreciation of the tectonic juxtaposition of material of very different provenance, dispersal of material from the same initial location, and the significance of metamorphic P-T-t data.

APPLICATIONS TO NATURAL OROGENS

Himalayan-Tibetan system

A number of studies have suggested that various forms of channel flow have been important during the evolution of the Himalayan-Tibetan system (e.g., Grujic et al. 1996; Royden et al. 1997; Clark & Royden 2000). In this section we explore the similarities and differences between some of the models presented here and various parts of this large orogenic system.

Ductile extrusion associated with channel flow has been proposed to explain the relationship between thrust-sense and normal-sense ductile fabrics and associated metamorphism and exhumation within and above the Main Central Thrust (MCT) zone in the Bhutan Himalayas (Grujic et al. 1996, 2000). Figure 16 shows the geometry of model HTRL-25 at 60 My in the vicinity of the erosion front, compared to simplified crustal-scale geometries for the Himalayas and southern Tibet (Davidson et al. 1997; Wu et al. 1998). Deformed and metamorphosed Indian upper crust of the Greater Himalayan sequence (also referred to as the High Himalayan Crystallines or Greater Himalayan Belt), is thrust over lower grade Indian crust along the MCT zone, and is separated from the overlying weakly deformed and metamorphosed Tethyan sequence by the South Tibetan Detachment system (STD) (e.g., Hodges 2000 and references therein). A laterally extensive suite of ca. 20-25 Ma leucogranites is concentrated immediately beneath the STD. Geochronological data indicate that top-to-the south (thrust-sense) motion within the MCT zone and top-to-the-south (normal-sense) motion along the STD system were broadly coeval (Hodges 2000, and references therein). At depth, seismic data suggest that Indian mantle lithosphere continues northward beneath the Tibetan plateau for a considerable distance. In the vicinity of 92-93°E, the surface expression of the Indus-Tsangpo suture lies ca. 200 km north of the STD, and ca. 400 km south of the point where Indian lithosphere is inferred to be detached and subducted into the mantle (e.g., Owens & Zandt 1997), although new data suggest that the latter distance may be subject to revision (Tilman & Ni 2000).

In general, the geometries of models HTRL-25 at 60 My and HTRL-74 at 36 My in the vicinity of the erosion front resemble the upper crustal geometry inferred for the MCT, Greater Himalayan sequence, and STD (Fig. 16). The lower surface of the extruded channel flow zone shows reverse-sense shear, while the upper surface shows normal-sense shear. The extruded region between the bounding shear zones is a relatively narrow zone of strongly deformed pro-crustal material that has experienced high temperatures. In the model, the extruded region

continues at depth into the "melt weakened" region where channel flow is active; extrusion of the channel material has occurred in response to focused erosion. Similar geometries and exhumation processes have been inferred for the Greater Himalayan sequence in various parts of the Himalayas (e.g., Searle & Rex 1989; Grujic et al. 1996, 2000; Searle et al. 1997; Wu et al. 1997; Hodges 2000).

However, in detail there are some significant differences between the geometries of model HTRL-25 and HTRL-74 and present-day geometries in the Himalayan-Tibetan system. In the models, the exhumed region is roughly symmetrical, with both underlying and overlying material bowed up adjacent to the steeply dipping extrusion zone (Fig. 16). This is in contrast to the strongly asymmetric, moderately dipping structures observed in the present-day MCT zone and adjacent regions. We note that the model most appropriately represents the geometry of the extrusion zone at a time of rapid exhumation, which in the case of the MCT zone was most likely at 15-20 Ma (e.g., Najman et al. 2000; White et al. in press). Work is in progress to investigate the effect of a change in erosion rate in the models on the geometry and cooling history of the exhumed region. Focused erosion has been proposed to account for present-day rapid exhumation of the Nanga Parbat massif in the western Himalayan syntaxis (e.g., Zeitler et al. 2001, and references therein). Although the three-dimensional geometry of the syntaxis contrasts with the two-dimensional models used here, near-surface structure in the Nanga-Parbat region is more consistent with model HTRL-25 than is the geometry of the present-day MCT zone.

In model HTRL-25, the surface position of the suture (vertical marker "0") lies several hundred kilometres retro-ward of the S-point. In the context of the Himalayan-Tibetan system, assuming that India (pro-side, on right) moves northward relative to a "fixed" Asia (retro-side, on left), model HTRL-25 predicts that the suture should lie north of the mantle detachment point, whereas it actually lies several hundred kilometres to the south. However, in model HTRL-74, which incorporates subduction zone advance, the relative positions of the suture and mantle detachment point (Fig. 13a, 48 My) are consistent with observations from the Himalayan-Tibetan system. For comparison of model HTRL-74 with the Himalayan-Tibetan orogen, in a reference frame fixed to the undeformed part of Asia, the panels of Figure 13 should be moved relative to one another by the offset in their S positions. When this is done, a marker in the undeformed pro-crust moves retro-ward by 2400 km in 48 My, approximating the northward motion of undeformed India since the collision. In the models, partitioning of V_p into equal components of

subduction and subduction zone advance is merely illustrative. A more general interpretation is that convergence of India on Asia at ca. 5 cm/y is partitioned between subduction of Indian mantle lithosphere at < 5 cm/y and northward advance of the detachment point at a finite rate. From an indenter perspective, the model can be interpreted to imply that the Indian mantle lithosphere has indented Asia by 1200 km while 1200 km of Indian mantle lithosphere has been subducted or underthrust at the advancing limit of the indenter.

Figures 17 and 18 compare metamorphism predicted by model HTRL-25 (HTRL-74 is similar) with observations from parts of the Himalayan-Tibetan system. Xenoliths in mafic volcanics from the central Tibetan plateau contain granulite facies assemblages that have yielded P-T estimates of 700-1100°C at 8-14 kbar (Hacker et al. 2000); very young ages from the same samples imply that the middle to lower crust beneath the Tibetan plateau is currently at granulite facies conditions. P-T-t paths and peak metamorphic P-T conditions for some points in model HTRL-25 (Fig. 17) overlap with the P-T range reported by Hacker et al. (2000), demonstrating that temperatures and pressures within the model orogen are compatible with the observations.

Particle paths (Fig. 17) show that the tracked points, which originated on the pro-side of the system, were transported hundreds of kilometres retro-ward, incorporated into the orogen, entrained in the channel flow zone, and finally exhumed by the combination of outward channel flow and surface denudation. This lateral transport is a characteristic feature of the models and can lead to juxtaposition of originally widely separated crustal sections across the erosion front (e.g., HTRL-25, HTRL-74; Figs. 5a, 13a). This may provide an alternative explanation for observed contrasts in pre-Himalayan zircon provenance and inferred geological history over short distances across the MCT zone (e.g., DeCelles et al. 2000), which have previously been attributed to pre-Tertiary movement on the MCT and related structures.

A characteristic feature of the Himalayan orogen is the upward increase in metamorphic grade across the MCT zone, generally referred to as "inverted" metamorphism (e.g., LeFort 1975). The origin of the inverted metamorphic sequence is very controversial (e.g., Hodges 2000, and references therein), and there is also considerable debate about the reliability of various P-T estimates from this zone (e.g., Macfarlane 1995; Fraser et al. 2000). Regardless of the merits of the various arguments on the origin of, and P-T variations within, the inverted metamorphic sequence, a viable model for the Himalayan-Tibetan system should produce a similar pattern of metamorphism at the model surface.

Figure 18a presents peak grade profiles for model HTRL-25 in the vicinity of the erosion front. These profiles show the maximum temperature, and corresponding pressures and times, reached by points that emerge at the surface of the model at a specified time, as a function of distance across the model surface. The profiles can be thought of as metamorphic traverses across the model surface. The results for HTRL-25 show that peak temperature increases structurally upward across the extruded channel zone (i.e., the profile is "inverted") and decreases again towards the plateau region. The peak temperature profile becomes narrower and moderately asymmetric with time. Peak temperatures are achieved latest in the central part of the extruded zone and earliest on the flanks. The peak pressure profile is distinctly less symmetric than the peak temperature profile, with a steep pressure increase on the pro-ward side. Figure 18b shows the peak temperature and pressure profiles for HTRL-25 at 67.5 My compared with observed P-T variations from the MCT zone and overlying Greater Himalayan sequence in the Langtang region of central Nepal (Inger & Harris 1992; Macfarlane 1995; Fraser et al. 2000). The model profiles agree well with the P-T data, given the uncertainties associated with the P-T estimates and discrepancies between studies. It is unfortunate that there are few reliable P-T data from the Lesser Himalayan sequence below the MCT zone, where the upward increase in metamorphic grade should be most noticeable.

The models show that ductile extrusion associated with channel flow can produce P-T variations at the model surface that are compatible with the observed variations in metamorphic grade. Our earlier work using a different style of tectonic model (Jamieson et al. 1996) also produced inverted metamorphic profiles consistent with observations from the MCT zone. However, that tectonic model predicted inverted metamorphism on the retro-side of the system and thrust-sense kinematics across the entire exhumed region, inconsistent with the larger scale regional tectonics of the Himalayan system. The inverted metamorphic pattern results from distributed syn-metamorphic ductile shear during exhumation (Grujic et al. 1996; Jamieson et al. 1996) and may be relatively insensitive to the details of the crustal-scale thermal-tectonic model used. Consistency not only with the metamorphism, but with other first-order features of the Himalayan-Tibetan system, should therefore be used as a test of tectonic models. In this respect, the present models are a significant improvement on our earlier work.

A number of studies have suggested that lower crustal channel flow accounts for present-day eastward migration of the edge of the Tibetan plateau (e.g., Royden 1996; Royden et al.

1997; Clark & Royden 2000; Kirby et al. 2000). We agree that mid-crustal channel flow in the absence of erosion, for example as shown in HTRL-30 (Fig. 9), could account for this phenomenon. In model HTRL-30, the channel propagates horizontally at depth at a rate controlled by the thermal and mechanical state of the region near the channel tip. As discussed above, the mechanism presented here differs from the model proposed by Clark & Royden (2000) in that lateral gradients in temperature, viscosity, and pressure control the channel flow. Although the two approaches differ, the conclusion is similar - outward flow of low viscosity mid-to lower crust could be responsible for the outward growth of the Tibetan plateau.

Grenville orogen

The Grenville orogen is widely regarded as a Proterozoic collisional orogen of Himalayan proportions. The structural and metamorphic style of the Central Gneiss Belt of the western Grenville orogen in Ontario suggest that the present-day erosion surface lay 25-30 km beneath an orogenic plateau at some time between ca. 1080-1020 Ma (Culshaw et al. 1997, and references therein). At the southwestern boundary of the Central Gneiss Belt, a shallow-dipping, relatively thin zone of highly migmatitic orthogneisses (Muskoka domain) separates less migmatitic overlying and underlying units with contrasting geological histories. Outcrop-scale features in the migmatites suggest that leucosome was present during southeast-directed (normal-sense) flow, and geochronological data suggest that the migmatites and associated extensional structures were broadly coeval with the ca. 1075-1060 Ma northwest-directed thrusting that is characteristic of the orogen as a whole (Timmermann 1998). Metamorphic P-T estimates from the Muskoka domain (CGB, Fig. 17; Timmermann 1998) are compatible with conditions beneath the model plateau regions. We speculate that extensive partial melting and associated viscosity reduction in the Muskoka domain gneisses led to southeast-directed outward channel flow within the convergent Grenville orogen. There is no evidence for rapid exhumation of the Muskoka domain, so it is likely that the postulated channel flow zone (if it existed) remained buried, as in model HTRL-30 (no erosion). In order to test this hypothesis, more work is needed on the field relationships and ages of migmatites and associated structures in the Muskoka domain, as well as on models using properties and geometries more appropriate to the Grenville orogen.

Younger, broadly coeval, thrust- and normal-sense shear zones are also known from the northern flank of the Grenville orogen. In particular, Rivers & Indares (1999) proposed that the ductile extrusion model of Grujic et al. (1996) might account for rapid exhumation of high-

pressure rocks in eastern Quebec and western Labrador. At the western end of the orogen, similar processes may have operated within the Grenville Front Tectonic Zone (e.g., Haggart et al. 1993), or within and beneath the ca. 1020 Ma extensional Shawanaga Shear Zone (e.g., Ketchum et al. 1998). We will test these ideas by investigating the effects of retro-side focused erosion on channel flow and exhumation in a large convergent orogen.

Southeastern Canadian Cordillera

Eocene leucogranite intrusions are closely associated with extensional shear zones in the southeastern Canadian Cordillera (Carr et al. 1987; Simony & Carr 1997). Metamorphism at deeper structural levels was diachronous, and associated with ductile thrusting that produced inverted metamorphic field gradients (e.g., Gibson et al. 1999). In the Valhalla region, less than 5 My separates thrusting at deeper structural levels from leucogranite intrusion and normal faulting at higher structural levels (Simony & Carr 1997). We speculate that channel flow may have affected the southeastern part of the Canadian Cordillera in the early Tertiary; we propose to test this hypothesis using model parameters appropriate for this region.

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FIGURE CAPTIONS:

Figure 1. Conceptual model of orogenesis that is controlled by thermal-mechanical coupling. (a) Type 1: Simple (cold, strong) orogens develop narrow, bivergent style when there are no major crustal heterogeneities and crust is strongly coupled to mantle. Tectonic style is controlled by the dynamics of the sub-orogenic lithosphere, corresponding here to mantle and lower crust subduction beneath S. (b) Type 2: Central core of the orogenic crust is weaker than exterior 'vise jaws'. Weak core may be intrinsically weak (e.g., sedimentary terranes), strain-softened, or hot. (c) Type 3: Large, hot orogen in which thick crust in the interior of the orogen is thermally weakened and decouples from the mantle to form a plateau. Gravitational forces control plateau development and internal channel flows, as discussed in this report. (d) Type 4: Large, hot orogens in which crust and mantle are decoupled and mantle lithosphere may be recycled by mechanisms other than subduction. For description of relative control of the dynamics by the Ampferer (Am) and Argand (Ar) force ratios, see Ellis et al. (1998).

Figure 2. Initial conditions for reference model HTRL-25 and most other models presented here (exceptions described in text). Upper panel shows undeformed, rectangular, passive marker grid; solid vertical lines at 200 km intervals are numbered (positive on pro-side, negative on retro-

side) for comparison with the deformed grids shown in later figures. Vertical marker "0" corresponds to "suture" between pro- and retro-crust at onset of deformation (see discussion in text). Upper and lower crustal layers are shown in different grey shades. Lower panel shows thermal structure of the model crust, with isotherms at 100°C intervals. Radioactive heat production in upper crust (0-20 km) is A_1 and in lower crust (20-35 km) is A_2 . Velocity vectors are shown as short heavy lines. Model is started in conductive steady-state, i.e., in thermal equilibrium with $V_p = 0$. Model boundary conditions: surface temperature = 0°C, basal temperature, $T_a = 1350^\circ\text{C}$; basal mantle heat flux, $q_m = 20 \text{ mW/m}^2$; no horizontal heat flux at the sides of the model. At the start of the model, pro-lithosphere converges with velocity $V_p = 2 \text{ cm/y}$, and detaches and is subducted beneath stationary retro-lithosphere; detachment point (S) is fixed in most models. Crustal thickening is isostatically compensated by two broken elastic beams (flexural rigidity 10^{22} Nm). Denudation rate function is proportional to slope (details in text) and corresponds to denudation rates of ca. 1 cm/y where slopes are 1:10 on the pro-side of the orogen. Slopes are calculated over a horizontal length scale of 10 km, the resolution of the finite element grid. The model is time-stepped by alternating mechanical and thermal steps. Thermal steps solve the time-dependent heat balance equation (shown above) with the current velocity field determined dynamically within the crust and by the boundary conditions. Radioactive upper and lower crust and the temperature field are advected and the updated temperature field calculated for the entire model region. This modifies the mechanics through the material flow laws for the upper ("wet quartzite x 5") and lower ("dry Maryland diabase") crust. Flow law parameters: η = effective viscosity (Pa.s), B^* = pre-exponential factor (Pa.s); $d\varepsilon/dt$ = strain rate (s^{-1}), n = stress exponent, Q^* = activation energy (kJ/mol), R = gas constant (8.314 J/°Kmol), T = °K. Heat balance equation parameters: ρ = density, C_p = heat capacity, K = thermal conductivity, $\kappa = K/\rho C_p$ = thermal diffusivity (all constant); $T(x,z,t)$ = temperature, t = time, $A(x,z,t)$ = rate of heat production / unit volume; $\underline{v}(x,z,t)$ = tectonic velocity (dynamic in crust, kinematic in mantle); $\nabla = \partial/\partial x + \partial/\partial z$ = gradient operator. Parameter values are given in Table 1 and changes between models are summarized in Fig. 8.

Figure 3. Thermal (left) and mechanical (right) properties of the reference model crust (Table 1, Fig. 2). A_1 , A_2 = radioactive heat production of upper (0-20 km) and lower (20-35 km) crust; q_m , q_s = basal and surface heat fluxes. Frictional-plastic Coulomb yield has an internal angle of

friction $\phi = 15^\circ$. Ductile rheology (viscous flow at stresses below plastic yield) follows the wet quartzite flow law (Gleason & Tullis 1995) with $B^* \times 5$ in the upper crust (WQz $\times 5$, 0-25 km; see text) and the dry Maryland diabase flow law (Mackwell et al. 1998) in the lower crust (25-35 km). Basal boundary condition is either subduction or advancing subduction (see Fig. 2, 8, and text). "Melt weakening" of upper crust (bold box and lower panel) reduces viscosity (stippled region) such that the material viscosity varies linearly from η_{700} to η_{750} from 700-750°C; for $T \geq 750^\circ\text{C}$, $\eta_{750} = 10^{19}$ Pa.s. This reduces the effective viscosity from the "Wet Quartzite" value to the "Newtonian Creep" value (lower panel).

Figure 4. Early (7.5-22.5 My) evolution of HTRL-25 with increasing convergence ($\Delta x = 150$ -450 km). During this interval there is no surface erosion. Vertical and horizontal scales are the same. (a) Velocity fields (upper panel of each pair) and deformed passive marker grids (lower panel of each pair), showing convergence (from right) driven by subduction of pro-mantle lithosphere at $V_p = 2$ cm/y. S is the detachment point; velocity of retro-mantle, $V_r = 0$. Velocity is in reference frame of fixed retro-limit of the model. Deformation of passive marker grid shows detachment of upper crust and formation of lower crustal antiform. Bold vertical marker lines are spaced 200 km apart before deformation. "Suture" (marker 0) is located directly above S at the start of the model; negative/ positive marker numbers are on retro-/pro-sides respectively. (b) Progressive deformation of upper and lower crustal radioactive layers (grey shades in upper panel of each pair; see Fig. 2 for initial conditions) and corresponding temperature distribution (isotherms in lower panel of each pair), contoured at 100°C intervals. q_s and q_m are the surface and basal heat fluxes in the undeformed crust.

Figure 5. Later (30-60 My) evolution of HTRL-25 with increasing convergence ($\Delta x = 600$ -1200 km). Panels as in Fig. 4. (a) Mid-crustal channel flow (velocity and deformation panels) increases with time and couples with crustal exhumation at the erosion front. Denudation model is described in Fig. 2 and text. Passive marker grid shows exhumation and extrusion of "melt weakened" channel material by 60 My in the near-symmetric "decapitated" protrusion at the erosion front. (b) Radioactive crust distribution and temperature fields show exhumation of lower crust at the erosion front by 60 My, and horizontal advection of 800°C and hotter isotherms. Upward advection compresses the temperature field into a thin, high-gradient, thermal

boundary layer beneath the erosion front, which would produce very rapid near-surface cooling in any material exhumed through this region.

Figure 6. Detail of HTRL-25 (60 My) showing relationship between the channel (stippled region) and erosion front. (a) Note opposite sense of shear across upper and lower boundaries of the channel, uplift of the footwall below the channel and uplift of the hangingwall above the channel. (b) Rapid exhumation of material from the channel (velocity vectors) has advected and condensed the near-surface thermal boundary layer to give high geothermal gradients below the erosion front. Velocities are shown with respect to stationary retro-lithosphere in (b) and with respect to stationary pro-lithosphere in (c).

Figure 7. Decomposition of the velocity field for HTRL-25 at 60 My. This example shows the total velocity (top panel) and the contributions to the total from flow driven by the basal subduction boundary condition alone (acceleration due to gravity = 0; middle panel) and by gravitational forces acting alone (basal velocity = 0; lower panel). Calculations are made by first iterating for the non-linear model properties (viscosity and plasticity) that satisfy the total boundary conditions and gravitational forcing, then solving the system using these model properties with either the boundary conditions or gravitational forcing set to zero. Note that the outward channel flow is driven by gravity. If convergence stops ($V_p = 0$), the velocity field would change from that shown in the top panel to that shown in the third panel (see model HTRL-32, Fig. 12a) and there would be outward flow of the upper crust in addition to the outward channel flow.

Figure 8. Flow chart showing properties of reference model HTRL-25 (upper box) and variations on these standard parameters used in the other model experiments reported here (lower boxes).

Figure 9. Evolution of HTRL-30 (45-75 My) with increasing convergence ($\Delta x = 900$ -1500 km). Panels as described for Fig. 4. Model parameters as for HTRL-25 except that there is no surface erosion. The outward migration of the channel tip is achieved by lateral injection into the mid-crust; thickening and heating of the crust to reach a threshold temperature of 700-750°C are

required before injection occurs. 800°C and higher isotherms are advected horizontally by the channel flow.

Figure 10. Evolution of HTRL-29 (52.5-75 My) with increasing convergence ($\Delta x = 1050$ -1500 km). See Fig. 4 for description of panels. Model experiment is the same as HTRL-25 except that there is no "melt weakening" of upper crustal material. Model develops a sluggish channel flow (52.5-60 My) followed by detachment and outward flow of the entire upper crust (75 My) when denudation has thinned the strong near-surface boundary layer below the erosion front.

Figure 11. Decomposition of the velocity field of HTRL-26 at 60 My. The model is the same as HTRL-25 except that the minimum viscosity achieved by "melt weakening" is 2×10^{18} Pa.s. As shown in Fig. 7, the velocity field is separated into components driven by the basal boundary conditions alone and gravitational forces alone. Note that the velocity of the channel flow is very similar to that of HTRL-25 (Fig.7) despite the five-fold reduction in minimum viscosity.

Figure 12. Evolution of HTRL-32 (52.5-67.5 My) following cessation of convergence at 52.5 My. See Fig. 4 for description of panels. Model parameters are the same as for HTRL-30 (no erosion) except that convergence stops ($V_p = 0$). Model shows outward flow of channel material and upper crust driven by gravitational forces alone (part a). Note that the channel does not propagate outwards until the 700-750°C threshold temperature is reached; outward flow substantially thins the plateau crust, while thickening the exterior crust. Channel flow velocities are initially high enough to invert temperatures in the lower crust but inversions dissipate as flow rates decrease with time (part b).

Figure 13. Evolution of HTRL-74 (24-48 My) with increasing convergence ($\Delta x = 1200$ -2400 km). See Fig. 4 for description of panels. In this model $V_p = 5$ cm/y (convergence of pro-lithosphere relative to stationary retro-lithosphere), partitioned into subduction zone advance ($V_s = 2.5$ cm/y), shown as the velocity of the S-point, and subduction at 2.5 cm/y. The high V_p means that the orogen grows more rapidly than in the other models; erosion begins at $\Delta x = 750$ km (15 My), i.e., at the same amount of convergence but an earlier time than in the other models. Results are shown in a reference frame attached to S, and the absolute position of S within the model is labelled in km. Velocities are shown with respect to stationary retro-lithosphere and the

vector scale is 5 cm/y. The suture is labelled "0". An outward channel flow coupled to the erosion front develops by 36 My. The main difference from the other models is the underthrusting beneath the lower crustal antiform and the motion of S relative to the suture, as discussed in the text.

Figure 14. Consequences of rising temperatures, the formation of a plateau, and the onset of channel flow, for steepening slopes at the plateau edge (erosion front), leading to positive feedback between surface denudation at the erosion front and channel flow.

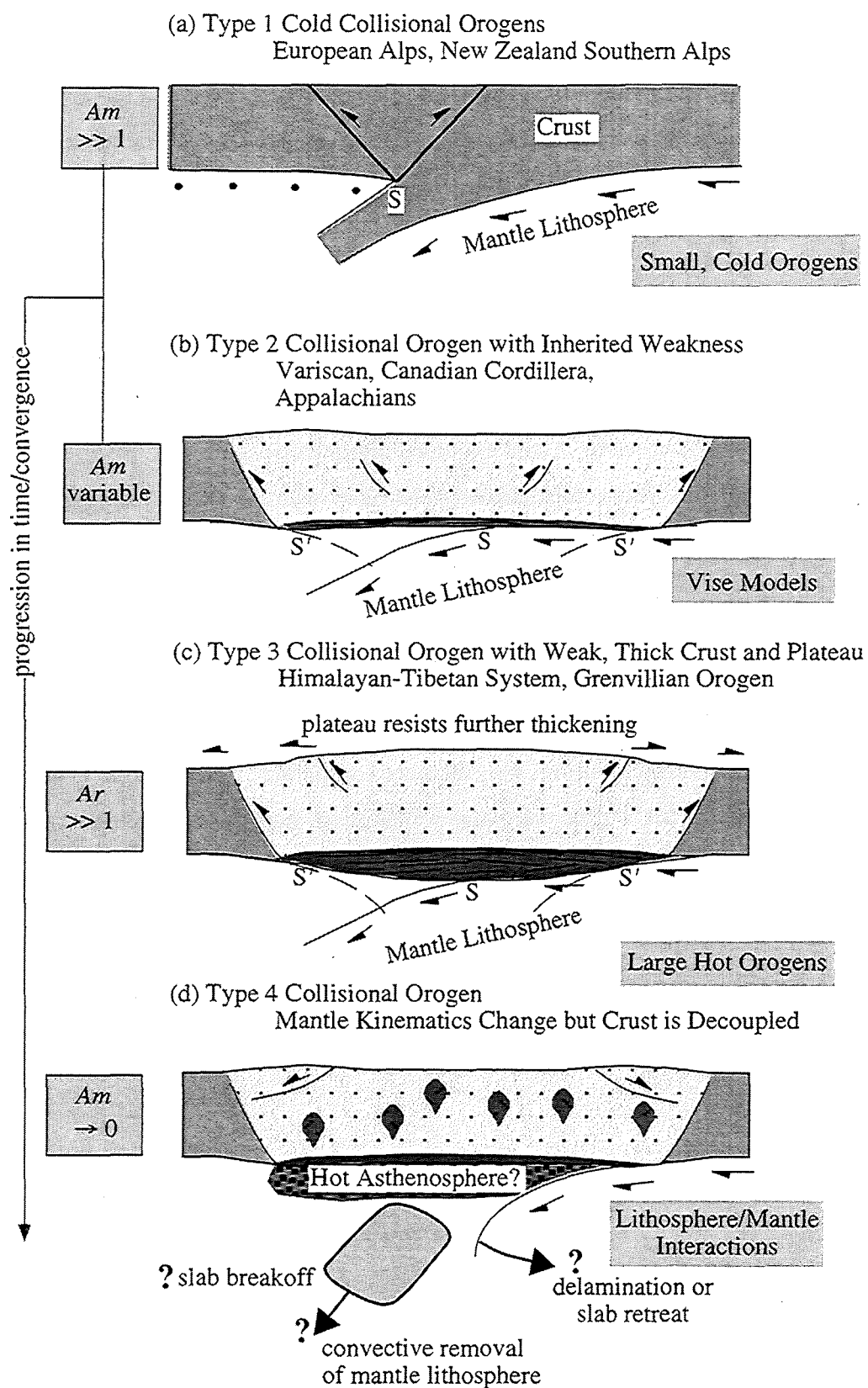
Figure 15. Simplified flow modes seen in the model experiments and their classification. See text for discussion.

Figure 16. Comparison of HTRL-25 (60 My; Fig. 6), with simplified crustal-scale geometry and inferred deformation/flow regimes for the MCT zone, STD system, and intervening Greater Himalayan sequence. Horizontal scales are approximately equal, but note large vertical exaggerations in (a) and (b). (a) Interpretation of Davidson et al. (1997), showing variable velocities and associated ductile extrusion across the High Himalayan Crystalline (HHC) belt. More recent work (Grujic et al. 2000) suggests that the mid-crustal "wedge" geometry is incorrect, and that the region between the MCT and STDS forms a moderately dipping, narrow channel with roughly parallel sides extending a considerable distance beneath the Tibetan plateau. (b) Interpretation of Wu et al. (1998), showing continuity and inferred age progression between partial melt zone beneath the Tibetan plateau and High Himalayan leucogranites (older closer to mountain front); diagram also emphasises the role of surface denudation in exhuming the ductile, partially molten region between the MCT and STD. (c) Geometry of model HTRL-25 at 60 My. See discussion in text.

Figure 17. P-T grid showing conditions beneath present-day Tibetan plateau (*TP*; Hacker et al. 2000) and Central Gneiss Belt of the Grenville orogen (*CGB*; Timmermann 1998) at ca. 1075 Ma compared with P-T-t paths for two tracked points (P19, P23) in model HTRL-25. Shaded fields show approximate metamorphic facies boundaries; labelled reaction lines are for the Al_2SiO_5 system (a = andalusite; k = kyanite; s = sillimanite) and muscovite dehydration (muscovite (m) + quartz (q) = K-feldspar (Kf) + Al_2SiO_5 (a) + H_2O (w)). Ranges for fluid-

present ("wet melt") and fluid-absent ("dehydrated melt") melting are also shown. Lower panels show the corresponding particle paths through the model orogen for the same points. Note that lateral transport is much greater than vertical transport. Path reversals correspond to entrainment of particles in the channel flow zone; exhumation of the particles results from the combination of outward flow in the channel coupled with surface denudation.

Figure 18. (a) Peak temperatures (T_{max} ; upper panel) and corresponding pressures ($P @ T_{\text{max}}$; lower panel) and times (labels, My since model start) for points at the model surface at specified times, plotted as a function of distance across the model in the vicinity of the erosion front. Peak T_{max} increases and profile gets narrower and less symmetric with time. Increase in peak temperature structurally upwards across exhumed region corresponds to "inverted" metamorphic sequence. Note that P distribution is narrower and more asymmetric than corresponding T distribution. Latest times of peak T , corresponding to "young" metamorphic ages, occur on pro-ward side of exhumed zone. (b) Comparison of model peak grade profiles with current metamorphic T (upper panel) and P (lower panel) distributions across the MCT zone and Greater Himalayan sequence in the Langtang region of central Nepal. Model results are in reasonable agreement with the observations, given reported uncertainties in P - T estimates (vertical error bars). However, the model conditions may be more comparable to a time (ca. 15-20 Ma) when exhumation rates in the MCT zone were higher than today (see discussion in text).



HTRL-25: Reference Model, Initial Conditions

$t = 0 \text{ My}$; $\Delta x = 0 \text{ km}$

Denudation rate = Slope $\times f(t) \times g(x)$

$f(t) = 0.107 \text{ m/y}$

$t < 37.5 \text{ My}$

$g(x) = 0$

$t > 37.5 \text{ My}$

$x \leq 1500 \text{ km}$:

$g(x) = 0$;

$1500 < x < 1550 \text{ km}$: $g(x)$ increases linearly with x to 1;

$x \geq 1550 \text{ km}$:

$g(x) = 1$

Marker Grid

600 km



retro-side

Temperature and Velocity

Denudation rate = Slope $\times f(t) \times g(x)$

$q_s = 70 \text{ mW/m}^2$



Conductive Steady-State

$q_m = 20 \text{ mW/m}^2$

Kinematic Subduction

$V_p = 2 \text{ cm/y}$

$T_a = 1350 \text{ }^\circ\text{C}$

Rheological Properties:

Upper crust: $\phi = 15^\circ$

η (Wet Quartzite $\times 5$)

Lower crust: $\phi = 15^\circ$

η (Dry Mld Diabase)

Thermal Properties:

$A_1 = 2.0 \text{ } \mu\text{W/m}^3$

$A_2 = 0.75 \text{ } \mu\text{W/m}^3$

$K = 2.0 \text{ W/m}^\circ\text{K}$

$\kappa = 1.0 \times 10^{-6} \text{ m}^2/\text{s}$

Effective Viscosity:

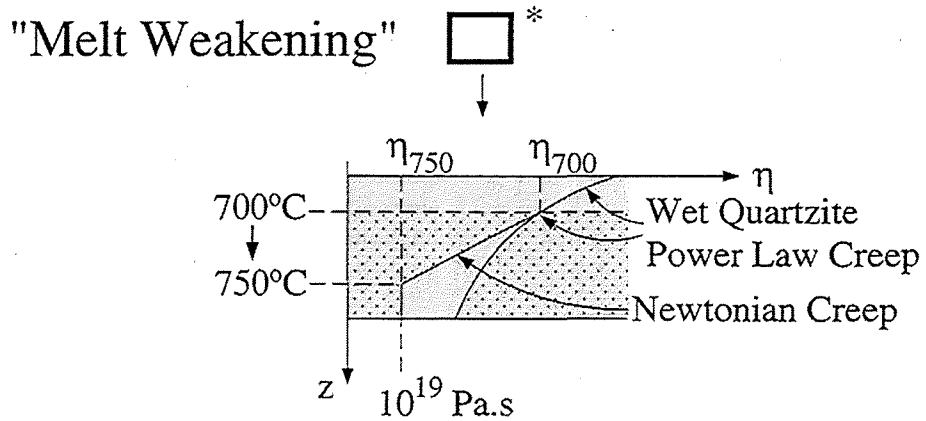
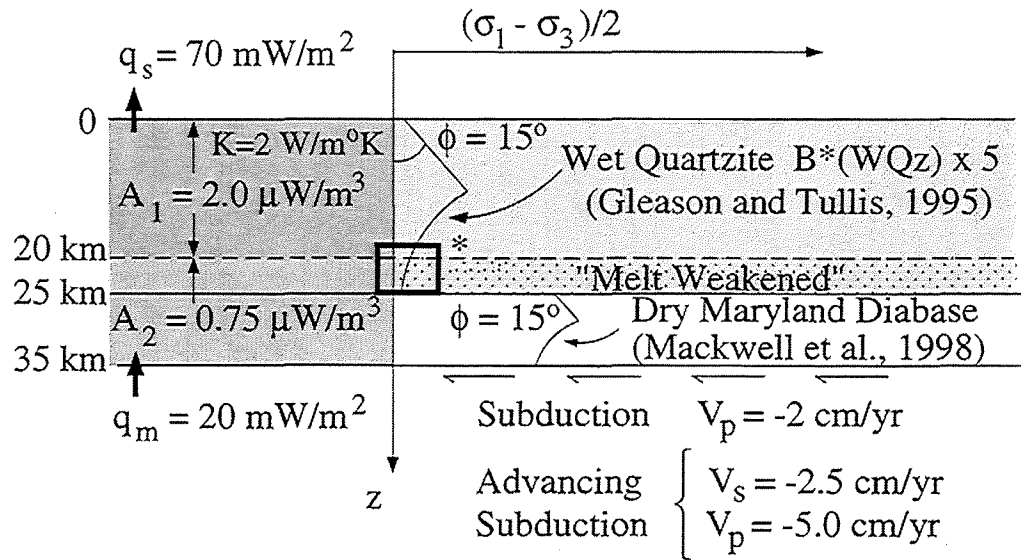
$$\eta = B^* \dot{\epsilon}^{\left(\frac{1}{n}-1\right)} \exp(Q^*/nRT)$$

Heat Balance:

$$\rho C_p (\partial T / \partial t + \underbrace{\mathbf{v} \cdot \nabla T}_{\text{Advection}}) = \underbrace{K \nabla^2 T}_{\text{Diffusion}} + \underbrace{A}_{\text{Heat Production}}$$

Figure 2

Reference Model



HTRL-25: Development of Crustal Channel/Extrusion Counter Flow

$t = 7.5 \text{ My}$; $\Delta x = 150 \text{ km}$

Velocity

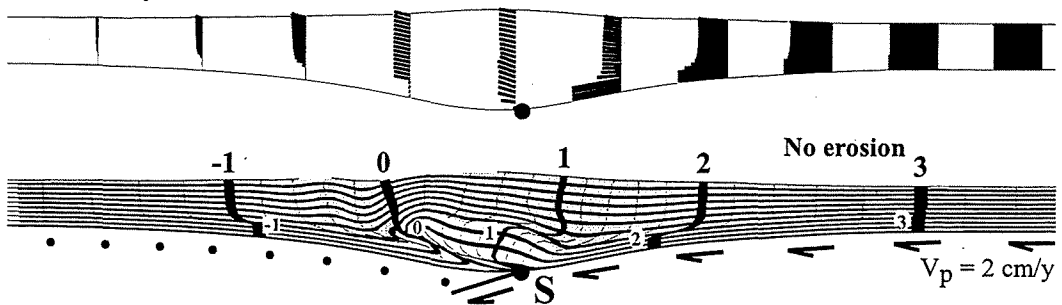
800 km

1600 km

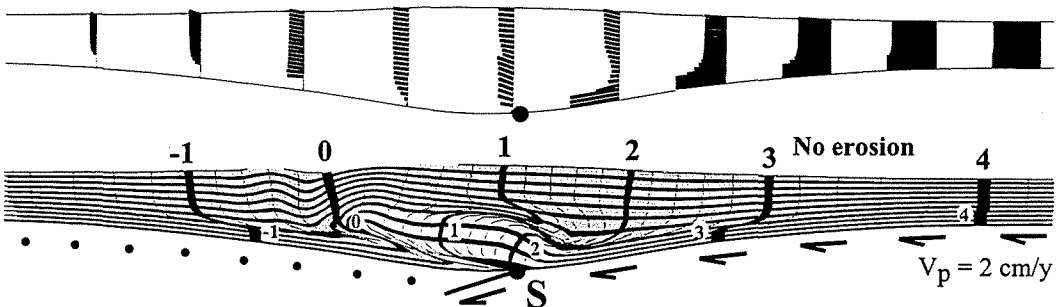
Deformation



$t = 15 \text{ My}$; $\Delta x = 300 \text{ km}$



$t = 22.5 \text{ My}$; $\Delta x = 450 \text{ km}$



cb_lithoprobe_HTRL-25_lag_vel_1.pdf

Figure 4a

HTRL-25:

**Distribution of Radioactivity in Crust and
Associated Temperatures**

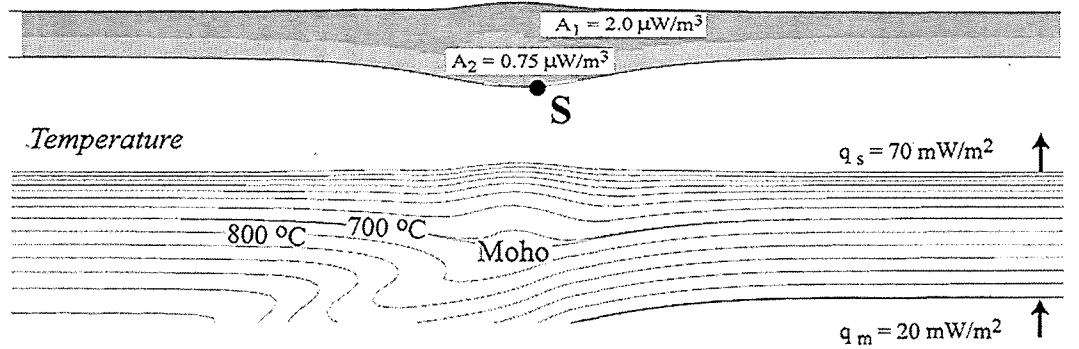
$t = 7.5 \text{ My}; \Delta x = 150 \text{ km}$

Radioactive Heat Sources

800 km

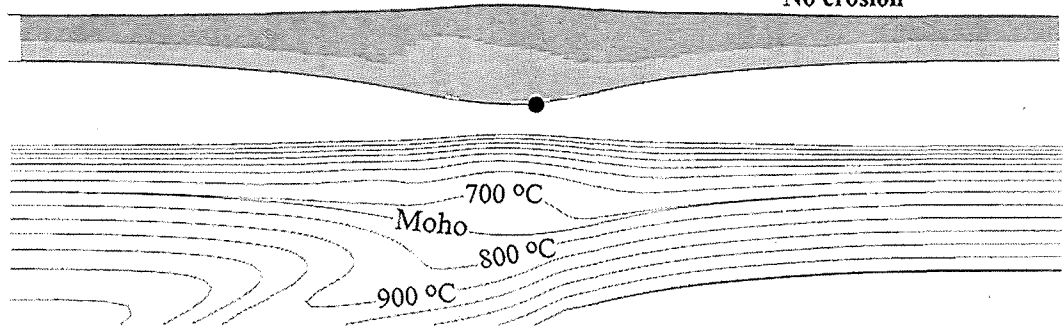
No erosion

1600 km



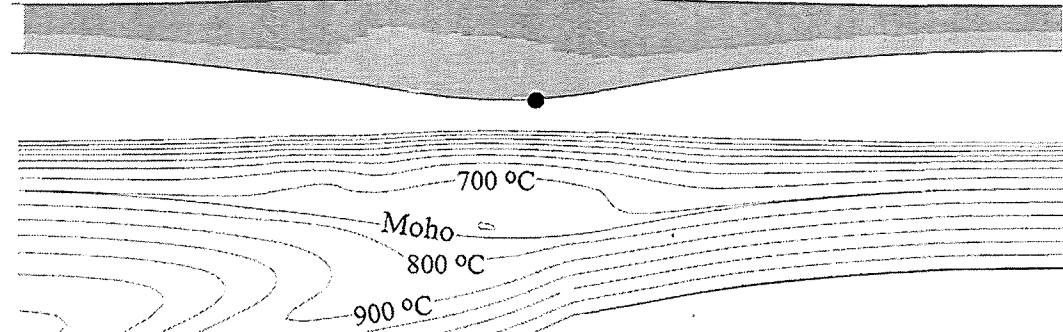
$t = 15 \text{ My}; \Delta x = 300 \text{ km}$

No erosion



$t = 22.5 \text{ My}; \Delta x = 450 \text{ km}$

No erosion



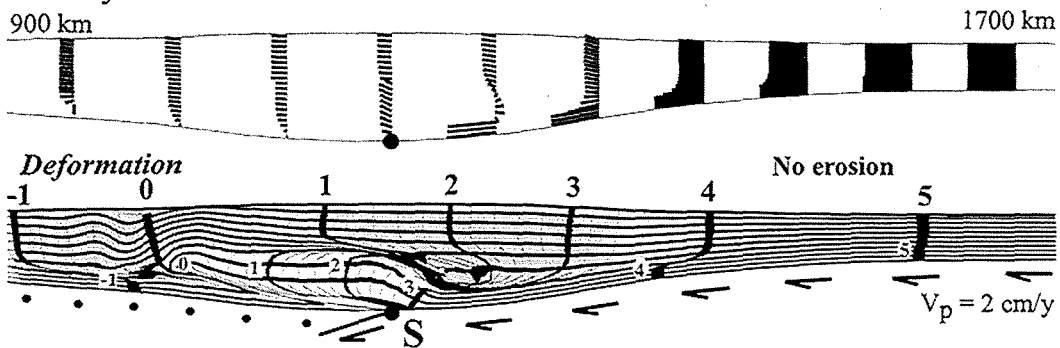
cb_lithoprobe2001_HTRL-25_mat_therm_cont1.pdf

Figure 4b

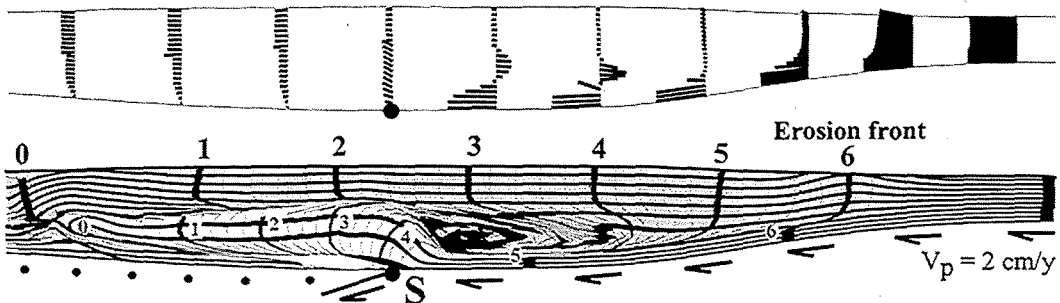
HTRL-25: Development of Crustal Channel/Extrusion Counter Flow

$t = 30 \text{ My}$; $\Delta x = 600 \text{ km}$

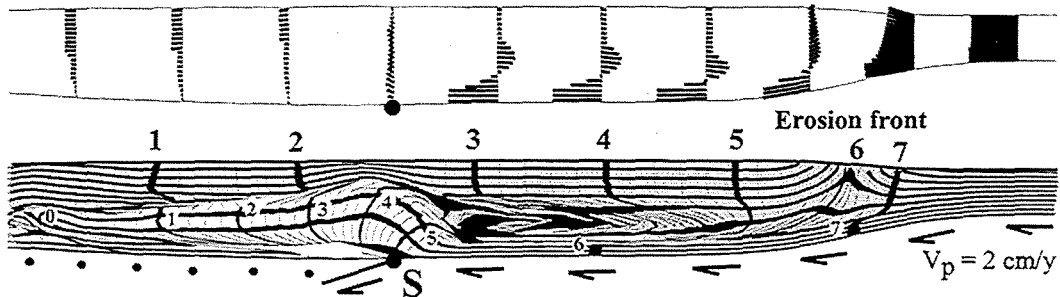
Velocity



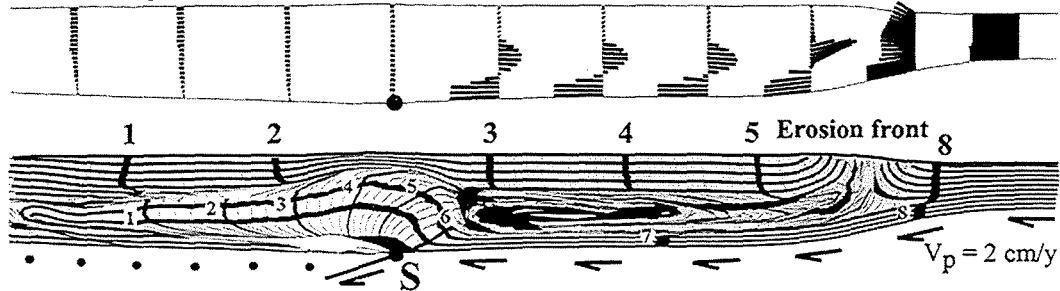
$t = 45 \text{ My}$; $\Delta x = 900 \text{ km}$



$t = 52.5 \text{ My}$; $\Delta x = 1050 \text{ km}$



$t = 60 \text{ My}$; $\Delta x = 1200 \text{ km}$



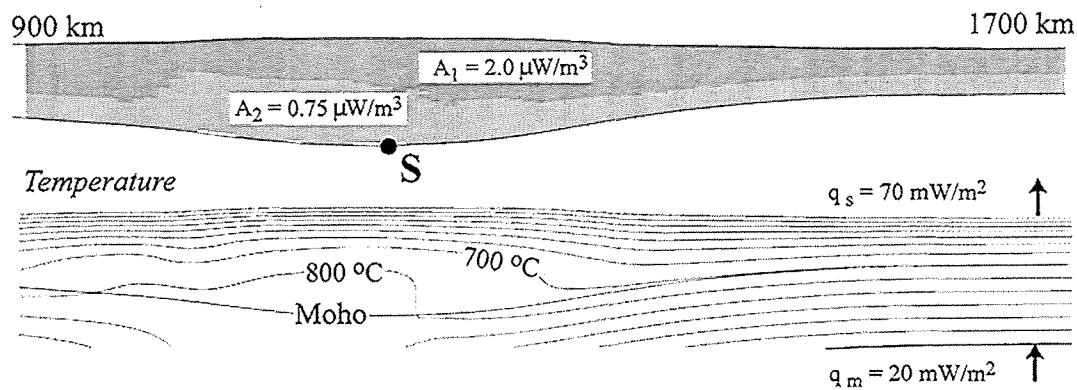
cb_lithoprobe2001_HTRL-25_lag_vel_2.pdf

Figure 5a

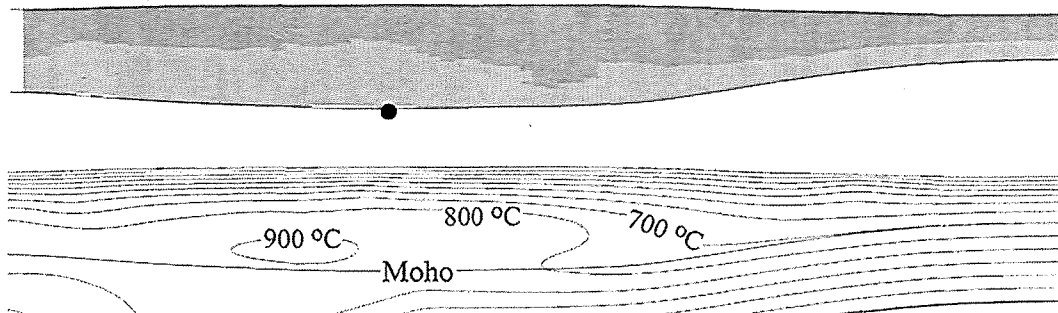
HTRL-25: Distribution of Radioactivity in Crust and Associated Temperatures

$t = 30 \text{ My}; \Delta x = 600 \text{ km}$

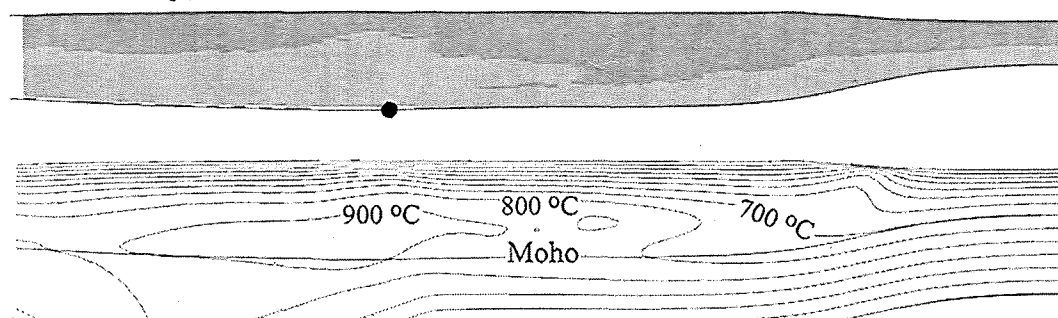
Radioactive Heat Sources



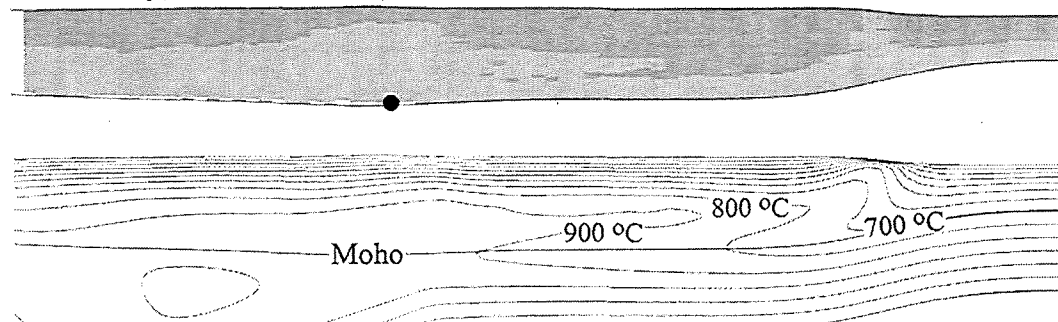
$t = 45 \text{ My}; \Delta x = 900 \text{ km}$



$t = 52.5 \text{ My}; \Delta x = 1050 \text{ km}$



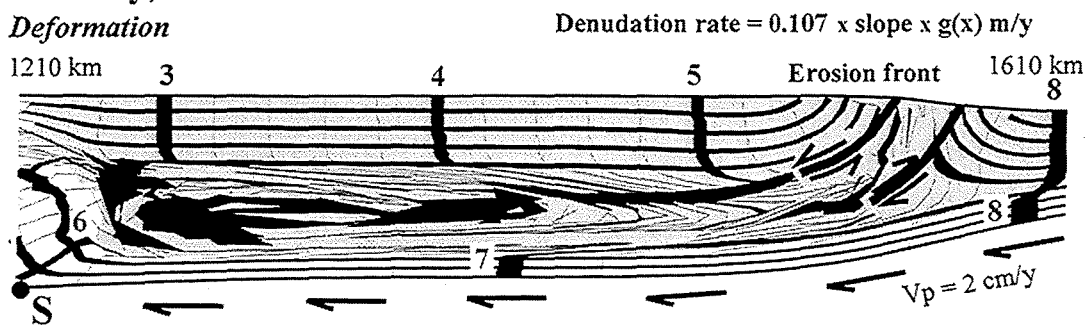
$t = 60 \text{ My}; \Delta x = 1200 \text{ km}$



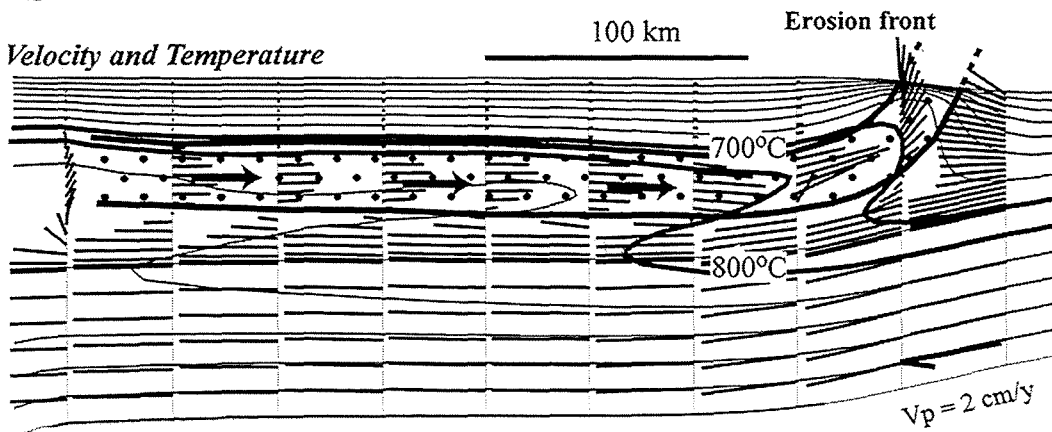
HTRL-25: Mid-Crust Extrusion/Channel Flow Channel response to erosion

$t = 60 \text{ My}$; $\Delta x = 1200 \text{ km}$

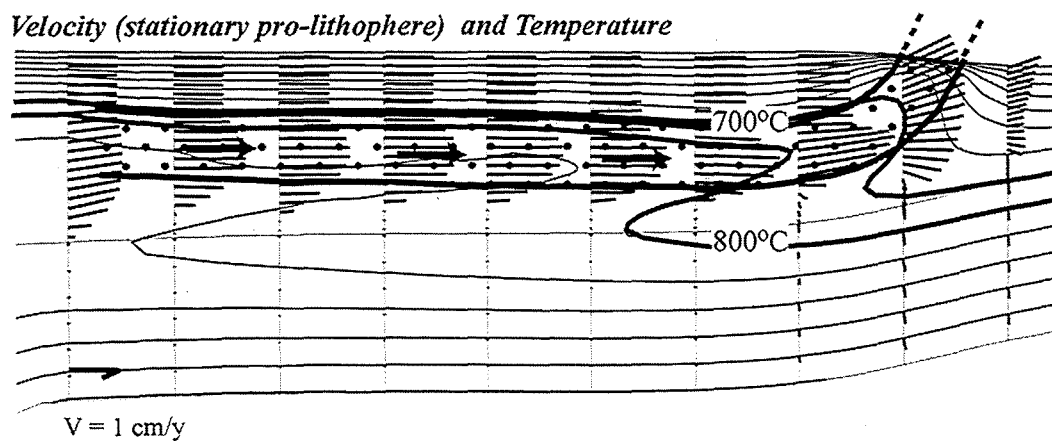
a) *Deformation*



b) *Velocity and Temperature*



c) *Velocity (stationary pro-lithosphere) and Temperature*



Cb_lithoprobe_HTRL25_ts000_lag_therm_vel.pdf

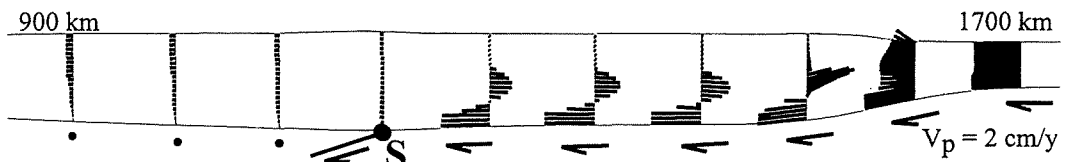
Figure 6

HTRL-25

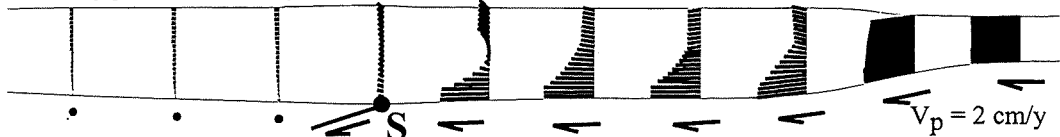
Boundary Conditions and Gravitational Contributions to the Flow Regime

$t = 60 \text{ My}$; $\Delta x = 1200 \text{ km}$

Total Velocity



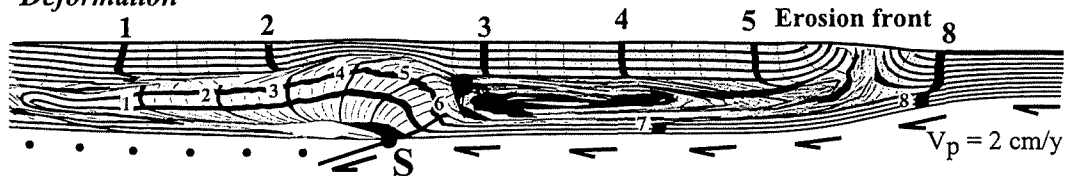
Velocity from boundary conditions alone



Velocity from gravity forcing alone

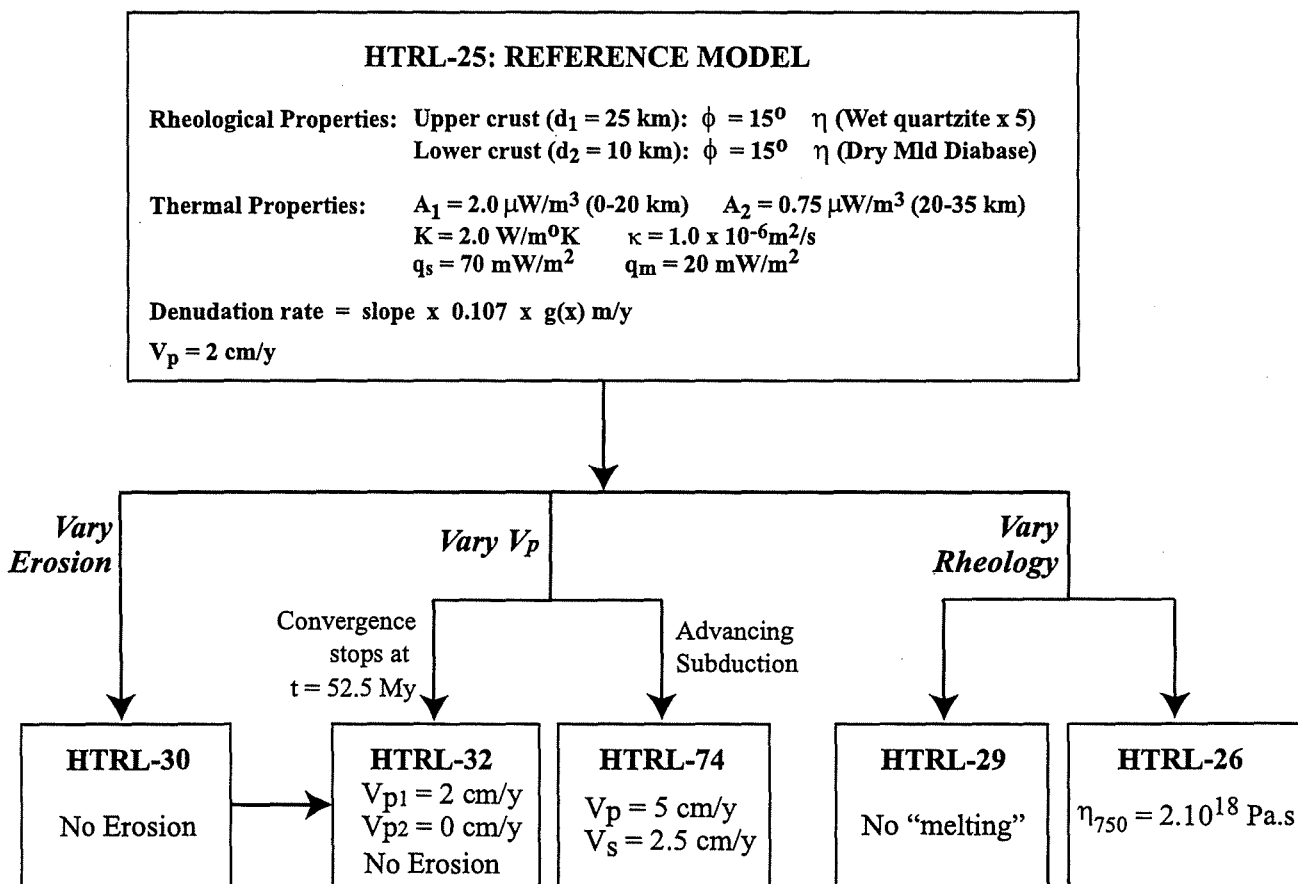


Deformation



cb_lithoprob2001_HTRL-25_lag_3-vcl.pdf

Figure 7



cb_lithoprobe2001_flowchart.ai

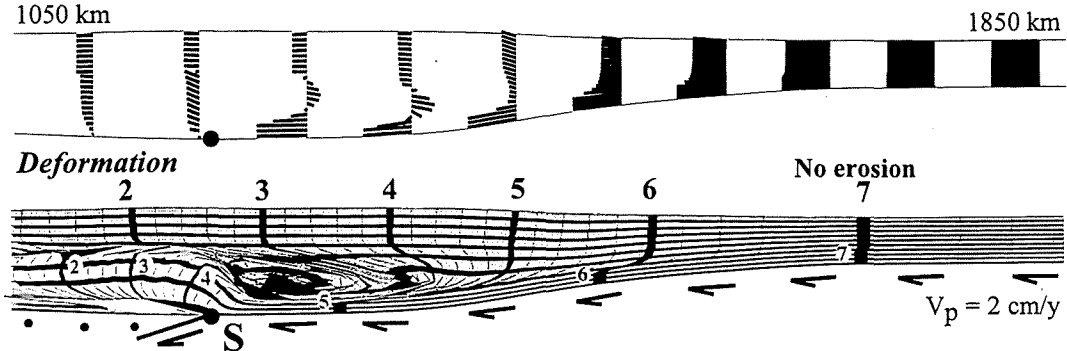
Figure 8

HTRL-30: No Erosion Mid-Crust Channel Flow

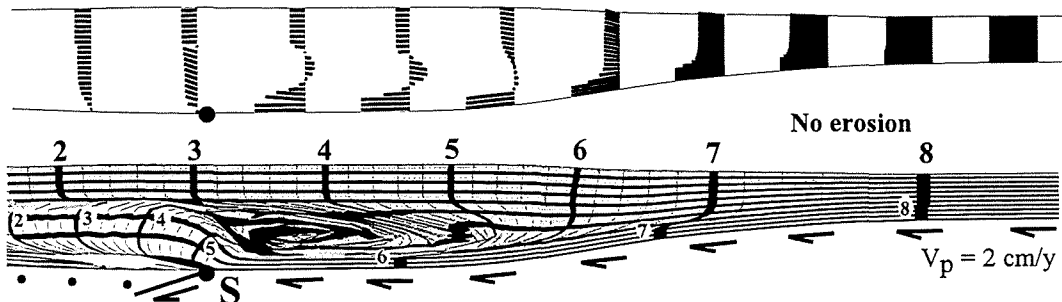
$t = 45 \text{ My}$; $\Delta x = 900 \text{ km}$

Velocity

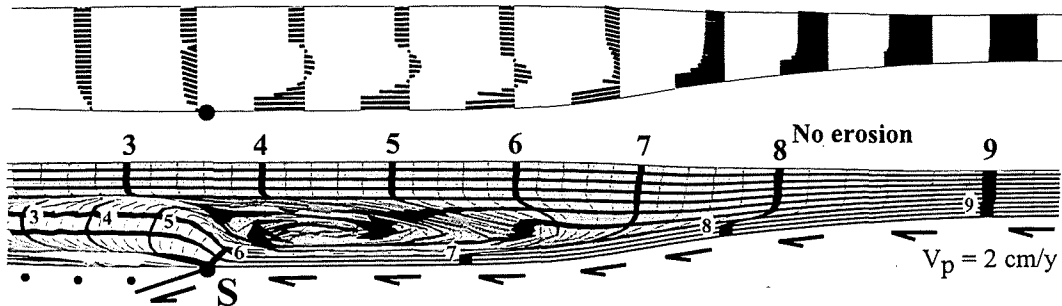
1050 km



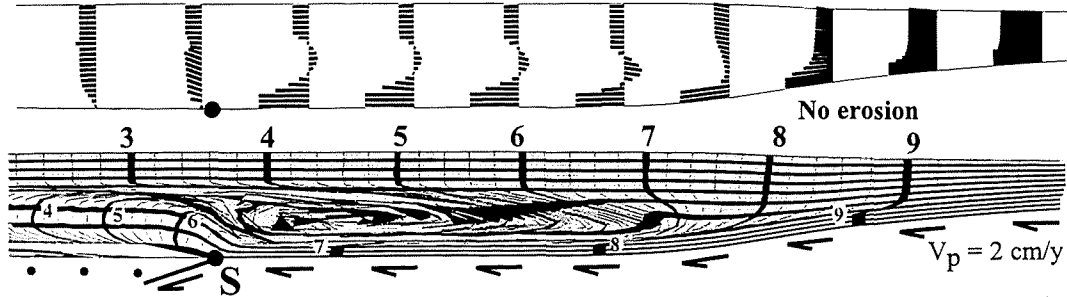
$t = 52.5 \text{ My}$; $\Delta x = 1050 \text{ km}$



$t = 60 \text{ My}$; $\Delta x = 1200 \text{ km}$



$t = 75 \text{ My}$; $\Delta x = 1500 \text{ km}$



cb_lithoprobe_HTRL-30_lag_vel.pdf

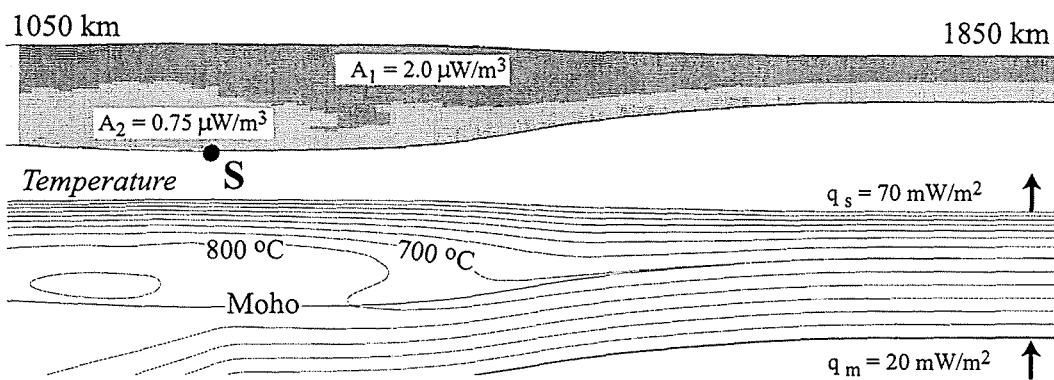
Figure 9a

HTRL-30:

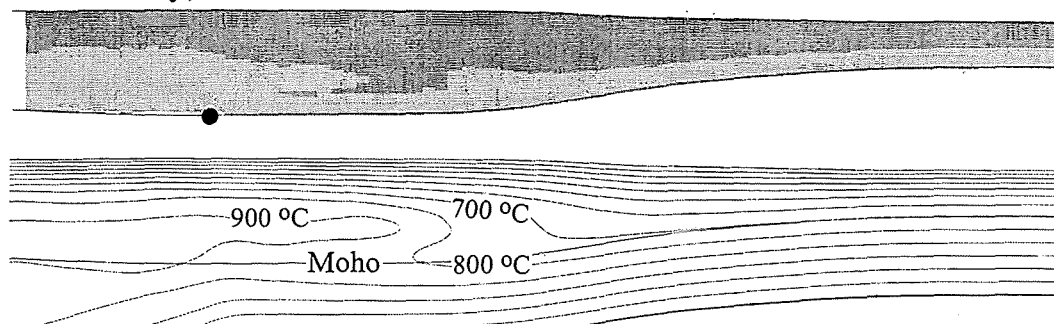
Distribution of Radioactivity in Crust and Associated Temperatures

$t = 45 \text{ My}$; $\Delta x = 900 \text{ km}$

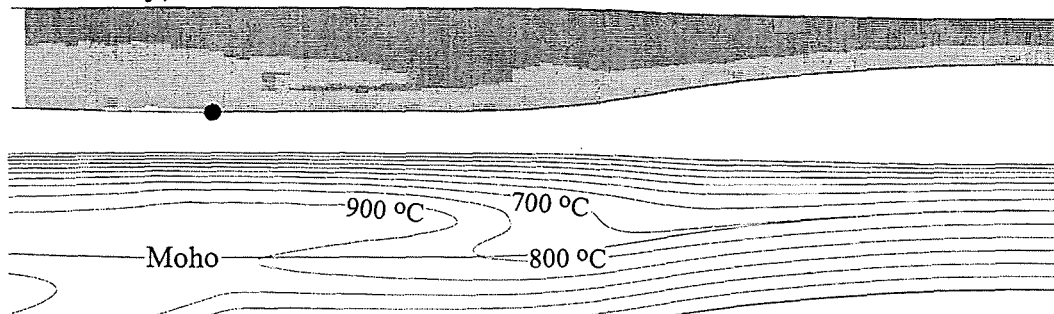
Radioactive Heat Sources



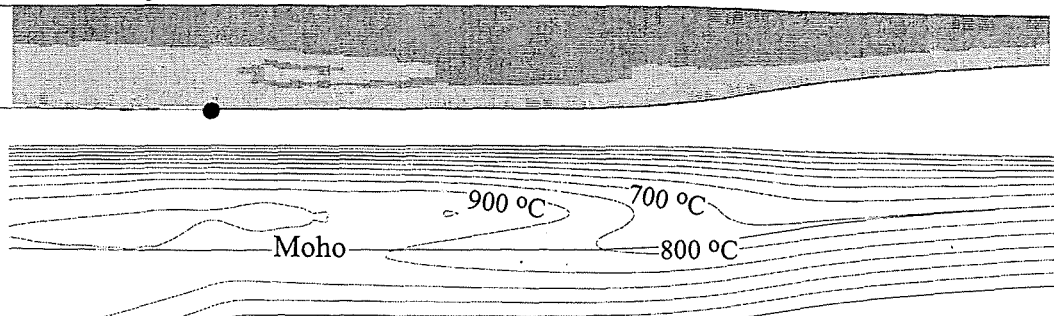
$t = 52.5 \text{ My}$; $\Delta x = 1050 \text{ km}$



$t = 60 \text{ My}$; $\Delta x = 1200 \text{ km}$



$t = 75 \text{ My}$; $\Delta x = 1500 \text{ km}$



cb_lithoprobe2001_HTRL-30_mat_therm_cont.pdf

Figure 9b

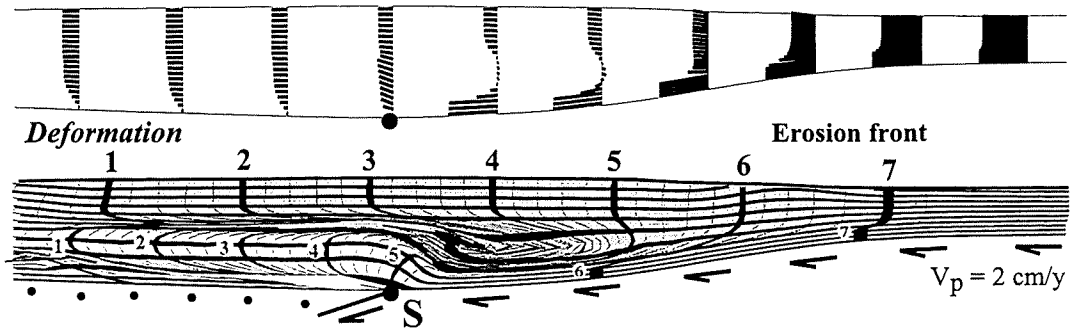
HTRL-29: Crustal Flow Regime for No Melt Weakening

$t = 52.5 \text{ My}$; $\Delta x = 1050 \text{ km}$

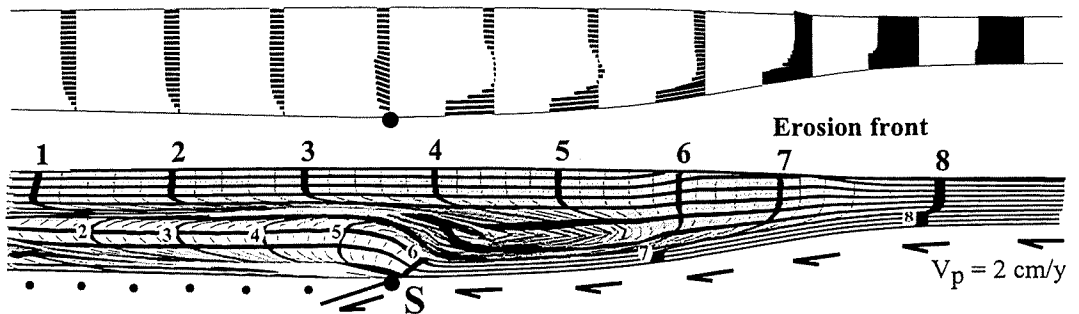
Velocity

900 km

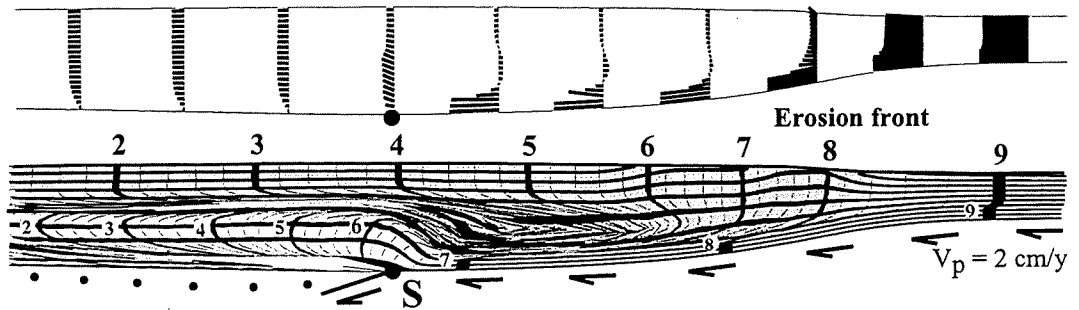
1700 km



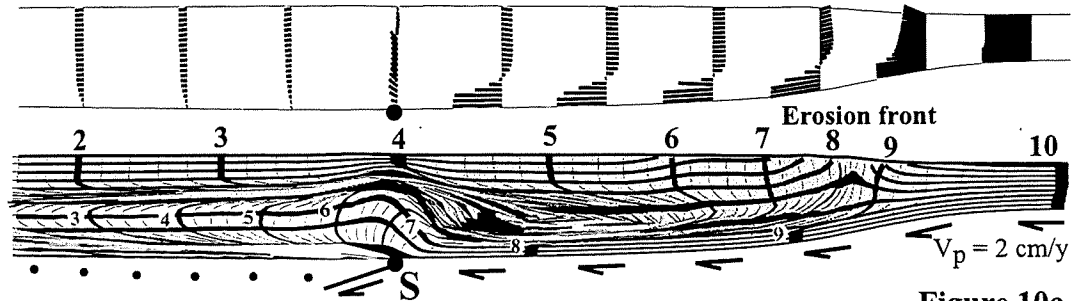
$t = 60 \text{ My}$; $\Delta x = 1200 \text{ km}$



$t = 67.5 \text{ My}$; $\Delta x = 1350 \text{ km}$



$t = 75 \text{ My}$; $\Delta x = 1500 \text{ km}$



cb_lithoprobe2001_HTRL-29_lag_vel.pdf

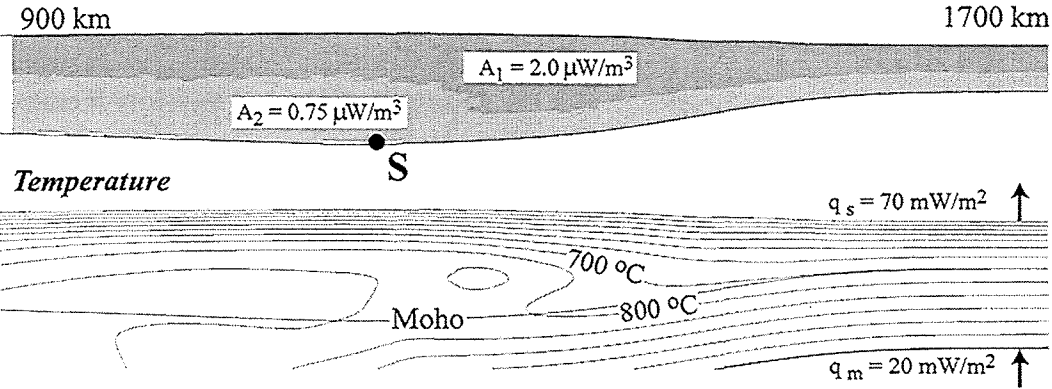
Figure 10a

HTRL-29:

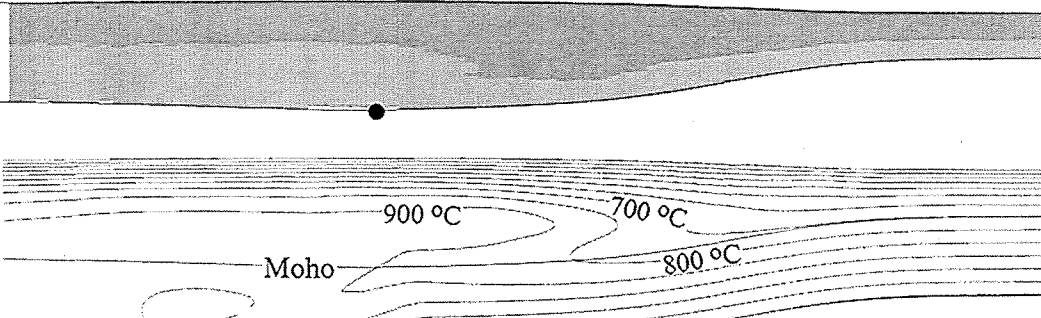
Distribution of Radioactivity in Crust and Associated Temperatures

$t = 52.5 \text{ My}$; $\Delta x = 1050 \text{ km}$

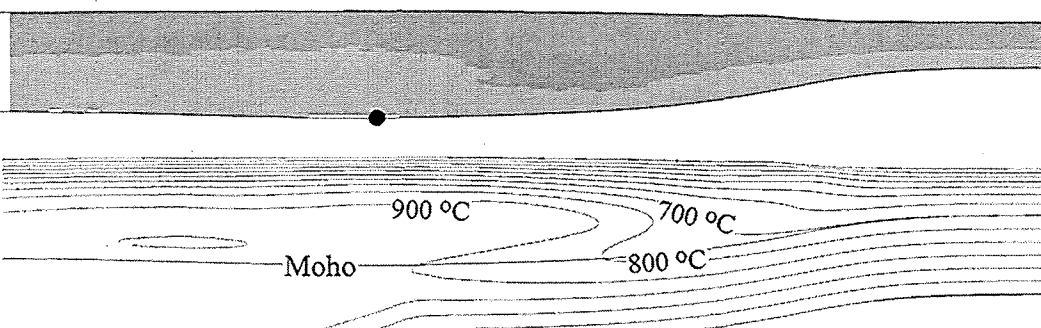
Radioactive Heat Sources



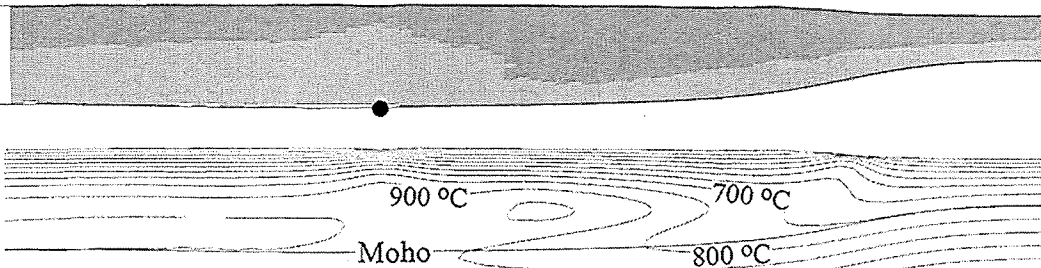
$t = 60 \text{ My}$; $\Delta x = 1200 \text{ km}$



$t = 67.5 \text{ My}$; $\Delta x = 1350 \text{ km}$



$t = 75 \text{ My}$; $\Delta x = 1500 \text{ km}$

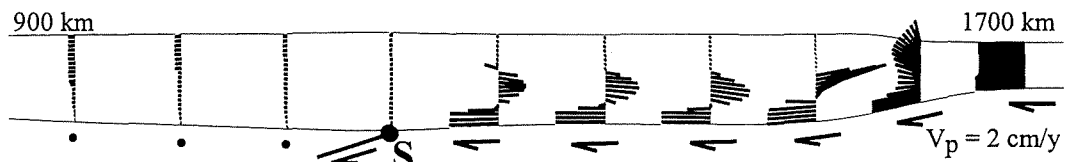


HTRL-26

Boundary Conditions and Gravitational Contributions to the Flow Regime

$t = 60 \text{ My}$; $\Delta x = 1200 \text{ km}$

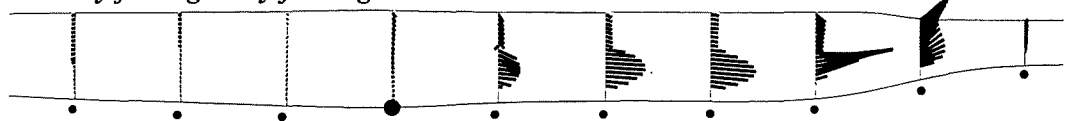
Total Velocity



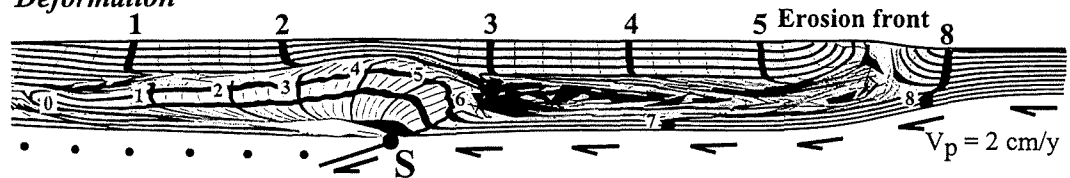
Velocity from boundary conditions alone



Velocity from gravity forcing alone



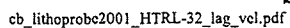
Deformation



cb_lithoprobe2001_HTRL-26_ts8000_lag_3-vel.pdf

Figure 11

1000 km



162

HTRL-32:

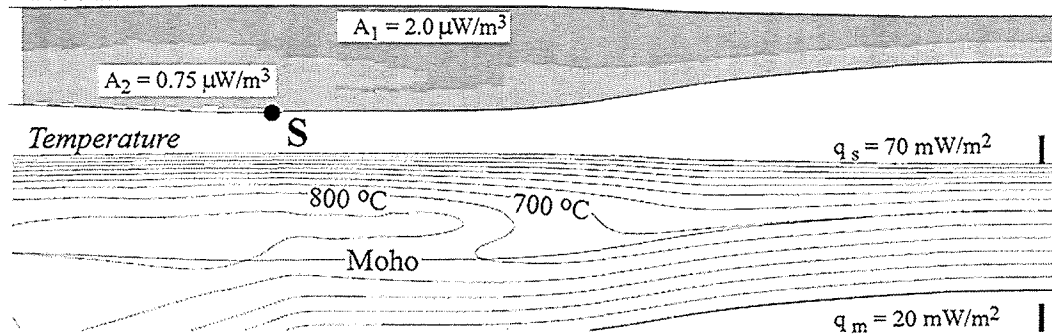
Distribution of Radioactivity in Crust and Associated Temperatures

$t = 52.5 \text{ My}$; $\Delta x = 1050 \text{ km}$

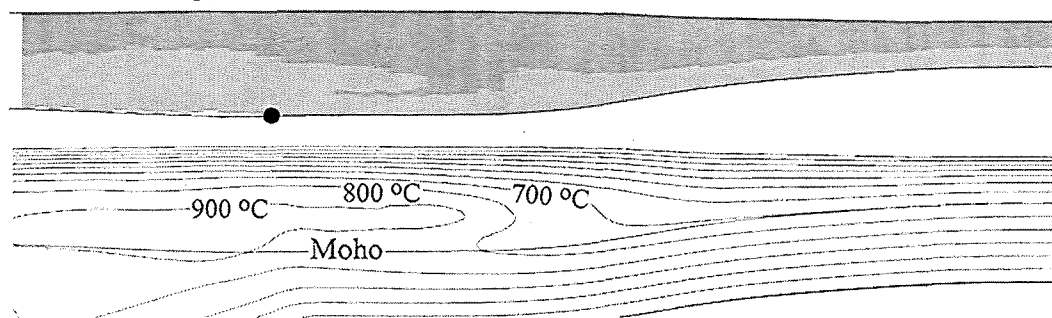
Radioactive Heat Sources

1000 km

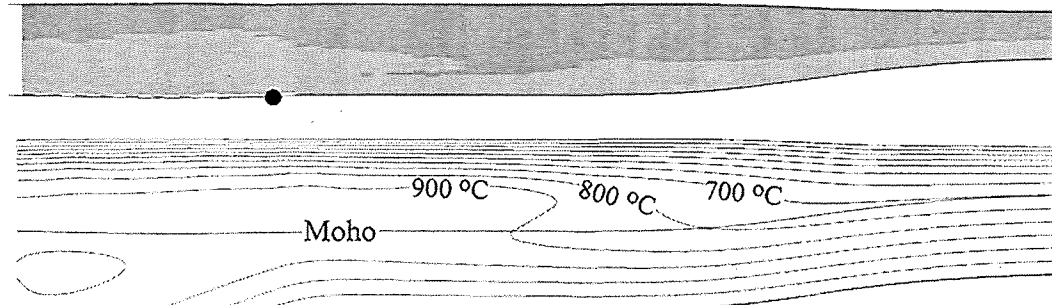
1800 km



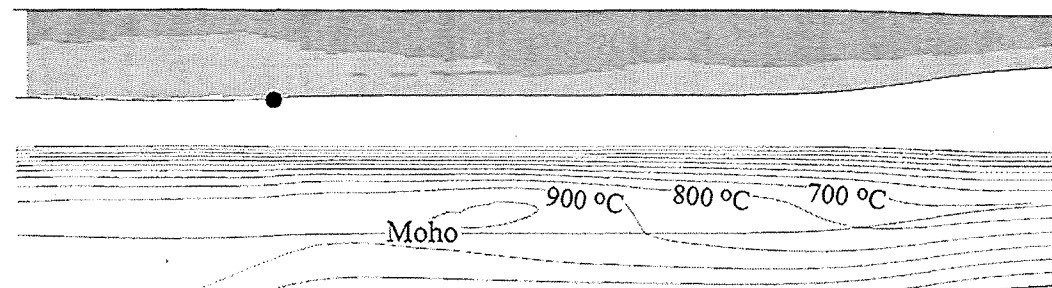
$t = 52.5^+ \text{ My}$; $\Delta x = 1050 \text{ km}$



$t = 60 \text{ My}$; $\Delta x = 1050 \text{ km}$



$t = 67.5 \text{ My}$; $\Delta x = 1050 \text{ km}$



cb_lithoprob2001_HTRL-32_mat_therm_cont.pdf

Figure 12b

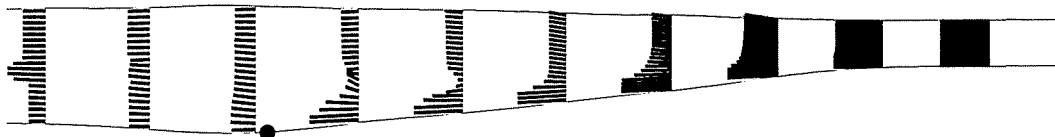
HTRL-74:

Advancing Subduction Model

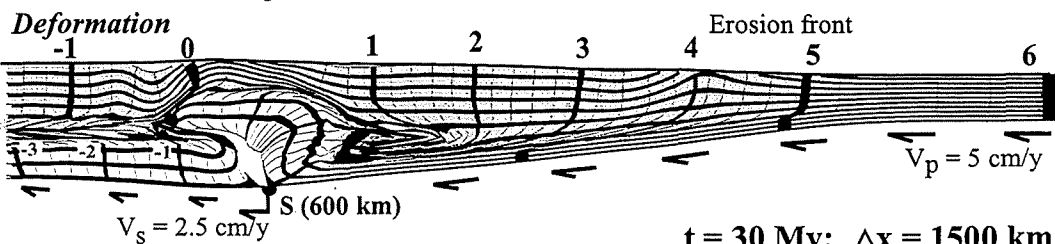
Velocity

$t = 24 \text{ My}; \Delta x = 1200 \text{ km}$

400 km 600 km 1200 km

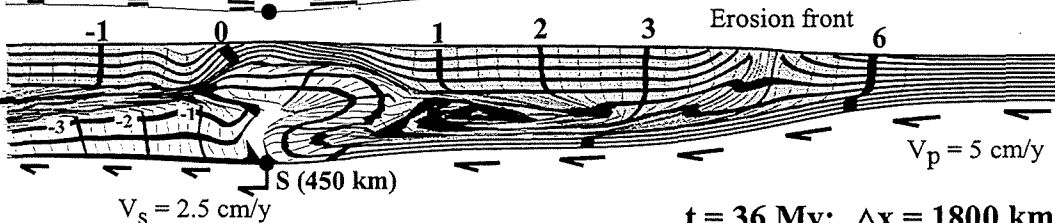


Deformation



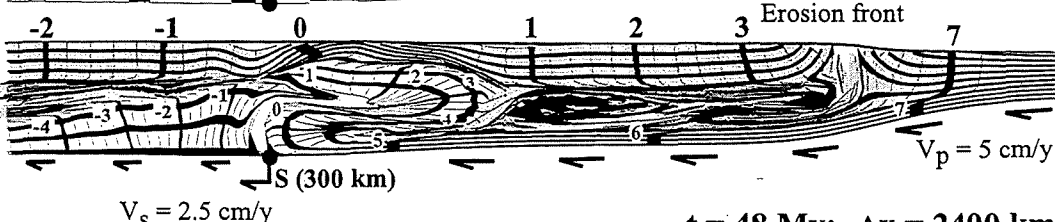
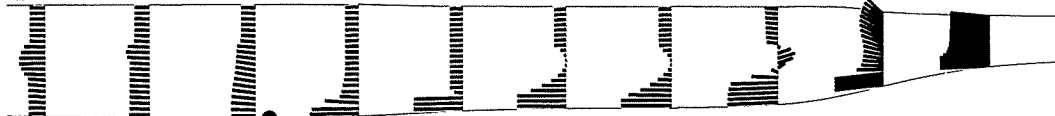
$t = 30 \text{ My}; \Delta x = 1500 \text{ km}$

250 km 450 km 1050 km



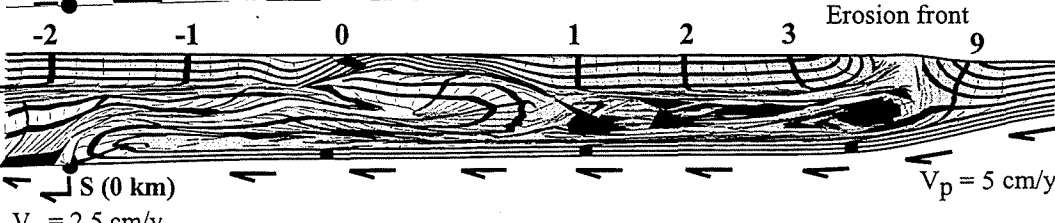
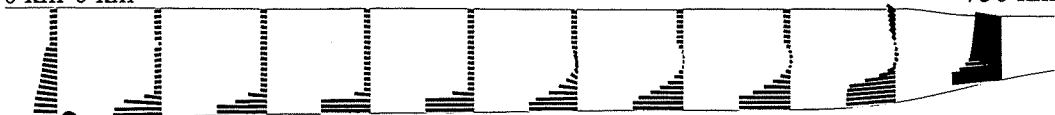
$t = 36 \text{ My}; \Delta x = 1800 \text{ km}$

100 km 300 km 900 km



$t = 48 \text{ My}; \Delta x = 2400 \text{ km}$

-50 km 0 km 750 km



Cb_lithoprobe2001_HTRL-74_lag_vel.pdf

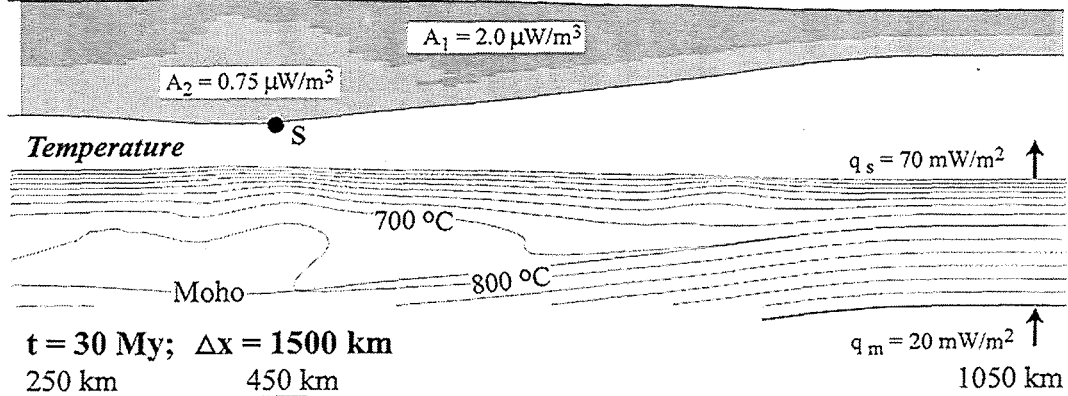
Figure 13a

HTRL-74: Distribution of Radioactivity in Crust and Associated Temperatures

$t = 24 \text{ My}; \Delta x = 1200 \text{ km}$

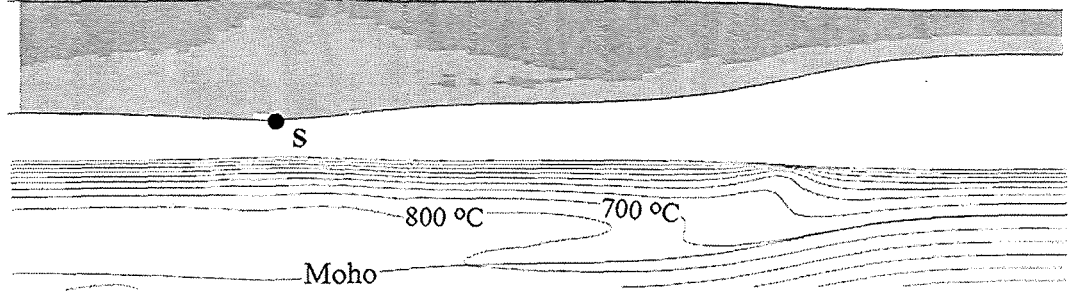
Radioactive Heat Sources

400 km 600 km 1200 km



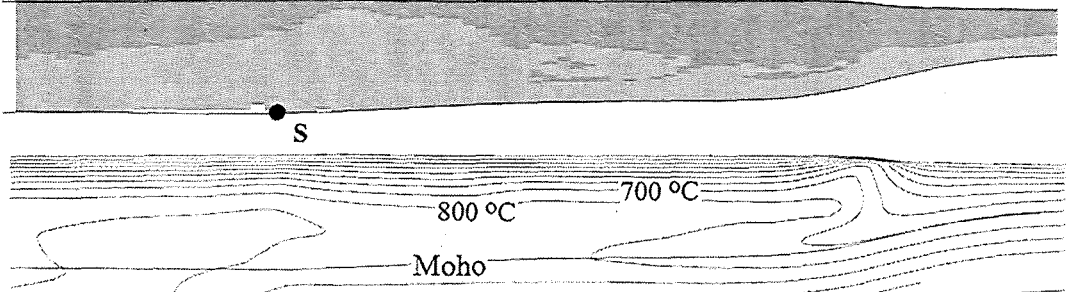
$t = 30 \text{ My}; \Delta x = 1500 \text{ km}$

250 km 450 km 1050 km



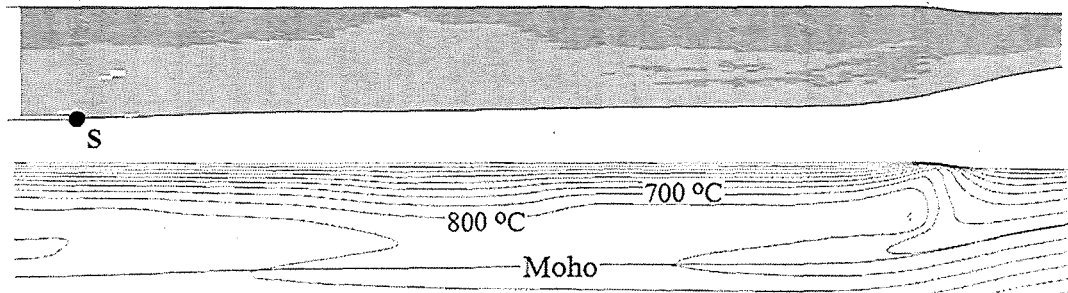
$t = 36 \text{ My}; \Delta x = 1800 \text{ km}$

100 km 300 km 900 km



$t = 48 \text{ My}; \Delta x = 2400 \text{ km}$

-50 km 0 km 750 km



Feedback Loops - Erosion/Channel Coupling

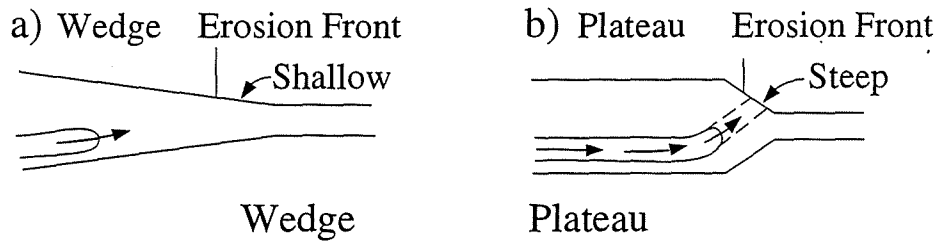


Figure 14

Flow Modes

Mode 1 Pure Shear

$$v_u \sim v_l$$

Mode 2 Detached
Upper Crust

$$v_u < v_l$$

Mode 3 Upper Crustal
Channel/Extrusion

$$v_u < v_l, v_c \text{ (-ve)}$$

Mode 4 'Gravity Sliding'
of Upper Crust

$$v_u \text{ (-ve)}$$

Mode 5 Gravity Sliding and
Channel Flow of
Upper Crust

$$v_c \text{ and } v_u \text{ (-ve)}$$

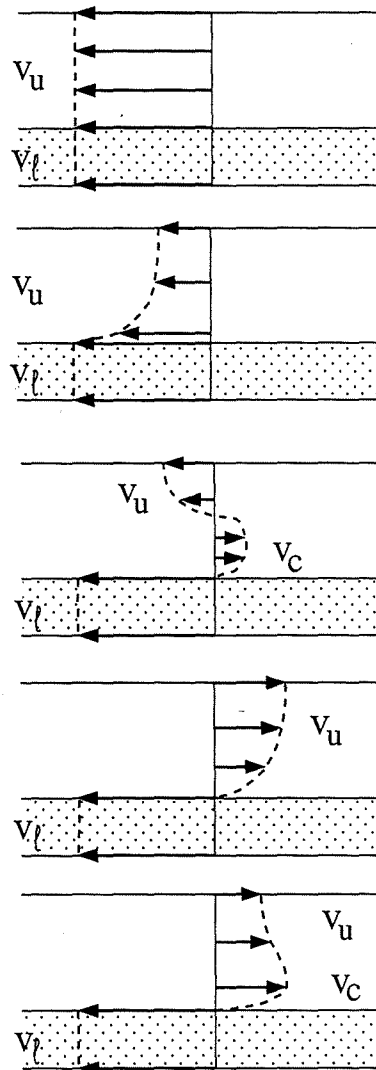
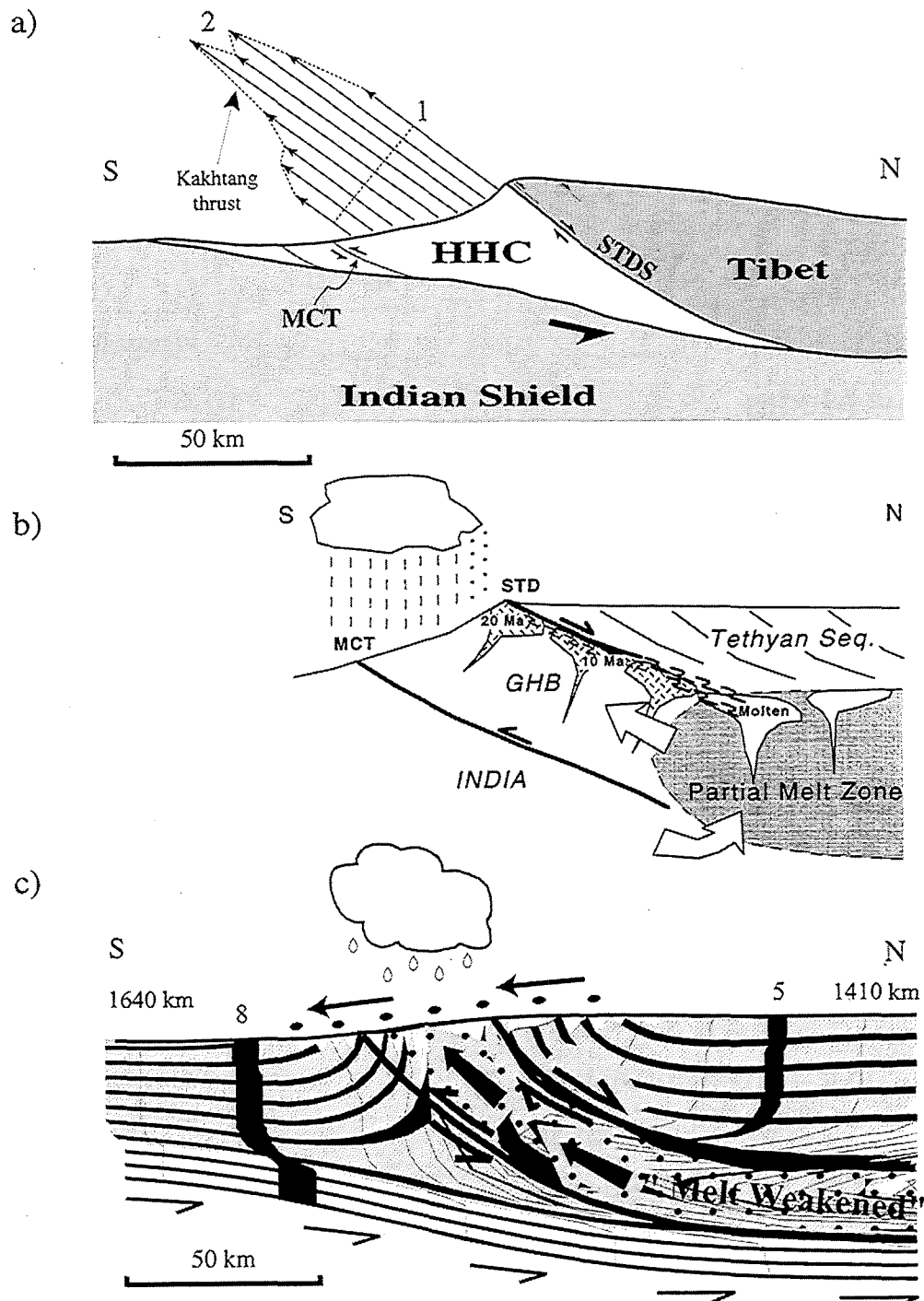


Figure 15



cb_lithoprobe2001_Fig16.pdf

Figure 16

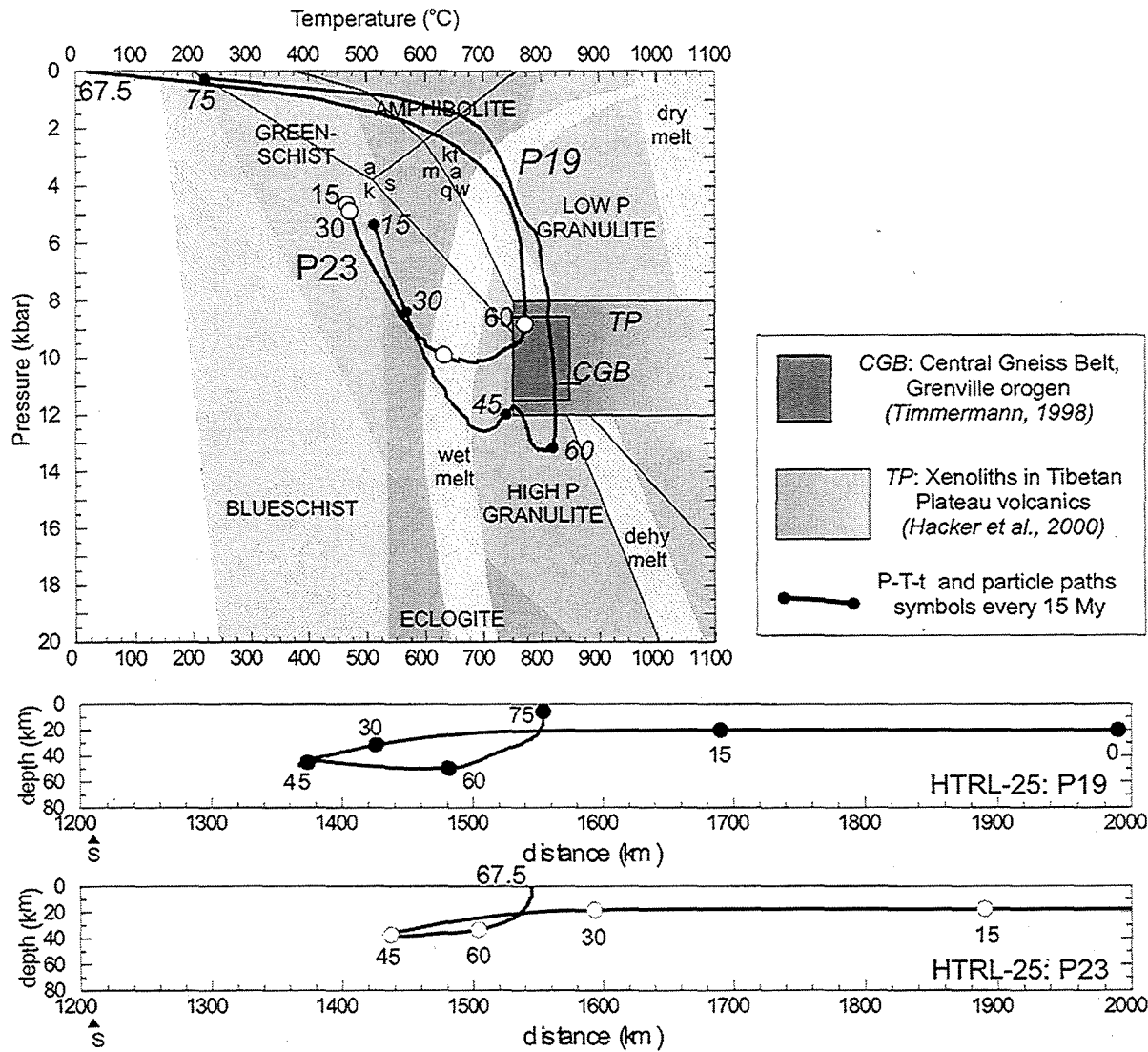
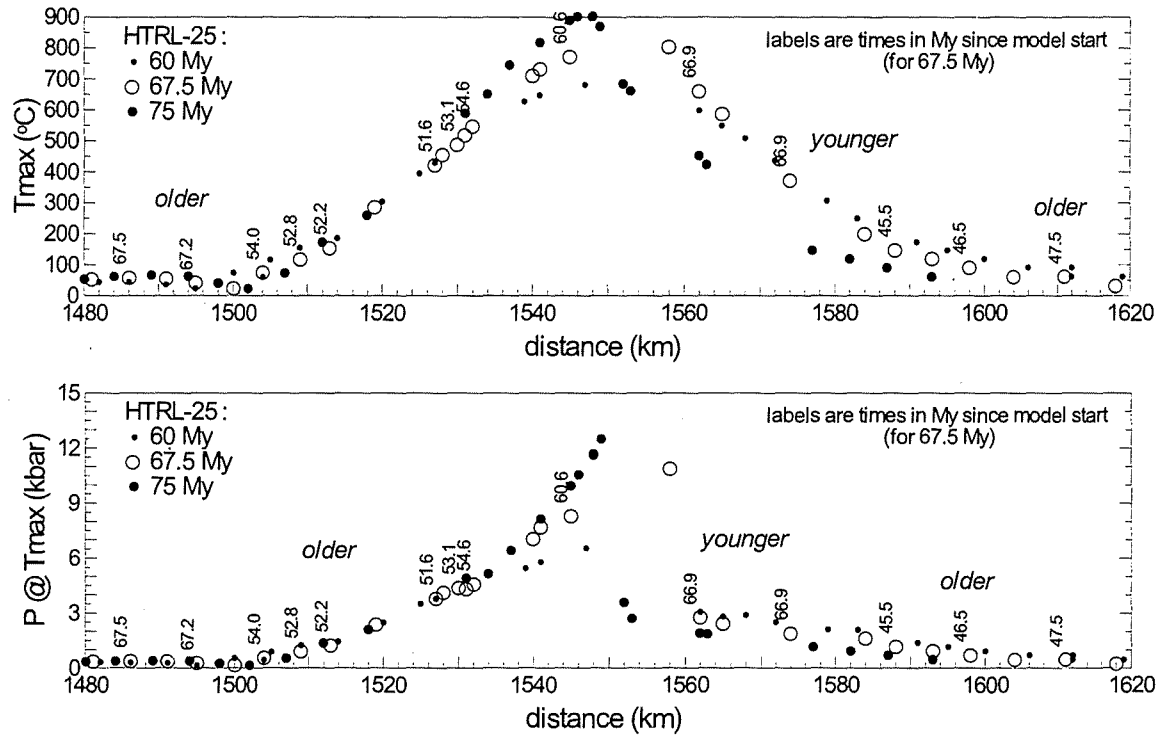


Figure 17

a) Peak grade profiles, HTRL-25



b) Comparison with P-T data from MCT zone, central Nepal

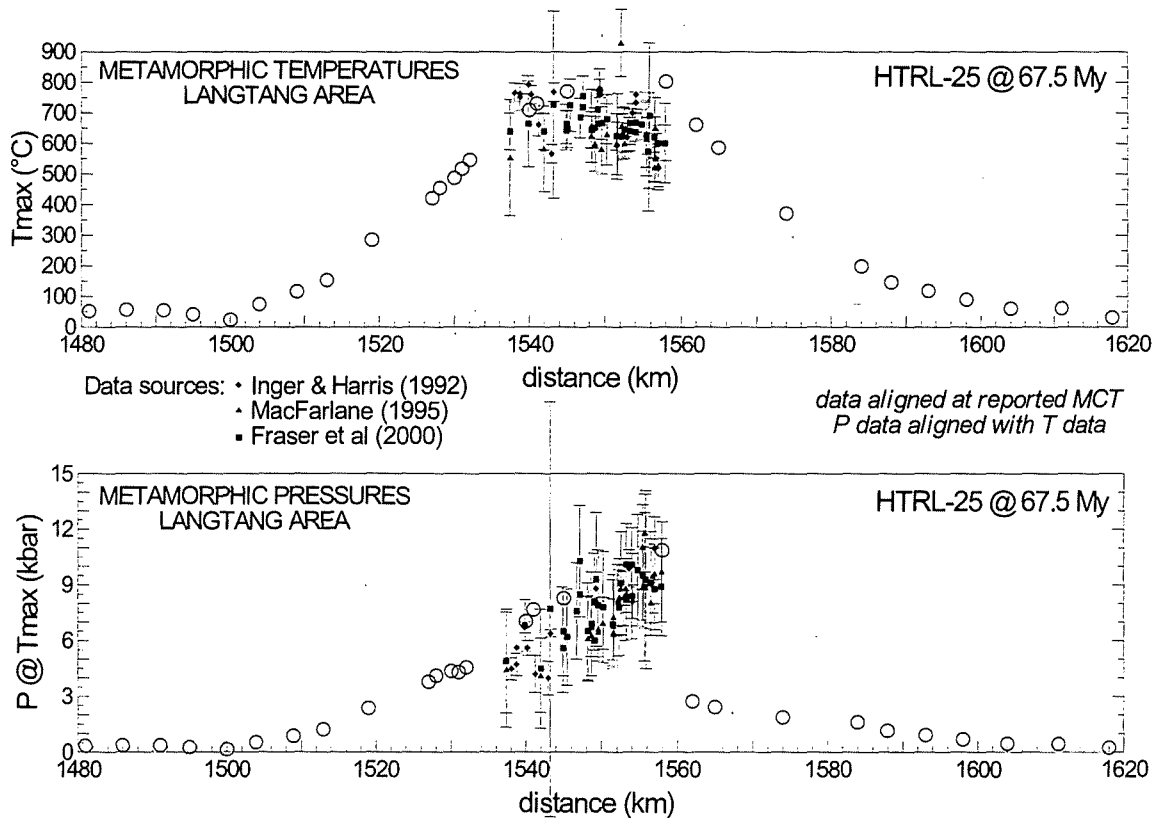


Table 1 Standard Model Properties

Upper Crust

Thickness	d_1	25 km	
Density	ρ_c	2700 kg/m ³	
Thermal Conductivity	K	2.0 W/m ^{°K}	
Heat Production (0-20 km)	A_1	2.0 μ W/m ³	
Internal Angle of Friction	ϕ	15°	
Viscosity (PLC ⁺) parameters (Wet Quartzite)		$B^* = 2.92 \times 10^6 \text{ Pa.s}^{1/4.0}$ $Q^* = 223 \text{ kJ/mol}$	$n = 4.0$
(In model $B^* \rightarrow B^* \times 5$ (see text))		$B^* = 1.46 \times 10^7 \text{ Pa.s}^{1/4.0}$	
Melt Weakening		700°C $\rightarrow$ 750°C	
$\eta_{700} \rightarrow \eta_{750}$		$\eta(\text{PLC}) \rightarrow 10^{19} \text{ Pa.s}$	

Lower Crust

Thickness	d_2	10 km	
Density	ρ_c	2700 kg/m ³	
Thermal Conductivity	K	2.0 W/m ^{°K}	
Heat Production (20-35 km)	A_2	0.75 μ W/m ³	
Internal Angle of Friction	ϕ	15°	
Viscosity (PLC ⁺) parameters (Dry Maryland Diabase)		$B^* = 1.91 \times 10^5 \text{ Pa.s}^{1/4.7}$ $Q^* = 485 \text{ kJ/mol}$	$n = 4.7$
No Melt Weakening			

Mantle Heat Flux	q_m	20 mW/m ²
Surface Heat Flux	q_s	70 mW/m ²
Mantle Density	ρ_m	3300 kg/m ³
Moho Temperature	T_{MOHO}	704°C

Initial Mechanical Model Domain $0 \leq x \leq 2000 \text{ km}$
 $0 \leq z \leq 35 \text{ km}$
Initial Thermal Model Domain (see Fig. 2)

Finite Element Grids		
Mechanical, Eulerian	nx, nz	201, 41
Mechanical, Lagrangian	nx, nz	501, 41
Thermal, Eulerian	nx, nz	201, 68

Initial Thermal State
Conductive Equilibrium Thermal Peclet Number = 0

Basal Boundary Conditions
Subduction $V_p = -2 \text{ cm/y}$
Advancing Subduction $V_s = -2.5 \text{ cm/y}$ $V_p = -5 \text{ cm/y}$
(- sign is strictly correct in Fig. 2 coordinate system)
Denudation (see Fig. 2)

*Definition of PLC (Power Law Creep) $\eta = B^* \dot{\epsilon}^{\left(\frac{1-n}{n}\right)} \exp(Q^*/nRT)$

THE PLEASANT VALLEY THRUST (WESTERN OMINECA CRYSTALLINE BELT): REVISITED AND REDEFINED.

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The Pleasant Valley Thrust (PVT), located in the Cariboo Mountains, British Columbia has been interpreted to be one of the largest west-verging thrust faults within the Omineca Crystalline Belt. The PVT has a remarkably large thrust flat where stratigraphic units in the hangingwall and footwall are parallel to the thrust over an area of at least 250 km by 50 km. The PVT is interpreted to place Neoproterozoic Windermere Supergroup rocks on top of Paleozoic rocks of the Snowshoe Group Assemblage and Semipelite – Amphibolite (SPA) Assemblage (correlative with the Mica Creek Assemblage). It is also interpreted to pre-date the regional fold structures of the Cariboo Mountains as it is folded by the large D₂ fold structures. Detailed field and petrographic work in the Hobson Lake area of the Cariboo Mountains has found that the trace of the PVT is different than presently interpreted. This raises the question of whether the PVT is in fact as significant a structure as implied by its map pattern.

The type location for the PVT occurs near Wells, British Columbia, on the westernmost edge of the Cariboo Mountains. The fault places Neoproterozoic Windermere Supergroup rocks on top of Paleozoic Snowshoe Group rocks (Struik, 1986). The Sir Arthur Meighen shear zone, found in the eastern Cariboo Mountains, is found to have Windermere Supergroup stratigraphy in its hangingwall and rocks of the SPA Assemblage in its hangingwall, and is interpreted to be correlative with the PVT (Murphy, 1999). The origin and age of the SPA Assemblage is currently unresolved but there are two possible interpretations. The first is that the SPA Assemblage sits stratigraphically below the Windermere Supergroup and is Proterozoic in age (Pell and Simony, 1987). An alternate hypothesis is that the SPA Assemblage is correlative with the Paleozoic Snowshoe Group and is therefore younger than the Windermere Supergroup (Murphy, 1999). The latter interpretation supports the Sir Arthur Meighen shear zone being correlative with the PVT and that the contact between the SPA Assemblage and Windermere Supergroup is a structural, not stratigraphic boundary. This interpretation would also imply that the PVT must occur throughout the Cariboo, Monashee and Selkirk Mountains of the southern Canadian Cordillera.

A major shear zone has been traced nearly continuously for approximately 40 km from Hobson Lake to east of Angus Horne Lake in the central Cariboo Mountains. The shear zone, herein referred to as the Hobson Lake shear zone, correlates locally with the trace of PVT / Sir Arthur Meighen shear zone. The Hobson Lake shear zone has a west-northwest trend and a dip between 45-60° towards the north. It is found to post-date the

regional metamorphism and cross-cut the regional D₂ fold structures. -Along its entire mapped length, the Hobson Lake shear zone is found to juxtapose rocks of the middle Windermere Supergroup (Isaac Formation) in the hangingwall against a footwall of lower Kaza Group rocks, which belong to the lowermost unit of the Windermere Supergroup. There is also a distinct break in the metamorphic grade across the shear zone that is most apparent at Hobson Lake where rocks of the Isaac Formation contain muscovite-chlorite-pyrite-siderite and locally chloritoid and the lower Kaza Group rocks in the footwall contain muscovite-biotite-garnet-staurolite and kyanite. The stratigraphic and metamorphic contrast across the shear zone indicates a down to the northeast normal sense of throw of 2-4 kilometers and fabrics in the shear zone also suggest a component of dextral-oblique motion. On the footwall side of the Hobson Lake shear zone the lower Kaza Group rocks lie on rocks of the SPA Assemblage and that contact would therefore mark the trace of the PVT. Examination of this contact on well exposed transects suggests a gradational, interbedded contact with no evidence of deformation. This implies that the position of the Pleasant Valley Thrust, from Hobson Lake eastward, is incorrect and supports the idea of Pell and Simony (1987) that the SPA Assemblage is stratigraphically below the lower Windermere Supergroup rocks.

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MULTI-DISCIPLINARY CORRELATION OF MID-CRETACEOUS VOLCANIC PACKAGES ACROSS THE INSULAR/INTERMONTANE SUPERTERRANE BOUNDARY: IMPLICATIONS FOR THE BAJA B.C. HYPOTHESIS

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The Baja B.C. controversy emanates from two conflicting models regarding the paleo-positions of the Insular and Intermontane Superterranes in the mid-Cretaceous. Previous paleomagnetic studies indicate the Insular Superterrane and associated entities have undergone ~2000 km northward translation relative to the Intermontane Superterrane between ~90-50 Ma. Conversely, geologic evidence supports <500 km of aggregate displacement between the superterranes. Both models cannot be correct. Analysis of a mid-Cretaceous volcanic package linked to eastern Insular Superterrane in the Empire Valley-Churn Creek area, south-central British Columbia tests the displacement models. We correlate the Empire Valley-Churn Creek package (EV-CC) to the Spences Bridge Group (SBG), which overlies Intermontane Superterrane based on six independent lines of evidence (Haskin, 2000):

1. Fault-restored geographic proximity: restoration of 110-140 km of dextral motion on the Eocene Fraser Fault places the two volcanic packages roughly along strike in the mid-Cretaceous.
2. Similar lithostratigraphy: The EV-CC volcanic succession is divisible into two distinct lithologic units similar to the SBG Pimainus and Spius Formations. Both have a lower unit dominated by volcanoclastic and pyroclastic rocks with intercalated mafic to felsic volcanic rocks, and an upper unit composed of mafic to intermediate volcanic rocks with minor pyroclastic rocks.
3. Comparable mineralogy: The upper unit of the EV-CC and the SBG packages both contain two distinct textural and mineralogic classes. Assemblage 1 (EV-CC) and 'Spius-type' (SBG) mineralogies are characterized by iddingsite after olivine pseudomorphs. Assemblage 2 (EV-CC) and 'Pimainus-type' mineralogies contain plagioclase, clinopyroxene, and orthopyroxene glomerocrysts.
4. Indistinguishable age: A U-Pb zircon date of 104.1 ± 0.3 Ma from the EV-CC lower unit matches a 104.5 ± 0.3 Ma date from the Pimainus Formation of the SBG.
5. Correlative geochemistry: both packages are subalkalic, calc-alkaline to transitional tholeiitic subduction-related arc packages with significant overlap in incompatible trace elements and overlapping and colinear trends on rare earth element plots.
6. Similar paleomagnetic signatures: Magnetic carriers in both packages are dominantly single domain magnetite and hematite. Magnetic behavior is characterized by well-behaved and well-defined single component magnetizations. Paleomagnetic poles are far sided and rotated clockwise with respect to the mid-Cretaceous North American pole and yield statistically indistinguishable displacement estimates (EV-CC 700 ± 600 km and SBG 1100 ± 600 km).

The correlation of the EV-CC succession with the SBG has major implications for large-scale displacement models. Shallow inclinations indicate the EV-CC/SBG package was ~700 to 1200 km south of present position at 104 Ma. The EV-CC succession is unconformably overlain by rocks correlated to the Albian-Santonian Silverquick conglomerate/Powell Creek volcanics of the Insular Superterrane, which paleomagnetic studies suggest was deposited ~3000 km south (Mahoney et al., 1996, 1999; Danielson et al. 1997; Riesterer et al in press; and Enkin et al., 2000). This stratigraphic relationship requires at least 2000 km of extremely rapid, sinistral, late Albian latitudinal translation followed by ~3000 km of Late Cretaceous to Tertiary northward translation of the Insular and Intermontane Superterranes. This model is difficult to reconcile with published plate motions and velocities of Engebretson et al. 1985 as it requires sinistral motions in a regime where dextral strike slip is more likely and necessitates velocities up to 19 cm/year. This study demonstrates the difficulty in resolving paleomagnetic and geologic data and is a critical step in refining our understanding of Cordilleran tectonics.

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Regional Terrane Relationships along the Superterrane Boundary: Implications for Large Scale Translation

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Multidisciplinary investigations into geologic relationships between terranes along the Insular/Intermontane superterrane boundary (Wrangellia, Methow, Cadwallader, Bridge River, Quesnellia, Bridge River terranes) strongly suggest regional terrane linkages during Cretaceous and Tertiary time.

Stratigraphic, geochemical, geochronologic, and paleomagnetic data indicate that volcanic strata of the Spences Bridge Group forms a regional overlap assemblage linking Quesnellia, Cache Creek and Stikinia(?). Haskin et al (2000, this volume) demonstrate that Spences Bridge Group strata extend across the Fraser fault system into the southern Chilcotin Plateau, where they are overlain by the Silverquick/Powell Creek succession.

Stratigraphic, structural, geochronologic, geochemical, palynological and paleomagnetic data demonstrate that the Silverquick/Powell Creek succession forms a regional overlap assemblage linking Wrangellia, Cadwallader, Bridge River, Methow and Quesnellia together by Albian-Santonian time. The Silverquick/Powell Creek succession was deposited in a series of piggyback basins developed adjacent to the Insular/Intermontane superterrane boundary. These rocks unconformably overlie Spences Bridge Group strata in the southern Chilcotin Plateau, definitively linking terranes of the Insular superterrane with those of the Intermontane superterrane by at least 85 Ma.

Additional constraints on terrane linkages are provided by geologic relations adjacent to the Pasayten fault in the southern Methow terrane, where Maastrichtian (ca. 70 Ma) Pipestone Canyon Formation unconformably overlies both the southern Methow terrane and Quesnellian basement rocks. This relationship negates substantial translation between the Insular and Intermontane superterranes after 70 Ma.

These linkages have significant implications for models invoking large-scale translation between the Insular and Intermontane superterranes between 90-50 Ma (the Baja BC hypothesis). Paleomagnetic data from the Spences Bridge Group indicate the package was ~700 to 1200 km south of its current position at 104 Ma. Paleomagnetic data from the Silverquick/Powell Creek succession indicates these strata were ~3000 km south of their current position between 100-85 Ma. The unconformable contact between the Spences Bridge Group and overlying Silverquick/Powell Creek succession requires complex and geologically unreasonable plate motions. Paleomagnetic data requires that the Albian Spences Bridge Group was deposited 1100 km to the south, latitude ~41°N, near the northern end of the Sierra Nevada system, at 100-105 Ma (Late Albian). This rock package must have been subsequently translated sinistrally approximately 1800 km south, to approximately the latitude of central Mexico (~27°N), prior to the deposition of the Albian-Santonian Silverquick/Powell Creek succession, which is discordant with respect to cratonic North America by 3000+/- 500 km, at ~86-92 Ma. The combined Spences Bridge/Silverquick package must then have been dextrally translated northward 1800 km, to be in its present position by Eocene time. Regional considerations, such as the paleomagnetic signature of the Carmacks Volcanic Group (Wynne et al., 1998) or the demonstrated stratigraphic linkage between the Intermontane and Insular superterranes at 70 Ma (Peterson et al., 1997; Peterson, 1999), requires a two-stage northward transport, with the Silverquick/Powell Creek succession being at ~41°N at 70 Ma, then in its current position (~52°N) by Eocene time.

The required tectonic scenario is implausible for two reasons: 1) ~1800 km of sinistral translation is required during the Late Albian (post 100, pre 92 Ma), a time of orthogonal convergence between the Farallon and North American plates (Kelley and Engebretson, 1994); 2) the rates of translation (~23 cm/yr from the south, 100-92 Ma; ~19 cm/yr north, 86-70 Ma) are far in excess of those proposed for Late Cretaceous relative plate motions (Kelley and Engebretson, 1994).

Rather than resolving the conflict, this investigation has sharpened the conundrum that exists between geologic and paleomagnetic estimates of latitudinal terrane motion. It is clear, however, that geologic relationships documented in the southern Canadian Cordillera necessitate revision of currently proposed models of large-scale terrane translation.

Spatial coincidence of the Kula-Farallon slab window with trench-distal volcanism of the Eocene Magmatic Belt in the southern Cordillera

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Introduction

Variations in composition and spatial distribution of the Early to Middle Eocene Magmatic Belt (Figure 1), particularly between its northern and southern components, must be accountable under a single tectonic umbrella; the issue of magma genesis must either lie within the realm of arc (subduction) processes (Armstrong, 1988), or not (Morris *et al.*, 2000). In its northernmost segment, (Alaska to northern BC), the belt's morphology, structural style and composition are typical of known arcs. In its central portion (between Lat. 48°N and Lat.55°N), voluminous volcanism occurs at a more trench distal position, with an extensional style of emplacement and often with mixed arc/non-arc geochemical affinity. These characteristics have been accommodated under an arc model by invoking a slightly flatter dip on the subducting slab in concert with back-arc spreading throughout this region (Ewing, 1981b; Grainger, 2000) or by incorporating the effects of the Kula-Farallon slab window (Thorkelson and Taylor, 1989). For the majority of the successions in the United States (south of Lat. 48°N), lithospheric decompression is the favoured model for magma genesis, on the grounds of detailed geochemical, isotopic and petrological studies, and also because of the improbability of producing arc magmatism under a flatly subducting slab, as would be required to accommodate such trench-distal centers as the Montana alkaline province and the Absaroka volcanic field (Düdás, 1991; Norman & Mertzman, 1991; Morris *et al.*, 2000). This leaves us with an enigmatic model for magma genesis for a set of grossly similar volcanic

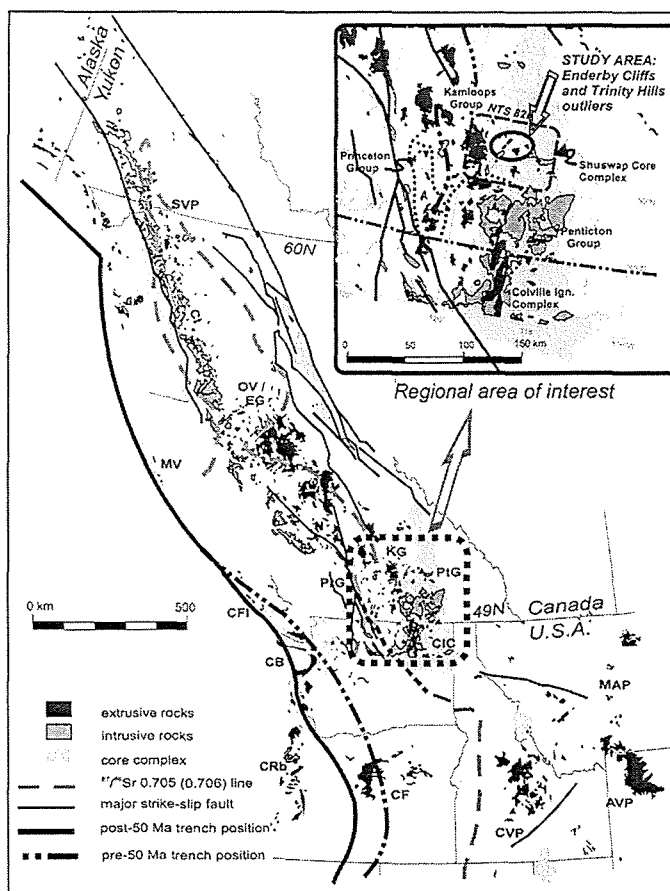


Figure 1 Early to Middle Eocene Magmatic Belt and related geologic features. Modified after Wheeler and McFeely (1991) and Burchfiel (1993). Initial $^{87/86}\text{Sr}$ line after Armstrong (1991), after Ghosh (1995) in Southern BC, and 0.706 line after Carlson *et al.* (1991) in the United States. Paleo-trench positions after Dickinson (1991). Magmatic centers include forearc, arc and foreland environments: SVP Sloko volcanic province; CI Coast Intrusives; OV/EG Ootsa and Endako Groups; KG Kamloops Group; PrG Princeton Group; PtG Penticton Group; CIC Colville Igneous Complex; CF Clarno Formation; CVF Challis volcanic field; AVF Absaroka volcanic field; MAP Montana alkaline province; CRb Coast Range basalts; CB Crescent Basalts; CF Catface intrusions; MV Massett Volcanics.

successions at these various parts of the belt, with arc magmatism occurring above the Kula Plate north of Lat. 48°N, and a lack of support for arc magmatism above the Farallon Plate south thereof.

Subduction of the Kula-Farallon spreading ridge from late Cretaceous to Middle Eocene times (approx. 83 to 40? Ma) beneath western North America (Engelbreton *et al.*, 1985; Stock and Molnar, 1988; Atwater, 1989) would have produced a slab window (Dickinson and Snyder, 1979; Thorkelson, 1996) under the northwestern United States or southern British Columbia (Thorkelson and Taylor, 1989). Recognition of the effect that the Kula-Farallon slab window would have on the overriding plate is critical to understanding the nature of this belt. Slab windows permit hotter, drier asthenospheric material to upwell and displace or interact with the overlying mantle wedge, which would have been previously metasomatized by the downgoing slab prior to its migration away from that locus. Lateral mantle flow may also occur. These perturbances in the mantle may produce anomalous magmatism in the overriding plate (Johnstone and Thorkelson, 1997). Diverging tractions between the downgoing slabs and the continental margin on opposite sides of the ridge-trench junction may produce extension in the overriding plate from fore-arc to back-arc. A slab window environment, then, induces a stress regime similar to that of back-arc extension, which may lead to mantle or crustal anatexis resulting in mafic to felsic magmatism, respectively.

This study addresses the paleo-position of the Kula-Farallon slab window with respect to trench distal volcanic successions of this magmatic belt in an attempt to link the on-land geological record with currently accepted plate motion models. The paleo-geographic reconstruction presented herein updates that of Thorkelson and Taylor (1989) to incorporate the effects on ridge-transform geometry resulting from plate motion changes, to incorporate revisions to ages of the magnetic anomalies (Berggren *et al.*, 1995) which affect the calculation of relative motion vectors from rotation poles, and to incorporate minor modifications to the geometry of the paleo-trench (Dickinson, 1991).

Method

The position of the window was calculated at 1 Ma increments from the Kula-Farallon breakup, here modelled as at end of Chron 34 (Engelbreton *et al.*, 1985), equivalent to around 83 Ma (Gradstein *et al.*, 1995). Three break-up geometries were modelled, within the region of uncertainty suggested by Engelbreton *et al.* (1985), those being a single-break, or 'northern' option, a single transform, and a double transform option (Figure 2). The more southerly options are based on the notion that spreading would occur along pre-existing transform segments of the

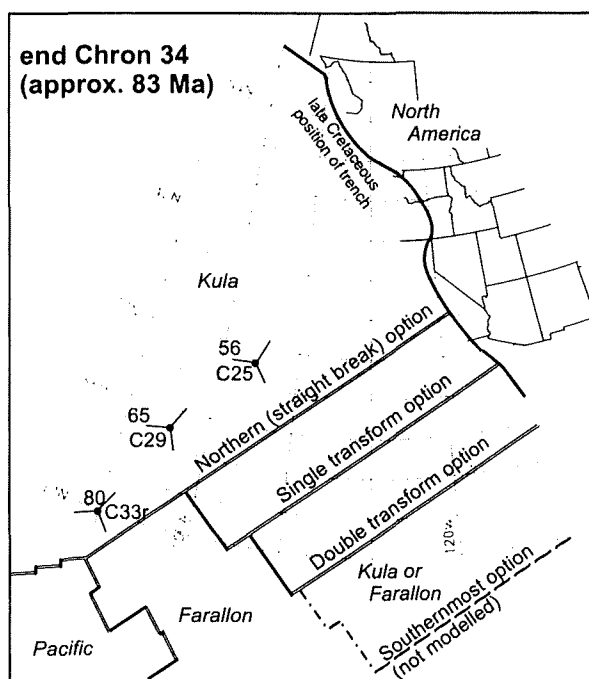


Figure 2 Possible geometry of incipient Kula-Farallon break within the area of uncertainty set out by Engelbreton *et al.* (1985). Ridge configurations are shown in double solid lines. Paleo-trench position after Dickinson (1991).

Pacific-Farallon, and so the longitudinal position of this ridge-transform segment may not be adjusted. Relative motion vectors for the Kula and Farallon plates were recalculated using the chron ages of Berggren *et al.* (1995) and Gradstein *et al.* (1995) from the rotation poles of Engebretson *et al.* (1985) for the six stages reported therein, those being end of Chron 34 to 33 (83 to 74 Ma), Chron 33 to 30 (74 to 66 Ma), Chron 30 to 28 (66 to 61 Ma), Chron 28 to 25 (61 to 56 Ma), Chron 25 to 21 (56 to 48 Ma), and Chron 21 to 18 (48 to 39 Ma).

At each stage boundary, the new spreading ridge must accommodate itself between the edges of the departing plates, which have a geometry defined by the previously stable spreading ridge. This forces a ridge-transform ratio defined by the difference between angles of the stable ridges for each of the two stages; a new ridge would develop either by ridge propagation or ridge rotation (Hey *et al.*, 1988). Regardless of mechanism, the ridge-transform geometry will continue to adjust itself for a period of around 3 Ma (cf. Kula-Pacific, duration of Chron 24, Lonsdale (1988), and Pacific-Farallon, Caress *et al.*, 1988). The ridge-transform boundaries were thus maximized within the inter-plate space at 3 Ma after the stage boundary, then fixed for the duration of the stage. Ridge-transform boundaries have significant effect on both the shape and migration rate of a slab window (Thorkelson, 1996); thus the inclusion of the five stage boundaries encountered during the lifetime of Kula-Farallon spreading become a fundamental element in the reconstruction of the window's geometry and position. The evolution of these ridge-transform boundaries and their relative ratios are a function of plate trajectories. The only adjustable parameter, then, is that pertaining to the geometry and position of the initial Kula-Farallon break (Figure 2).

Discussion

The reconstructions presented demonstrate several significant aspects of the slab window's position with respect to the overriding (North American) plate. Firstly, regardless of starting geometry, transform faults produced at stage boundaries slow or stall the slab window's migration from about 76 to 56 Ma. Under the straight-break option, the window hovers beneath the Idaho Batholith, while the transformed options place it just north of (single transform option) or directly beneath (double transform option) the foreland alkaline magmatic center in Colorado. Either of these late Cretaceous to Paleocene positions is supportable.

Secondly, the portion of ridge which is the same under all options (from the Pacific-Kula-Farallon junction north-east to the first transform under the southern options, Figure 2) enters the trench by about 51 Ma at about Lat. 48°N under all options. This means that by 50 – 49 Ma, the window is necessarily positioned beneath the Penticton and Kamloops Group volcanics in the Intermontane Belt (Figure 3). Differences in the window's position between the straight and transformed models remain however in the more trench-distal portions of the Cordillera. The transformed options see the window swing south such that it encompasses the position of both Challis and Absaroka volcanic fields for the duration of magmatism (51 – 44 Ma). The straight-break (northerly) option sees the window swing to the north, encompassing no particular magmatic field. The transformed option is thus preferred, as it permits explanation of much of the trench-distal volcanism in the United States which has chemical similarity and is coeval to that of the Kamloops Group in British Columbia.

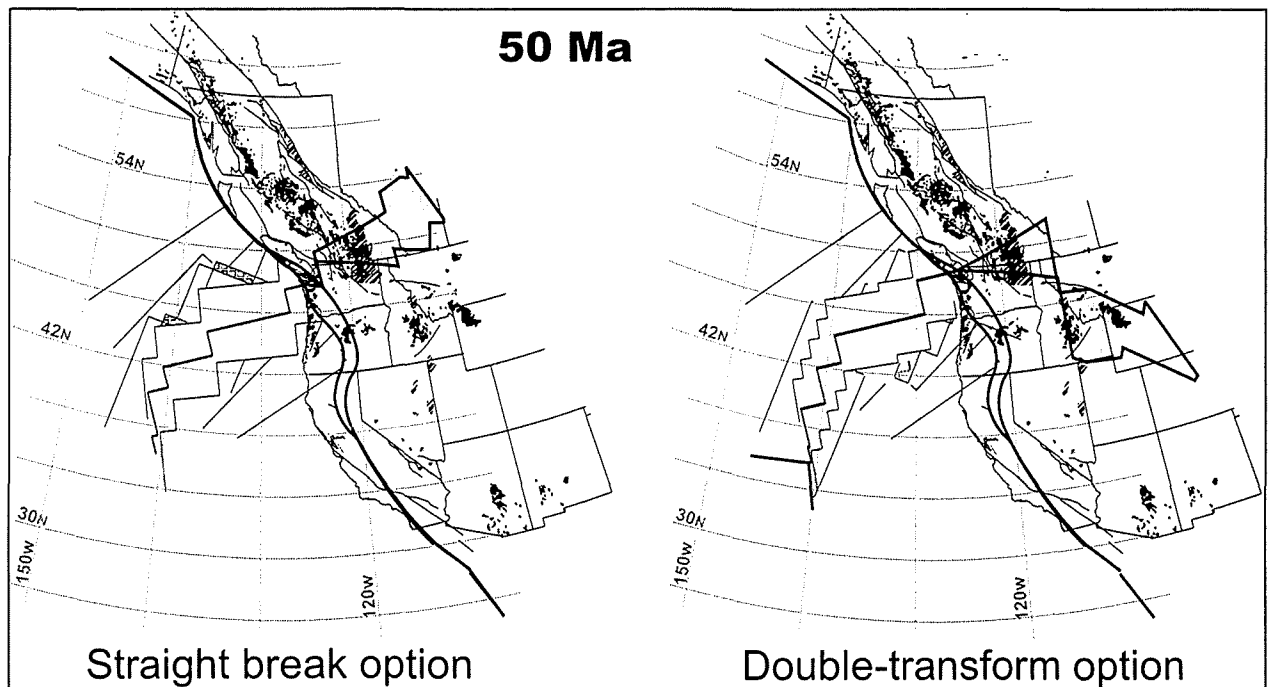


Figure 3 Reconstructed positions of the Kula-Farallon slab window at 50 Ma. The match to on-land geology shown by this preliminary double-transform reconstruction is reasonably good. Refinements to the Kula-Farallon incipient break geometry and removal of the effects of dextral strike-slip on the Fraser Fault system may improve this correlation.

Thirdly, the northernmost extent of the window encompasses the majority of the voluminous, extension related volcanism in central British Columbia. The Endako and Ootsa Groups of the Nechako Plateau however are not encompassed by the paleo-window at the correct times (~ 50 to 45 Ma). Further work examining the effect of post-Eocene dextral strike-slip on the Fraser Fault, on the order of around 100 km (Umhoefer and Schiarizza, 1996) could however, see these successions positioned in the vicinity of the window. This would require that the fault system was continuous with the Pinchi Fault system to the north, beneath the Miocene volcanic cover. Since the double-transform option correlates well with the spatial extent of volcanism from southern B.C. to Wyoming, a model encompassing the full belt (Nechako region) by way of dextral strike-slip restoration would offer a singular model to explain inland volcanism of the whole belt without negating the findings of a large degree of effort on the part of many workers who have examined successions of this belt.

Lastly, an interesting effect of the change in stage boundaries of the transformed options is that the long transform becomes compressive in nature. This area of plate which sees approx. 10 Ma of compression comes in contact with North America in the vicinity of western Oregon to Vancouver Island over the period of 54 to 50 Ma. The timing of this event coincides with the expected accretion of oceanic thickenings such as Siletzia. If this compression resulted in the initiation of subduction prior to removal of compression at 56 Ma, it would be in agreement with geophysical studies beneath those areas which have been interpreted as truncated, subducting slabs under southern Vancouver Island (Clowes *et al.*, 1998) or areas of thickened oceanic crust under the Washington and Oregon coast (Parsons *et al.*, 1999).

Conclusions:

This updated slab window reconstruction holds promise to answer outstanding issues regarding the nature and origin of the Eocene magmatic belt. These preliminary reconstructions show a good correlation to

the overall shape and timing of the belt. Further work includes i) refinement to the initial break geometry to enhance coincidence with the spatial distribution of the belt, ii) examination of the effects of dextral strike slip along the Fraser and Pinchi Fault systems to accommodate Eocene volcanism in the Nechako Plateau under this scenario, and iii) correlation of the window margins to contoured trace element geochemistry for the entire magmatic belt. This work forms part of a larger project detailing stratigraphy and geochemistry of the Eastern Kamloops Group.

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Geologic Setting of Paleozoic Strata and Mineral Occurrences in the Mount Tod-Adams Lake region, south central British Columbia

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The correlation and extent of pericratonic strata deposited on the western distal margin of ancient North America in south central British Columbia is problematic. Within the Thompson-Okanagan-Shuswap and Adams Lake regions, strata included within the pericratonic Kootenay Terrane are poorly exposed, obscured by intense deformation, and mapped only at reconnaissance scale. As a result, the boundary between pericratonic rocks of the Kootenay Terrane and rocks of the allochthonous Quesnel Terrane is a subject of debate. This boundary has been interpreted as a large scale, complex thrust zone. Alternatively, Mesozoic volcanic and volcanoclastic rocks of Quesnel Terrane may stratigraphically overlie pericratonic strata of the Kootenay Terrane.

Rocks of the Eagle Bay Assemblage, and the Sicamous, Tsalkom and Silver Creek formations are included within the pericratonic Kootenay Terrane in the Mount Tod – Adams Lake – Shuswap region. These rocks are of significant economic interest since they are host to several polymetallic massive sulphide deposits. For example, altered calcareous rocks of the Sicamous Formation and metavolcanic rocks comprising Unit EBK of the Eagle Bay Assemblage host the Steep skarn and Serpent massive sulphide deposits near Adams Lake. The Steep showing includes massive pyrrhotite with associated trace amounts of gold, and disseminated to massive pyrite, chalcopyrite, and magnetite. Extensive quartz-sericite-fuchsite-pyrite alteration and affiliated pyrite, chalcopyrite and pyrrhotite mineralization of metasedimentary and metavolcanic rocks of the Silver Creek Formation occurs adjacent to a quartz vein stockwork system within the Chase Silica quarry, south of McGillivray Lake.

Volcanic and volcanoclastic rocks of the Late Triassic Nicola Group and arc-related rocks of the Devonian to Late Permian Harper Ranch Group characterize Quesnel Terrane in the Thompson – Okanagan – Shuswap region. A sequence of carbonaceous phyllite and slate that is host to previously unrecognized mineralization separates the Nicola Group from the Silver Creek Formation in the Mount Tod area. These rocks may correlate with an assemblage of schist, marble, and quartzite separating the Silver Creek Formation from Triassic rocks correlative with the Nicola Group in the Vernon area. This assemblage is interpreted as a stratigraphic transition zone, implying that rocks assigned to Quesnel Terrane are stratigraphically linked with pericratonic rocks of the Kootenay Terrane. Alternatively, these rocks may be metamorphosed sediments at the base of the Nicola Group, or a Sicamous Formation equivalent. Establishing the stratigraphic and structural relationships of this sequence is important for understanding the tectonic setting of the ancient distal margin of North America.

Devonian-Mississippian VMS deposits of western Canada as an expression of continent margin tectonics

S. Paradis, and

S.L. Bailey, G.D. Bradshaw, K. Daughtry, T. Hoy, D.C. Murphy, J.K. Mortensen, J.L. Nelson, J. Peter, S.J. Piercey

Abstract

The metallogenic component of the Ancient Pacific Margin NATMAP project is a comprehensive analysis of the tectono-stratigraphic settings of syngenetic deposits within the Cordillera-long belt of complexly deformed and metamorphosed rocks of pericratonic character known as Kootenay and Yukon-Tanana terranes.

Metallogenic history of the western Canadian-Alaskan Cordillera during the Devonian-Mississippian time suggests the coeval development of hydrothermal activity from south to north along the outer plate margin of ancestral North America. This led to the generation of primitive to continental arc-related volcanogenic massive sulfide (VMS) and hybrid VMS-sedimentary exhalative sulfide (SEDEX) deposits in complexly deformed and metamorphosed metasedimentary and metavolcanic rocks of these terranes.

This paper focuses on Devonian-Mississippian syngenetic sulphide deposits of the Kootenay and Yukon-Tanana terranes, and addresses their similitudes in terms of timing and tectonic environments.

The Devonian-Mississippian rocks of the Eagle Bay Assemblage of the Kootenay Terrane host several types of syngenetic sulphide deposits: 1) Volcanogenic disseminated sulphides (Cu, Au, Ag, Mo) in bimodal calcalkaline metavolcanic rocks and metasedimentary rocks (e.g., Harper Creek, Lydia, VM, VAV); 2) Kuroko-type volcanogenic sulphides (Au, Ag, Zn, Pb, Cu, Ba) in alkaline felsic to mafic metavolcanic rocks (e.g., Homestake, Rea, Beca, Harper, Joe); and 3) Besshi-type volcanogenic sulphides (Cu, Zn, Ag, Au) in metasedimentary rocks (e.g., Mount Armour, Fortuna).

The Devonian-Mississippian rocks of the Finlayson district, Yukon-Tanana terrane comprises: 1) Besshi-type deposits (e.g., Fyre Lake) in mafic volcanic rocks of boninitic, low-Ti tholeiitic to calc-alkalic affinities; and 2) Kuroko-style and hybrid Kuroko-SEDEX-style deposits (e.g., Wolverine, Kudz Ze Kayah, GP4F) in felsic volcanic rocks of calc-alkalic affinity.

Devonian-Mississippian time is marked by tectonism and profound changes in the setting of the western edge of the ancestral North America. It involved continental arc magmatism and long-lived regional extension, which generated back-arc and intra-arc spreading and exhalative activity.

MESOSCALE DEFORMATION FABRICS IN A FAULT-PROPAGATION FOLD, SOUTHERN LIVINGSTONE RANGE, ALBERTA, PRELIMINARY OBSERVATIONS

Michael A. Cooley, John M. Dixon, Raymond A. Price, & T. Kurtis Kyser

ABSTRACT

Preliminary investigation of a fault-propagation fold exposed at Green Creek in the southern Livingstone Range revealed a series of mesoscale structures that formed during the development of the Green Creek anticline. The oldest structures are duplexes that developed at specific stratigraphic horizons in the hinge region of the fold. Subsequent deformation involved the formation of backthrusts in the steeply dipping forelimb, which may have been concurrent with tightening of the fold by bedding-parallel flexural slip. The hinge region duplexes and forelimb backthrusts are cut by the Green Creek thrust which propagated through the forelimb of the fold just below the hinge region, taking advantage of pre-existing fractures. The latest episode of deformation is represented by locally developed steeply-dipping small scale faults which strike ENE and cut the anticline at a high angle. A few fault planes exhibit subhorizontal or gently plunging slickensides and rarely exhibit an apparent dextral offset on the order of several metres.

Future work will include a comprehensive examination of faults, fractures and veins that formed during the evolution of the Green Creek anticline to model the hydrodynamics of formation fluids. Geochemical analyses of vein minerals associated with each episode of deformation is proposed to determine the fluid chemistry evolution, which will provide information on temperature and depth during particular stages of deformation.

INTRODUCTION

The Livingstone Range anticlinorium of southwestern Alberta is situated above a major longitudinal hanging-wall ramp on the Livingstone thrust (Figure 1).

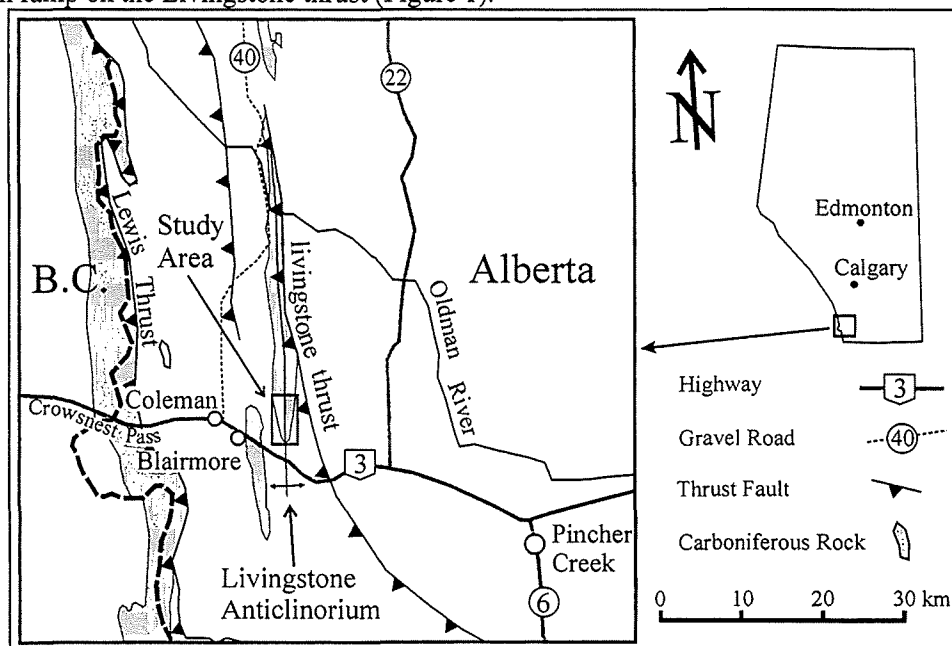


Figure 1 Map of southern Alberta showing the location of the study area at the southern end of the Livingstone Range. The Livingstone Range is an anticlinorium comprising Carboniferous and younger rocks in the hanging wall of the Livingstone Thrust.

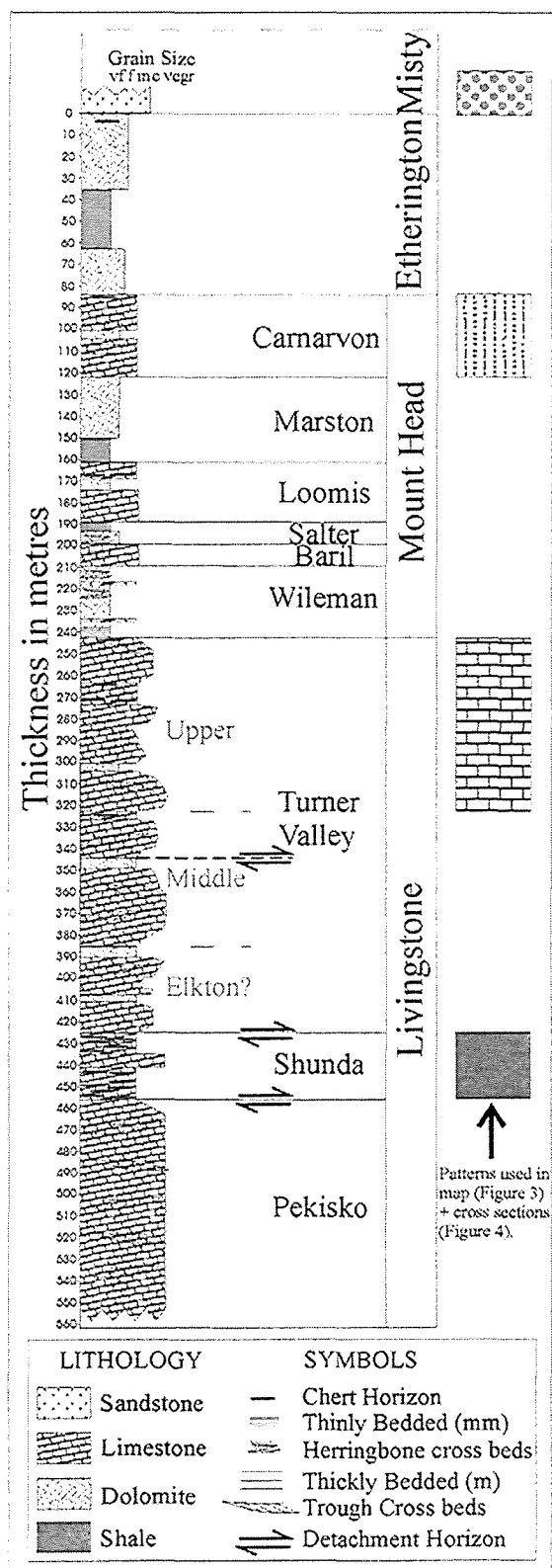


Figure 2. Stratigraphic column of rocks exposed in the southern Livingstone Range.

It involves an array of fault-propagation folds (Almendinger, 1998; Suppe and Medwedeff, 1990) associated with thrust faults that splay from the underlying Livingstone Thrust. The anticlinorium is analogous to petroleum reservoir structures that occur in the subsurface in other parts of the Rocky Mountain Foothills of southwestern Alberta that can only be studied by seismic imaging and deep drilling. However, unlike subsurface structures, the Livingstone Range anticlinorium has been exhumed and etched into relief by erosion, making it readily accessible for detailed study and analysis.

This project aims to elucidate the geometric, kinematic and dynamic relationships between displacements on the thrust faults and the development of the thrust-fault-related folds and fracture systems. The field project involves detailed structural mapping, structural analysis of fractures and other mesoscale tectonic fabric elements, and sampling of mineralized fractures for subsequent analysis in the laboratory. This project will study the associated geochemistry, hydrodynamics and geothermics of circulating formation fluids by means of isotopic analyses and fluid inclusion studies of vein minerals.

PREVIOUS WORK

Previous mapping in the area was conducted by Norris (1987) at a scale of 1:50000, (NTS map 1829A). The rocks that underlie the study area comprise mainly Carboniferous stratigraphy, including the uppermost Banff Formation, and the overlying Livingstone, Mount Head, Etherington, and Misty Formations. A stratigraphic column of the rocks mapped by this study in the southern part of Norris' area is presented in Figure 2.

SUMMER 2000 FIELD WORK

Field work in the summer of 2000 focused on mapping a fault-propagation fold in Carboniferous stratigraphy at the south end of the Livingstone Range, immediately north of Crowsnest Pass where the anticlinorium plunges abruptly to the south (see Figure 1). A preliminary geologic map of the study area is presented in Figure 3.

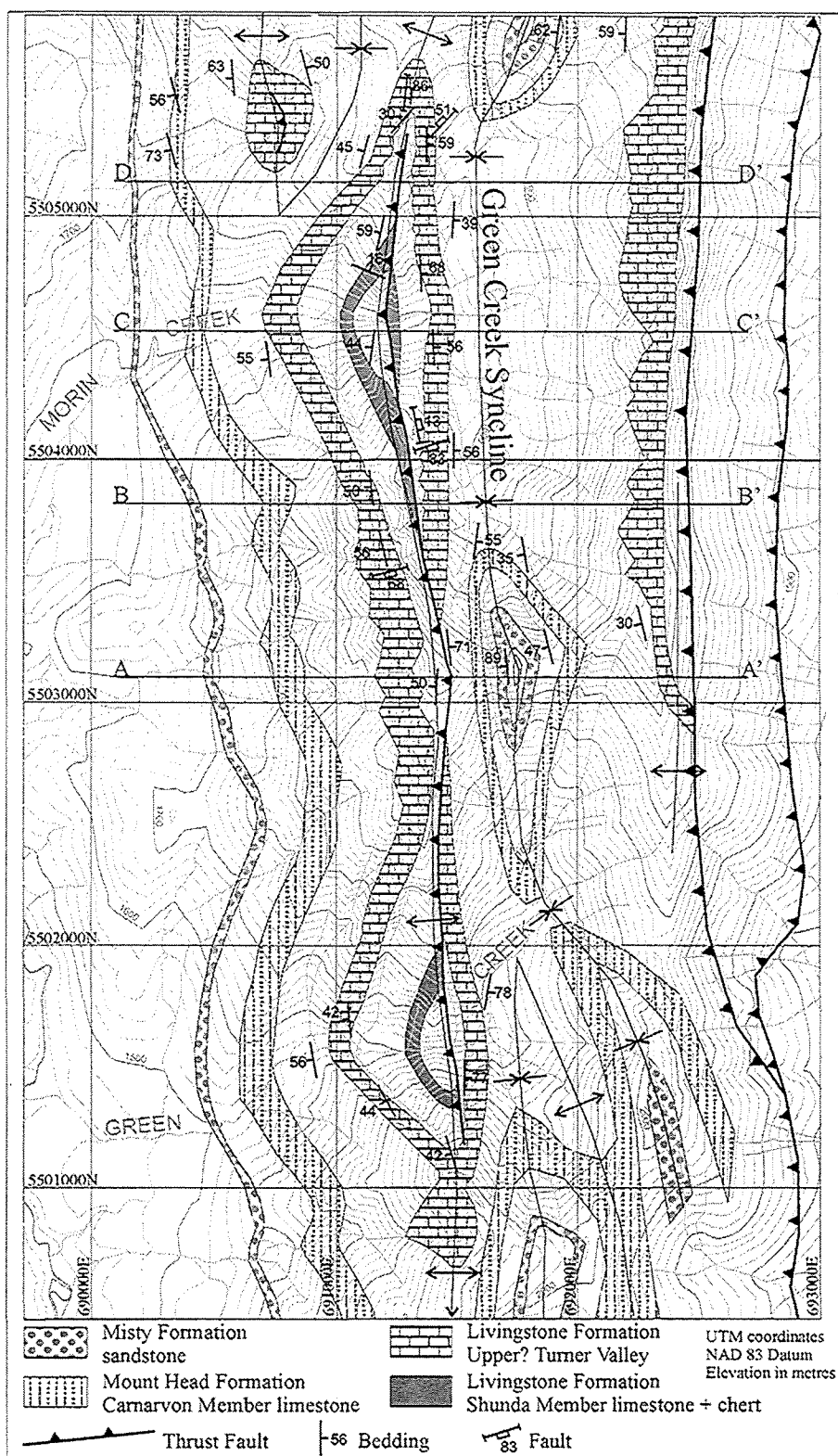


Figure 3. Geologic map of the field area. The Green Creek Anticline and Green Creek Thrust are at the centre of the map, west of the Green Creek Syncline.

The southernmost outcrops of Carboniferous rocks are well exposed in the Green Creek syncline (Norris, 1997), which is dissected by Morin Creek and Green Creek. West of this syncline lies the complimentary anticline, referred to as the Green Creek anticline by this study. The Green Creek fold pair was mapped at a scale of 1:10000 by this study. Mapping was facilitated by measuring six stratigraphic sections across the three limbs of the Green Creek fold pair. Attention was focused on the anticline to elucidate the architecture of the fault-propagation fold.

PRELIMINARY OBSERVATIONS

The Green Creek anticline plunges gently and has a surface strike extent of approximately 5 kilometres. An east-verging thrust fault occupies the axial surface of the anticline and has an approximate displacement of 140 metres at an axial culmination north of Green Creek and south of Morin Creek. To the north and south, displacement on the fault decreases until it dies out where the tip line of the fault occupies the core of the anticline.

A series of cross sections through the northern part of the fold are presented in Figure 4.

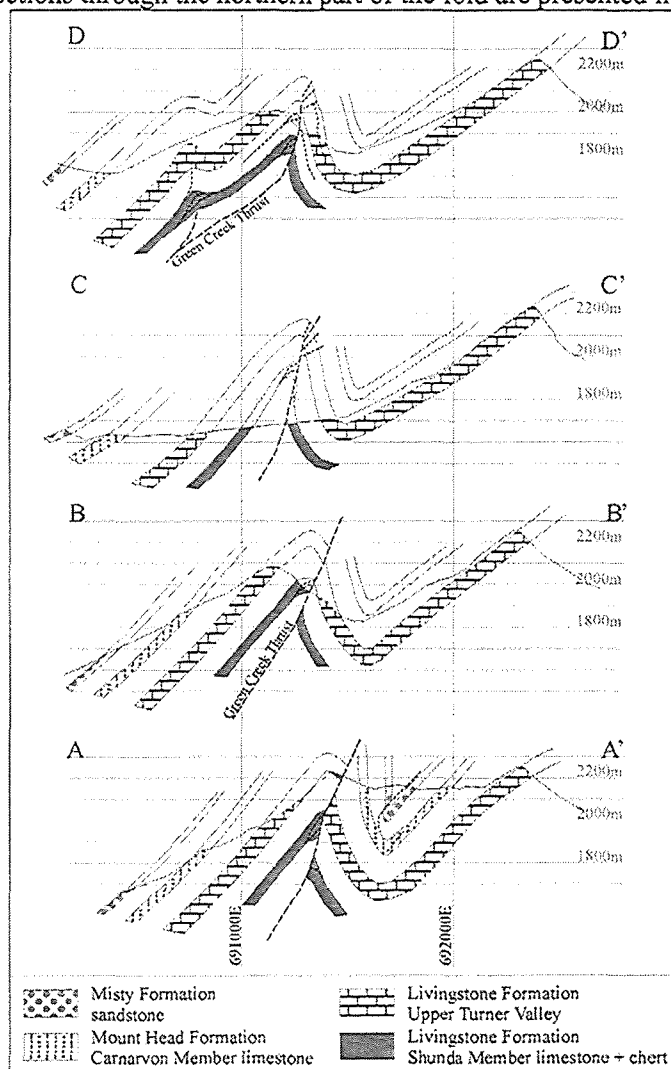


Figure 4. Cross sections A, B, C, and D through the northern part of the Green Creek fold pair.

The cross sections show how thrust displacement decreases to the north, and increases down the dip of the fault. The Green Creek anticline is asymmetric, with a steeply eastward-dipping forelimb and a shallowly dipping west limb. The core of the anticline is well exposed along strike, and is only present in the hanging wall of the Green Creek Thrust. In the west limb there are a few stratigraphic levels that are the location of preferred detachment horizons. These include the base of the Shunda Member, the base of the Elkton Member, and in the upper part of the Middle Member of the Turner Valley. In the field, these fault horizons are indicated by smooth fault planes parallel to bedding, by gullies where fault action has made the rocks more easily eroded, and by the presence of duplex structures in the core of the anticline (Figure 5).



Figure 5. View to the north at the core of the Green Creek Anticline exposed on the north side of Morin Creek. The hinge region of the fold contains a zone of duplexes that form an antiformal stack within the lower part of the Shunda Member.

Although several faulted horizons were identified in the west limb of the anticline, no ramps or thickening of the stratigraphy was observed.

The forelimb of the Green Creek anticline contains several small-scale back-thrusts, with displacements on the order of several metres, at various stratigraphic levels. One of these back thrusts is shown in Figure 6.

In the forelimb of the anticline near the core there are extension faults with an east-verging, thrust-sense of displacement. These extension faults appear to reactivate fractures that are generally at a high angle to bedding and strike roughly parallel to structural grain (Figure 7).

The extension faults offset the back thrusts, indicating that they formed later. The extension faults are depicted in cross section D-D' in Fig 4, and are represented by fault symbols on the map in Figure 3.

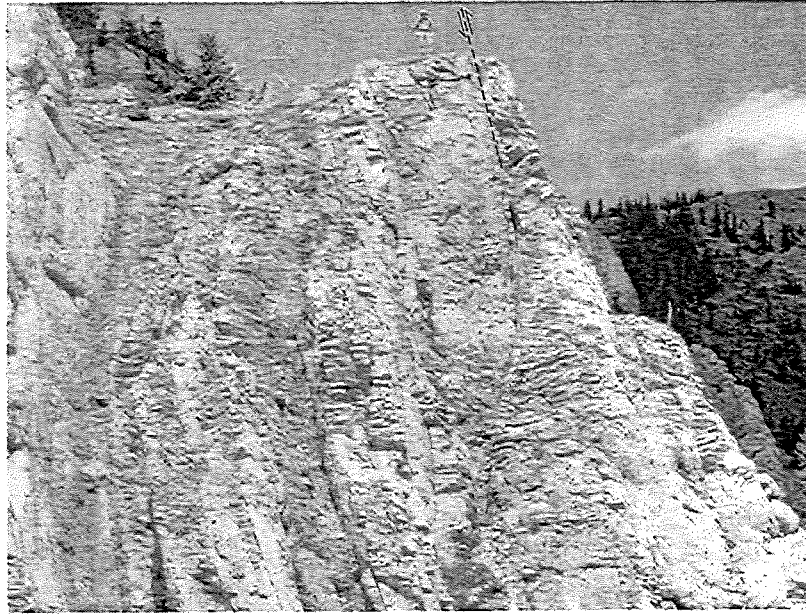


Figure 6. View looking north at a west-verging thrust fault (back thrust) in the forelimb of the Green Creek anticline, in the Upper Turner Valley Member.

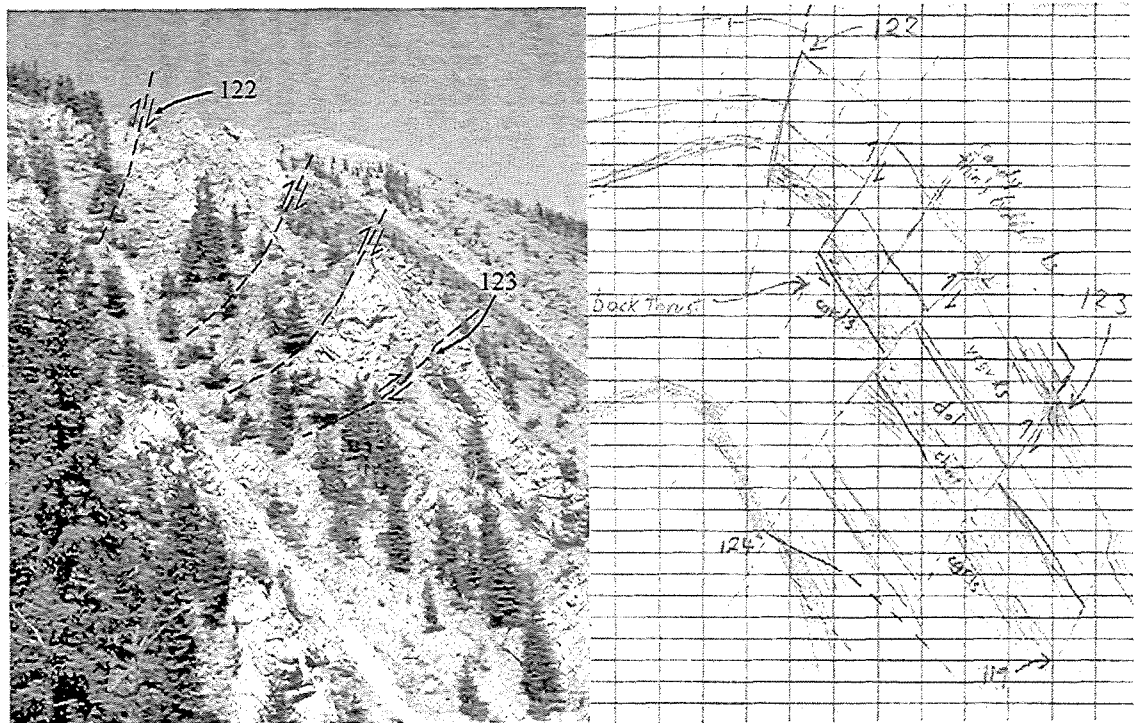


Figure 7. The photo on the left is a view to the north at extension faults exposed in the forelimb of the Green Creek anticline. The diagram to the right is a sketch from a field notebook showing the interpretation of the structures. The numbers refer to map stations.

In the forelimb and back limb near the axial culmination of the fold there are near-vertical small scale faults which strike ENE and cut the anticline at a high angle. Most have an apparent dextral offset on the order of several metres. These are represented by fault symbols immediately south of Morin Creek on Figure 3. These faults appear to cut the backthrusts in the forelimb, but their relation to the extension faults is uncertain.

A preliminary fracture analysis was undertaken across the anticline, with fracture measurements of orientation, trace length, aperture and density collected at several stations across the structure. Although a comprehensive examination of the data is not completed at this time, it was observed that fractures are more numerous and exhibit a greater variety of orientations in the footwall than in the hanging wall.

INTERPRETATION

From the observed mesoscale faults it is apparent that the Green Creek anticline formed from a series of deformation events, with different structures forming at particular times during its evolution. Preliminary observations suggest that the earliest mesoscale structures to form were the detachment horizons and the related duplexes in the hinge region of the fold. As the Large scale anticline developed, back thrusts formed in the steepening forelimb, and tightening of the fold was facilitated by bedding-parallel flexural slip. When the fault at the core became active it propagated through the forelimb of the anticline, taking advantage of available fractures with a structurally favorable orientation.

FUTURE WORK

Structural Analysis

An understanding of the mesoscale structures that are present along this structure should give insight into how this particular fault developed, and into how fault-propagation folds evolve in general. Further field work will entail additional detailed stratigraphic analysis in order to improve the resolution of the structural mapping. The field area will be expanded to the north to document the other fault-propagation folds and to determine how the folds overlap or interfinger each other along strike. The analysis of the geological model will be facilitated by 3D computer modeling with GoCad and ArcView computer programs.

Fluid Flow Dynamics; Modelling The Fluid History

Extensive sampling of vein minerals associated with each mesoscale structure will be undertaken to conduct fluid inclusion studies and to analyze the isotope geochemistry to characterize the fluids that were migrating through the structure at each phase of deformation. The geochemical results will be used to determine how the fluid system evolved during deformation, which should provide insight into the hydrodynamics of formation fluids, hydrocarbon maturation, migration and trapping. This project will attempt to model the fluid flow that may have occurred at particular times during evolution of the structure by determining which fractures and faults were open to flow at particular times during the evolution of the structure. The aperture (fracture separation) may be determined from the widths of veins if the fracture is filled with minerals, otherwise a fracture aperture must be assumed.

The results of this project may provide additional information about the eastward progression of fluids which chemically reset the paleomagnetism ahead of the deformation front as the deformation progressed from west to east (Enkin et. al, 1997).

ACKNOWLEDGMENTS

We gratefully acknowledge the financial support for this project by Sponsors of the Foothills Research Project and the Natural Sciences and Engineering Research Council of Canada. We also acknowledge Greg Cameron for his assistance in the field.

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Cretaceous sedimentation, burial, magmatism and metamorphism in the Leech River Complex – What does it all mean?

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The Leech River Complex (LRC), located on southern Vancouver Island, is a metasedimentary and metaigneous assemblage that is fault bounded between the Eocene Metchosin Igneous Complex, to the south, and Paleozoic-Jurassic rocks of Wrangellia to the north. Field and geochronologic investigations have revealed a previously unknown Cretaceous history for this assemblage. U-Pb dating of detrital zircon from metasedimentary rocks of the LRC indicates a varied source region for the protolith, based on discordant grains with $^{207}\text{Pb}/^{206}\text{Pb}$ ages of ca. 1.4 Ga, and grains with concordant ages of 185 Ma to 103 Ma. The youngest grain provides a maximum age for deposition of the LRC metasedimentary protoliths. This late Early Cretaceous age differs from previous interpretations of the depositional age using Rb-Sr isotopic systematics (Fairchild and Cowan 1982) as well as possible correlative relationships with rocks of the Pacific Rim Complex (Brandon 1989) and the Pandora Peak Unit (Rusmore and Cowan 1985), both constrained by fossil assemblages, all of which indicate a Jurassic depositional age. A titanite age of 88 Ma, interpreted as the crystallization age of the Jordan River metagranodiorite, records previously unrecognized Cretaceous magmatism in the LRC.

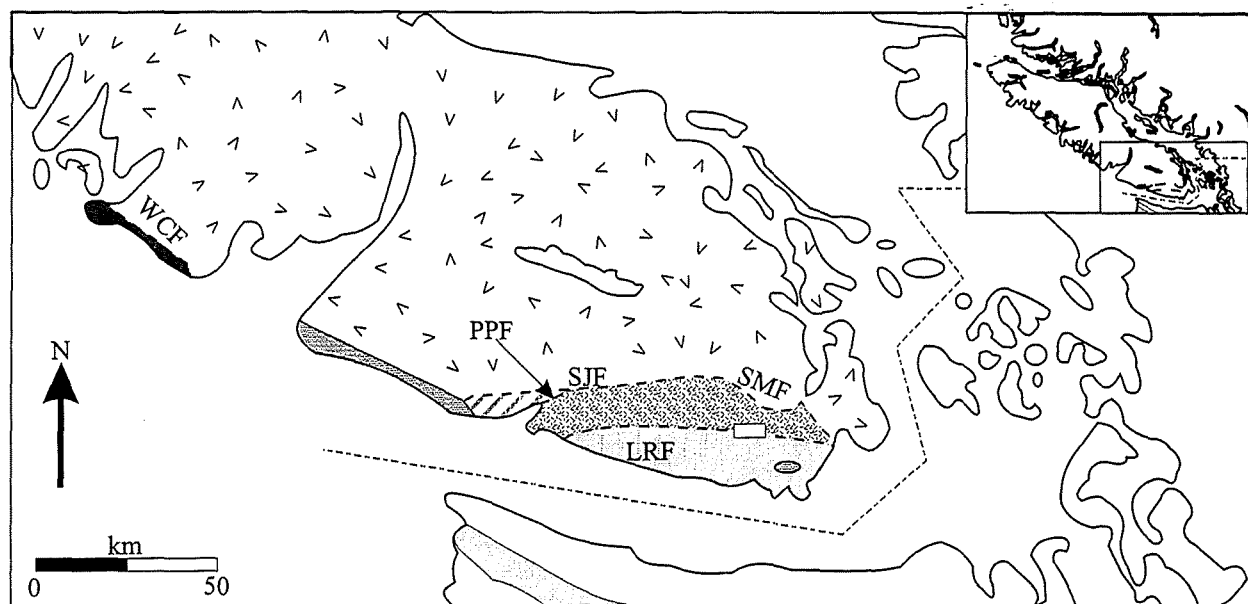
The Jordan River metagranodiorite, consisting of one stock and 5 - 10 related dykes, is slightly peraluminous, with an arc-like trace element signature, and an overall composition nearly identical to the host metasediments. Preliminary modeling suggests that melting or assimilation of the forearc sediments was important in its petrogenesis. The granodiorite records a period of magmatism soon after LRC protolith sediments were deposited. Amphibolite-facies metamorphism in the contact aureole and penetrative ductile deformation accompanied the magmatism. Regional greenschist facies metamorphism is also interpreted to partly overlap in time with magmatism. Contact metamorphic porphyroblasts range from concordant to discordant with respect to the regional fabric, indicating growth during and probably just after formation of this fabric; regional mineral assemblages are interpreted as syn-deformational.

The age determinations on the Jordan River metagranodiorite and the LRC metasediments indicate that these intrusions were emplaced into the LRC less than 20 m.y. after the sediments were deposited. Conditions in the metamorphic aureole around the Jordan River metagranodiorite indicate that the intrusion was emplaced at pressures of 2-3 kb. These lines of evidence indicate that after deposition of sedimentary protoliths the LRC was buried to a depth of ca. 10 km, intruded by the Jordan River metagranodiorite, deformed, and metamorphosed within a 20 m.y. time period.

Rapid burial of the LRC may be due in part to early Late Cretaceous west-directed thrusting along the Coast Mountains thrust system (Journeay and Friedman 1993). However, the presence of the Cretaceous peraluminous magmatism in the LRC, which was possibly close to the paleo-plate margin, requires special tectonic circumstances. Two alternative scenarios are envisioned for this Cretaceous magmatism. The first is that an oceanic spreading ridge interacted with this segment of the North American forearc during early Late Cretaceous time, a situation that has been proposed to explain the voluminous Eocene forearc magmatism in the LRC, elsewhere on Vancouver Island, and in southeast Alaska (Sisson and Pavlis 1993; Haeussler et al. 1995; Groome 2000; Groome et al. 2000). A second is through overthickening of the crust during thrusting along the Coast Mountains thrust system. However, given the normally low geothermal gradient of the forearc, the tectonic overthickening would have to have been quite significant to initiate partial melting. At present, not enough is known about the Cretaceous forearc geology of this region to draw any firm conclusions about the cause of this magmatism.

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Legend

Tertiary	Mesozoic	Paleozoic to Mesozoic
 Carmanah Group  Crescent Terrane (undivided)	 Pacific Rim Complex  Leech River Complex  Pandora Peak Unit	 Wrangellia (undivided)

Figure 1. Simplified tectonic assemblage map of southern Vancouver Island and adjacent portions of Washington State and mainland British Columbia. WCF=West Coast Fault, PPF=Pandora Peak Fault, SJF=San Juan Fault, SMF=Survey Mountain Fault, LRF=Leech River Fault. The study area for the project is outlined by the white box. After Wheeler and McFeeley 1991.

ABSTRACTS of
Poster
PRESENTATIONS

Eocene-Neogene Magmatism in the Nechako River area, central British Columbia.

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Eocene and Neogene magmatic episodes in central British Columbia were widespread, post-accretionary, and well inboard of the Tertiary Canadian Cordilleran continental margin. Collaborative studies in the northern Nechako map area (NTS 93F north half) established the distribution, age, structural setting, composition, and isotopic character of the newly defined Eocene Nechako Plateau group (53-45 Ma; e.g., Grainger, 2000; Grainger et al., 2001) and Neogene Cheslatta Lake suite mafic rocks (13-11 Ma), and associated intrusions (e.g., Resnick, 1999; Anderson et al., 2001). The Nechako Plateau group comprises Lower Eocene felsic rocks of the Ootsa Lake Formation (53-47 Ma), and the overlying Middle Eocene mafic rocks of the Endako Formation (51-45 Ma).

The faulted, Eocene bimodal volcano-plutonic complex developed coeval with terminal cooling of the nearby, extensional Vanderhoof Metamorphic Complex (VMC; e.g. Wetherup, 1997, 1998; Anderson et al., 1998). Faulting controlled distribution of felsic Ootsa Lake and overlying mafic Endako formations rocks as well as coeval, compositionally similar, high level, miarolitic biotite granite and diabase plutons. Post-Eocene movements on the small- and large-scale, NE- and NW-trending, syn- to post-volcanic fault systems that bound the complex also account for post-depositional, regional northeastward 20 degree dips of the strata.

Development of the Eocene volcano-plutonic complex just preceded important outboard reorganization of oceanic and continental plates and attendant extensional tectonics predicted in the North American plate (e.g., Engebretson et al., 1985) and is contemporaneous with intra-continental transtensional tectonics. (e.g., Struik, 1993). Oligocene-Miocene magmatism in south-central British Columbia was the result of back-arc partial melting of the upper mantle in the asthenosphere inboard of the Pemberton volcanic belt (Bevier 1983a, 1983b). A similar outboard arc (Masset formation; Hickson 1991) and inboard alkaline mafic volcanic field (Cheslatta Lake suite; Anderson et al., 2001) system extends north to central British Columbia.

Ootsa Lake formation comprises rhyolitic and dacitic, biotite-quartz-alkali feldspar phyric tuff, pyroclastic rocks, breccia, flows, and siliceous glassy domes and includes minor basal to medial basalt and andesite flows. The felsic volcanic rocks are silicic, calc-alkaline, rich

in potash and large ion lithophile elements, peraluminous, and have moderate to marked Eu* anomalies.

Fault-bounded, miarolitic leucocratic biotite granite and felsic quartz-biotite-alkali feldspar porphyry and aphanitic plutons are associated with, intrude, and slightly postdate (ca. 47 Ma) the Ootsa Lake formation volcanic country rocks and related block faults. The plutons are silicic, calc-alkaline, high-potash, metaluminous to peraluminous, mostly of within-plate granite affinity, and have chondrite-normalized rare earth element patterns identical to those of the Ootsa Lake Formation volcanic rocks.

The Endako formation consists of amygdaloidal, hypersthene-augite-plagioclase phyric andesite to basaltic andesite flows and hyaloclastite. The sequence was deposited with increasing disconformity on Ootsa Lake formation over one hundred kilometers from northwest to southeast towards the VMC. The age of onset of Endako formation magmatism also decreases and the hiatus in volcanism represented by its basal contact increases from 0-6 m.y. towards VMC. The relationships among: 1) the age and distribution of syn-volcanic faults; 2) the stratigraphic and temporal characteristics involving the onset of Endako formation volcanism, and; 3) the history of the VMC suggest the change from ductile to brittle deformation in the complex affected the localization and onset of fissure eruptions that produced Endako formation rocks (e.g., Anderson et al., 1998; Grainger et al., 2001).

Geochemical studies indicate that the unit is internally heterogeneous. It generally has mixed mildly alkaline transitional to calc-alkaline, hy-normative, high-potash, within plate continental to volcanic arc geochemical affinities distinct from younger Neogene basanite and alkali olivine basalt of the Cheslatta Lake Suite (e.g., Anderson et al., 1998, 2001).

The Cheslatta Lake Suite in the Nechako River area encompasses Miocene primitive alkaline magmatism, preserved as scattered erosional remnants of volcanic necks, dykes, and lava flows (Resnick, 1999). The centres are localized along re-activated Eocene block faults. The rocks range from (plagioclase-clinopyroxene-) olivine phyric basanite through to transitional basalt with a groundmass comprising plagioclase, clinopyroxene, Fe and Ti oxides, feldspathoid, olivine, and apatite. Ultramafic xenoliths are common and distinctive; plutonic and metamorphic xenoliths are rare but significant in refining interpretations of crustal composition (Resnick, 1999). The rocks range in age from mid-Miocene (13-11 Ma; Ar-Ar) in the west to Late Oligocene (ca. 28 Ma; Ar-Ar) for the Summit Lake centre in the east (e.g., Anderson et al., 2001; Grainger, 2000).

All the Neogene samples are alkaline, high-potash, and encompass within-plate and continental affinities. The samples contain > 10 wt. % normative olivine (Mg numbers: 72-42) and most are silica undersaturated with more than 3 wt. % normative nepheline; a few are hypersthene normative. Some compositions are consistent with near primitive mantle melt compositions. Systematic differences in Ca, Ti, Al, Fe, Zr, Ce, Sr, P, Nd, Sm, Sc, Lu, and Yb abundances which distinguish Neogene similar basanite or alkali olivine basalt rock types that occur in the Stikine vs. Cache Creek or Quesnel terranes in central B.C. likely indicate regional heterogeneities in the mantle source regions for the lavas. The Cheslatta Lake suite volcanic rocks in the Intermontane Belt differ from the Chilcotin Group in the study area and to the south

in the form of their centres, preserved thickness, phenocryst/xenocryst mineralogy, amygdule abundance, included xenoliths, isotopic age, and major and incompatible, high field strength, and rare earth trace element contents (e.g., Anderson et al., 2001).

Ootsa Lake and Endako formations rocks are isotopically primitive (Sri: 0.70401-42; epsilonNd: +1.3 to +3.2). Isotopic compositions for Miocene Cheslatta Lake suite rocks in Stikine Terrane and the Oligocene Summit Lake centre (in Quesnel Terrane) are terrane-specific (Grainger 2000).

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Massive Sulphide Genesis in an Ensialic Marginal Basin, Kootenay Terrane, B.C.

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The Eagle Bay Assemblage in the Johnson Lake Area is composed dominantly of deformed, greenschist grade metamorphic rocks derived from volcanic-sedimentary sequences that are thought to range in age from the Lower Cambrian to Devonian-Mississippian periods. This region is host to a number of base and precious metal occurrences of variable host lithology and style of mineralization

The Lower Cambrian sequence is composed of quartzite, pelitic schist, marble, greenstone and green schist derived from subalkaline mafic volcanics and is interpreted to be in thrust contact with the Devonian-Mississippian sequence in the Johnson lake area. Marbles are locally correlated with the archaeocyathid bearing Tschinikin limestone.

The Devonian-Mississippian sequence is composed of alkaline bimodal metavolcanics and metasedimentary rocks. The metavolcanic suite is derived from mafic to intermediate flows, pillow lavas, volcanoclastics, and quartz-feldspar porphyritic rhyolite to rhyodacite flows and volcanoclastics. The metasedimentary sequence is composed of schist and phyllite derived dominantly from graphitic shale, quartz wacke, quartz sandstone and quartz-lithic pebble conglomerate. U-Pb zircon geochronology performed on the felsic volcanic sequence has yielded dates of 360 Ma interpreted to represent a crystallization age.

Volcanogenic massive sulphides are associated with bimodal volcanics/volcanoclastics and are either hosted within the volcanic sequences (Homestake) or along the contact with metasediments (Rea,). These are polymetallic (Zn, Pb, Cu, Au, Ag, Ba) sulphide deposits with enrichments in precious metals. The Eagle Bay Assemblage records various stages of rifting of the Ancient Pacific continental margin. The Cambrian sequence is evidence of Lower Paleozoic rifting and is host to sedimentary exhalative massive sulphide occurrences. The Devonian Mississippian sequence records Mid-Paleozoic rifting and may represent the vestiges of ensialic back-arc basin formation. The lithology, mineral occurrences and proposed tectonic setting have similarities with the Tetagouche Group in the Bathurst District of New Brunswick, and Okinawa Trough, Japan.

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Lower Ordovician-Lower Devonian stratigraphic framework along a Macdonald Platform to Ospika Embayment transect, Trutch, Halfway River and Ware map areas (94G, B, F), northeastern British Columbia

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Lower Paleozoic studies along a northeast-southwest, platform to basin transect were initiated during two field seasons (1998, 1999) to address the early Paleozoic evolution of the western margin of Laurentia. These complement a similar transect in northernmost BC from the Muskwa River to the Kechika River and across the parautochthonous Cassiar Terrane. The stratigraphic framework for this northern area, which focused on the Lower Ordovician succession and introduced a revised stratigraphic nomenclature, has recently been described (Pyle and Barnes, 2000). For the current project, a total of 11 stratigraphic sections were examined in which over 8 850 m of section were measured and described and a total of 257 conodont samples collected (4-5 kg each). This transect extends from the Macdonald Platform southwest to the Ospika Embayment of the southern Kechika Trough, with emphasis on sections preserving the ancient shelfbreak (Fig.1). The stratigraphic units measured, described and sampled include: a) the platformal succession (uppermost Kechika, Skoki, Beaverfoot, Lower Silurian Limestone, Nonda), b) unnamed transitional facies of the Road River Group (shelfbreak-slope equivalents of the Beaverfoot, Silurian Limestone and Nonda including a Shale and Quartzite unit, Carbonaceous Limestone unit and Unnamed Breccia unit) and c) basinal facies of the Road River Group (Ospika and Kwadacha formations) (Fig. 2).

The platformal succession of Kechika, Skoki, Beaverfoot and Nonda formations studied at the Sections 1 and 2 in 1998 was complemented by the Sections 3 and 4 in 1999 (Fig. 3). The Skoki (over 600 m thick) comprises massive bedded dolostone that is overlain unconformably by massive dolostone and sandstone of the Beaverfoot (320 m). The Beaverfoot is overlain abruptly by thin bedded, uniform limestone with shale partings of the unnamed Silurian Limestone (220 m) that becomes more massive up-section and grades into the fossiliferous limestone and massive dolostone of the Nonda (140 m). At the Section 4, both the Silurian Limestone and Nonda are more argillaceous.

A spectacular section north of Mt. Kenny (Fig. 1) preserves an interfingering relationship of the Nonda Formation (200 m) and its off-shelf lateral equivalent, the unnamed Carbonaceous Limestone unit (250 m), which comprises thin bedded, dark grey limestone and a series of thick to massive, dolomitic, fossiliferous, debris flow breccias. Traced westwards, the Nonda appears to also correlate with the Unnamed Breccia, a composite of debris flow breccias that contain monomictic clasts of Nonda lithology as well as silicified corals and brachiopods typical of the Nonda.

The off-shelf equivalent of the Beaverfoot Formation is the unnamed Shale and Quartzite unit which unconformably overlies the Skoki at two localities (Sections 6, 7; Figs. 2, 3). The shale and quartzite comprises two distinct subunits: brown weathering, thin bedded argillaceous limestone and shale with Kechika-like striped weathering and graptolitic black shale punctuated by massive beds of quartzite. At both localities, the shale has yielded graptolites of the *clingani* Zone (Caradocian, Upper Ordovician; G. Ortega and S. H. Williams, pers. comm., 1999, 2000).

In the eastern Ware map area, west of the shelfbreak, two sections of the Road River Group were studied (Sections 9, 10; Fig. 1). A section of Ordovician Road River Group, Ospika Formation, included variegated shale and thin limestone-dolostone beds, volcanics and units of siltstone and limestone breccia (780 m) unconformably overlain by uniform, orange weathering, finely laminated siltstone of the Kwadacha Formation (500 m; formerly Silurian Siltstone). An unnamed unit of Devonian Road River Group, comprising a thick sequence of mainly crinoidal, carbonate turbidites and black shale, yielded polygnathid and icriodid platform conodonts that represent several zones in the Lower to Middle Devonian.

The detailed stratigraphic and conodont biostratigraphic work has focused on the unnamed Road River Group units and should clarify the east-west temporal and facies relationships of a stratigraphic interval for which there has been barely any biostratigraphic control (Fig. 2). The conodont biostratigraphic and paleoecologic studies will address several stratigraphic and tectono-stratigraphic questions. For example, from preliminary results, the age of the volcanics within the Skoki and Ospika formations is Middle Ordovician and the quartzite of the Unnamed Shale and Quartzite is Upper Ordovician and time equivalent to the Mt. Wilson quartzite in the southern Cordillera.

Taxonomy, biostratigraphy and paleoecology of the conodont faunas are currently in progress. Approximately 15 000 conodont elements have been recovered with moderate to excellent preservation with conodont colour alteration values in the range of CAI 3-5. The Kechika-Skoki contact lies within the *Oepikodus communis* Zone (Lower Arenigian, Lower Ordovician; Fig. 2). The Skoki Formation contains a sequence of faunas that range from the Arenigian into the Caradocian based on the zone species *Pygodus anserinus* which occurs near the top of the Skoki at the Section 8. There, the Skoki is overlain by Ordovician Road River Group black shale and limestone (Ospika Formation) that yield *Baltoniodus* cf. *B. alobatus* (early Caradocian). Conodonts from near the base and top of the Beaverfoot suggest a mid-Caradocian and a Llandovery (Lower Silurian) age, respectively. The Ordovician-Silurian boundary lies within the Carbonaceous Limestone unit; in the upper part of which occurs *Distomodus kentuckyensis*, indicative of the Llandovery. Distinctive platform conodonts from the Unnamed Breccia unit are assigned to the *Pterospiriferus amorphognathoides* Zone, close to the Llandovery-Wenlock boundary.

In summary, the stratigraphic framework for this North-central Rockies transect has been completed and a companion paper to that of Pyle and Barnes (2000) that dealt with the Northern Rockies transect is submitted. Both papers introduce a new stratigraphic nomenclature for the Upper Cambrian-Lower Devonian succession and document facies relationships from the platform carbonates to the east, across the interfingering facies of the shelfbreak, to the more condensed shale-limestone sequences of the Kechika Trough and Ospika Embayment. The detailed conodont biostratigraphy being developed will enable a more complete assessment of the sequence stratigraphy leading to a more constrained interpretation of the tectonic evolution of the Cordilleran margin of Laurentia in this Early Paleozoic post-initial rift phase.

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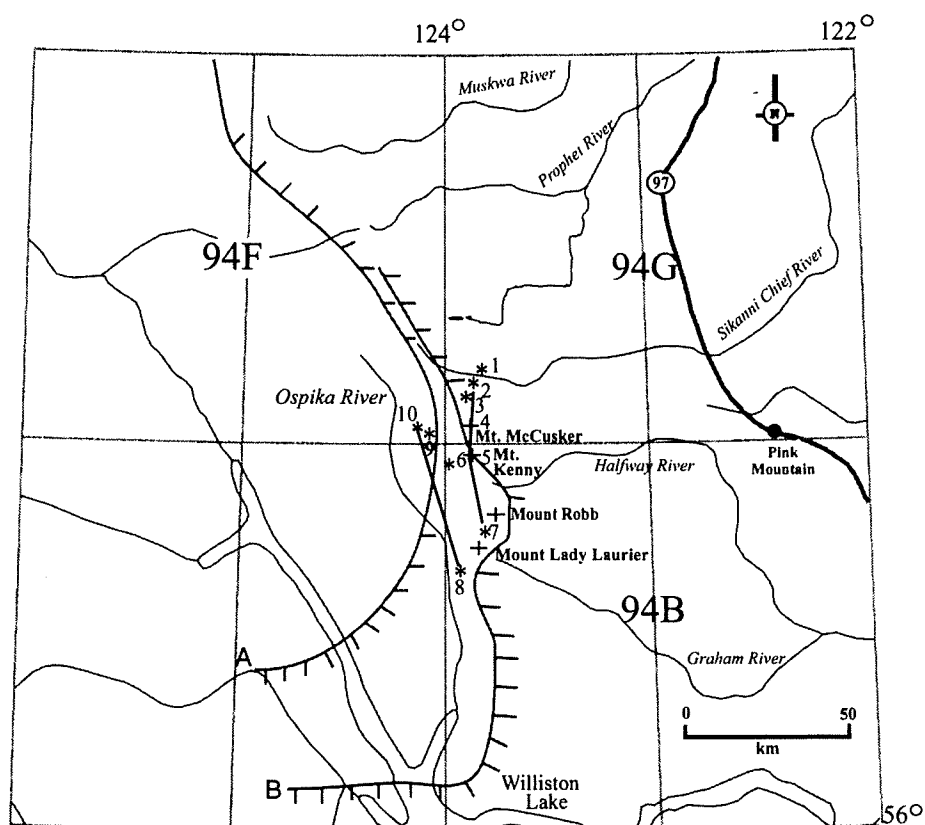


Fig. 1. Section localities, lateral shift of platform margin (teeth point toward shelf) and lines of cross-section in Fig. 3: 1. Sikanni Chief River North, 2. Sikanni Chief River South, 3. Gautschi Creek, 4. Mount McCusker, 5. Mount Kenny (creek and ridge sections), 6. Mount Kenny East, 7. Mount Robb, 8. Mount Lady Laurier, 9. Devonian Ospika River Section, 10. Ospika River North.
A. Eastern edge of Ospika Embayment, Mid-Ordovician (after Cecile and Norford, 1979);
B. Eastern edge of Ospika Embayment, Late Ordovician-Early Silurian (after Thompson, 1989).

Subsystem	AREA SERIES/ STAGE	OSPIKA EMBAYMENT (SHELFBREAK)		MACDONALD PLATFORM (PLATFORM-SHELF)
Lower Devonian		ROAD RIVER GROUP	Brown Siltstone Unit	Muncho-McConnell Formation
Upper Silurian	Pridoli		Unnamed Breccia Unit	Nonda Formation
	Ludlow		Unnamed Carbonaceous Limestone Unit	Unnamed Silurian Limestone
Lower Silurian	Wenlock		Unnamed Shale and Quartzite Unit	Beaverfoot Formation
	Llandovery		Ospika Fm.	Skoki Formation
Upper Ordovician	Ashgillian		Skoki Fm.	
	Caradocian			
Middle Ordovician	Darriwilian			
Lower Ordovician	Arenigian			
	Tremadocian		Kechika Formation	Kechika Formation
Upper Cambrian	Trempealeauan			

Fig. 2. Lower Paleozoic stratigraphic framework of platform to basin transect. Question marks indicate where biostratigraphic resolution is required.

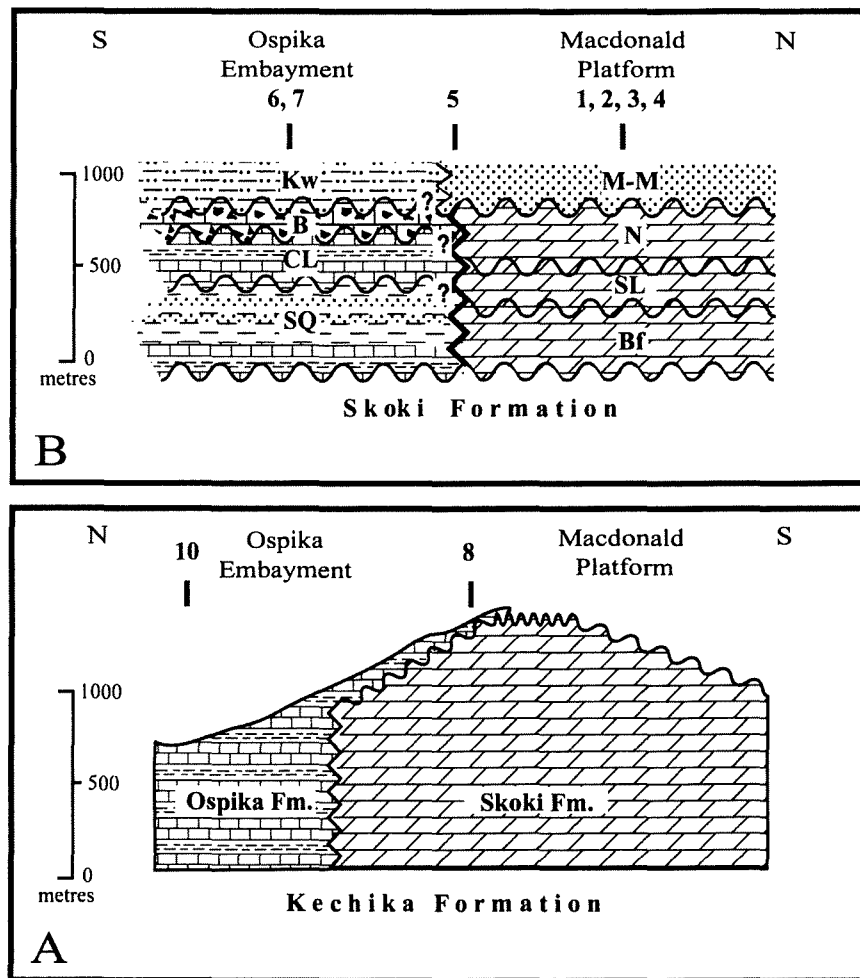


Fig. 3. Cross-sections (lines shown in Fig. 1) across the Macdonald Platform to Ospika Embayment: A. Mid-Ordovician abrupt platform to basin transition from Skoki to Ospika Fm. of Road River Group. B. Late Ordovician-Early Silurian platform (Bf = Beaverfoot Fm.; SL = Silurian Limestone; N = Nonda Fm.; M-M = Muncho-McConnell Fm.) to offshelf facies of Road River Group (SQ = shale/quartzite unit; CL = Carbonaceous Limestone unit; B = Breccia unit; Kw. = Kwadacha Fm.) (modified from Thompson, 1989).

New ^{40}Ar - ^{39}Ar biotite, muscovite and hornblende ages along the Rocky Mountain Trench between Golden and the Big Bend, British Columbia

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GEOLOGICAL SETTING

In the Golden and Big Bend region of southeastern BC the Rocky Mountain Trench (RMT) is the geomorphological boundary (in part fault controlled) between the Rocky Mountain Belt and the Omineca Belt. Early Tertiary normal faults in the RMT obliquely cut through and offset the out-of-sequence Purcell Thrust. The hanging wall of the Purcell Thrust, to the southwest, is dominated by mainly clastic rifted continental margin sediments of the Upper Proterozoic Horsethief Creek, Kaza and Cariboo Groups and their equivalents overlain by the Lower Cambrian Hamill Group. The footwall of the Purcell Thrust is dominated by the rifted and passive continental margin sediments of the Upper Proterozoic Miette Group, Lower Cambrian Gog Group, the passive continental margin sediments of the Middle Cambrian Chancellor Group and the Upper Cambrian / Lower Ordovician McKay Group (see Figure 1). The metamorphic grade increases from sub-greenschist facies (development of pressure solution cleavage) in the south near the Trans-Canada Highway to mid-amphibolite facies (migmatitic pelites with kyanite and kyanite + staurolite) in the northern Big Bend region.

SAMPLE LOCATION, PREPARATION AND ANALYSES

Seven samples were chosen for ^{40}Ar - ^{39}Ar analyses of biotite, hornblende and muscovite in an attempt to constrain timing of movement along the Purcell Thrust. Five samples from near the Big Bend of the Columbia River form a transect across the Purcell Thrust. The other two samples come from the hanging wall of the Purcell Thrust in the southern third of the field area (see Table 1, Figure 1).

The samples were crushed and separated at the University of Calgary, irradiated in the nuclear reactor at McMaster University in Hamilton Ontario and analysed in the ^{40}Ar -

^{39}Ar lab at Queen's University in Kingston Ontario. The mineral separates were step heated to fusion, in 8 to 14 steps.

^{40}Ar - ^{39}Ar RESULTS

The ^{40}Ar - ^{39}Ar systematic closure temperatures are 300 - 350°C for biotite (Purdy and Jager 1976, Harrison et al. 1985, Dodson and McClelland-Brown 1985), 350 - 400°C for muscovite (Hanson and Gast 1967, Purdy and Jager 1976, Cliff and Cohen 1980) and ~500°C for hornblende (Harrison 1981). Therefore apparent ages would be expected that decrease in order from hornblende to muscovite to biotite. In three samples, the biotite ages were 6 to 25Ma older than the muscovite ages (see Table 1), likely due to excess argon (McDougall and Harrison 1988). Step-heating of the biotite and muscovite separates resulted in flat plateaus (see Figure 2). The correlated, integrated and plateau ages all agree within error values which suggests that these are internally and statistically consistent ages.

This close correspondence between biotite and muscovite ages suggests rapid cooling from 400°C through 300°C between 54.4 and 57.6Ma. Such a rapid cooling could correspond to a rapid uplift that likely resulted from Early Tertiary regional extension. The ^{40}Ar - ^{39}Ar mica ages from the southern portion of the field area agree within error with the northern ^{40}Ar - ^{39}Ar mica ages. Therefore the biotite-muscovite ^{40}Ar - ^{39}Ar ages represent a rapid regional cooling and are thus unsuitable for dating movement along the Purcell Thrust.

Hornblende ^{40}Ar - ^{39}Ar ages from this region were more difficult to interpret because of "U" shaped step-heating profiles (see Figure 3) and low K content. Two hornblende separates were analysed. Most of the ^{39}Ar from one sample was released during four steps that gave a second plateau age of 164.84 +/- 4.2Ma (see Figure 3). This 164Ma age could correspond to movement along the Purcell Thrust in the south of the study area. The 98Ma first plateau age could represent partially reset ^{40}Ar - ^{39}Ar systematics due to pervasive fluid flow during Early Tertiary regional extension.

Further work will include the ^{40}Ar - ^{39}Ar analyses of whole rock and sericite separates from the one known exposure of the Purcell Thrust in the south of the field area

near the Trans-Canada Highway. More hornblende separates from calc-pelites throughout the field area will also be analysed to determine if there is consistent evidence for a 164Ma event in the field area.

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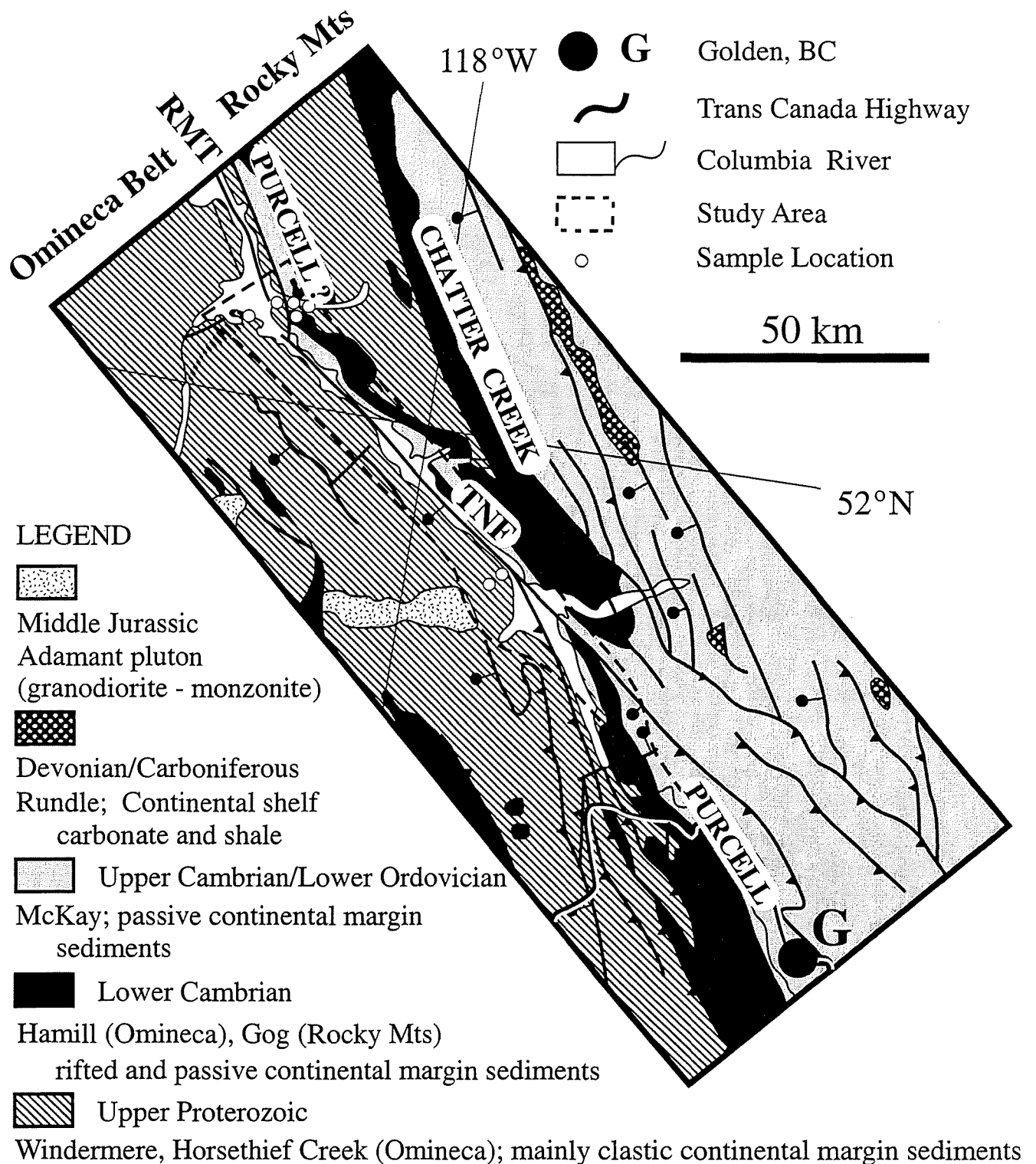


FIGURE 1: Regional Geological Map (modified after Wheeler and McFeely, 1991). Note that the location of the Purcell and Chatter Creek Faults are labelled, that RMT = Rocky Mountain Trench and that TNF = Trench Normal Fault.

Table 1: 40Ar-39Ar Analyses Results

Outcrop	FW / HW (N/S)	D to PT	Metamorphic Grade	Mineral Analysed	Integrated Age (Ma)	Correlation Age (Ma)	Plateau Age (Ma)	2nd Plateau (Ma)
KB0016	HW (N)	100m	Mid Amphibolite	Biotite	54.8+1.18	56.57+1.71	56.75+1.13	
KB0016	HW (N)	100m	Mid Amphibolite	Muscovite	56.32+0.94	57.62+2.26		
KB0035	HW (N)	6km	Mid Amphibolite	Biotite	58.9+1.2	57.56+1.03	58.85+1.19	
KB0035	HW (N)	6km	Mid Amphibolite	Muscovite	54.42+1.19	54.01+0.72	54.42+1.18	
KB0071	FW (N)	100m	Mid Amphibolite	Biotite	56.31+1.52	55.81+3.55	56.51+1.42	
KB0071	FW (N)	100m	Mid Amphibolite	Muscovite	57.24+1.09	56.96+1.21	57.10+1.08	
KB0075	FW (N)	2.1km	Mid Amphibolite	Biotite	62.32+1.03	61.45+0.82	62.56+1.04	
KB0075	FW (N)	2.1km	Mid Amphibolite	Muscovite	56.63+1.07	55.7+4.14	56.5+1.03	
KB0075A	FW (N)	2.1km	Mid Amphibolite	Hornblend	108.8+3.9	93.26+5.04	98.38+3.25	
KB0077	FW (N)	1.2km	Mid Amphibolite	Biotite	65.28+1.35	71.03+2.1	67.82+1.42	
KB0077	FW (N)	1.2km	Mid Amphibolite	Muscovite	57.84+0.97	56.38+1.46	57.75+1.0	
KB6001	HW (S)	1.2km	Greenschist	Biotite	58.68+1.04	58.36+0.46	58.46+1.04	
KB6001	HW (S)	1.2km	Greenschist	Hornblend	198.86+4.28	40.25+19.48	82.75+3.16	164.84+4.2
KB6001	HW (S)	1.2km	Greenschist	Muscovite	59.82+1.16	55.3+1.42	56.42+1.11	
KB6111	HW (S)	4.6km	Mid Amphibolite	Biotite	80.75+2.37	82.72+2.11	82.66+1.45	
KB6111	HW (S)	4.6km	Mid Amphibolite	Muscovite	56.86+1.93	55.38+9.66	56.66+1.66	

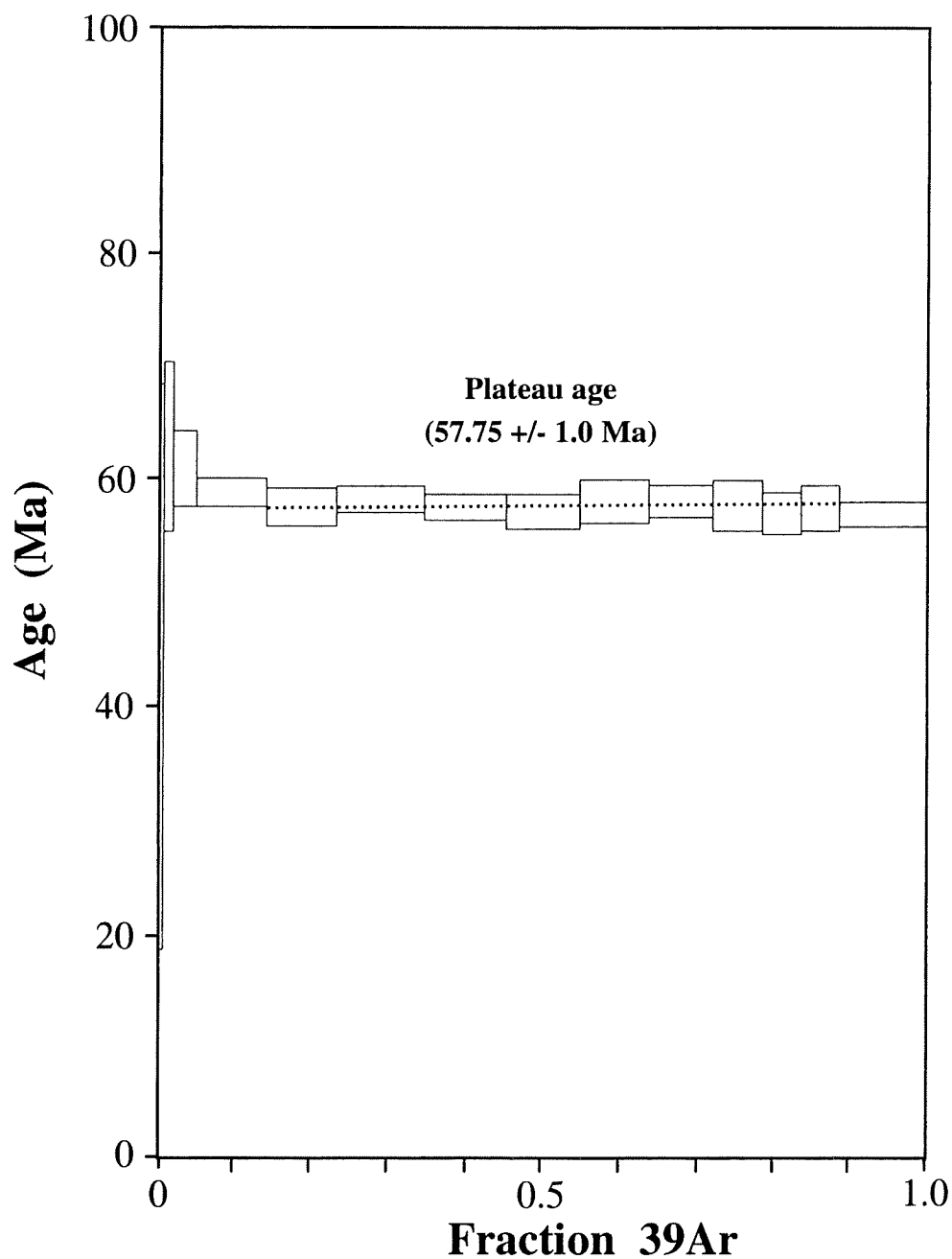


FIGURE 2: The step heating profile for muscovite from outcrop 0077. The profiles for the muscovite and biotite in this study were flat. In this analyses, atmospheric Ar was burned off during the first four steps, therefore, these steps were not used to calculate the plateau age. Note (see Table 1) that for biotite and muscovite the correlated, integrated and plateau ages all agree within error.

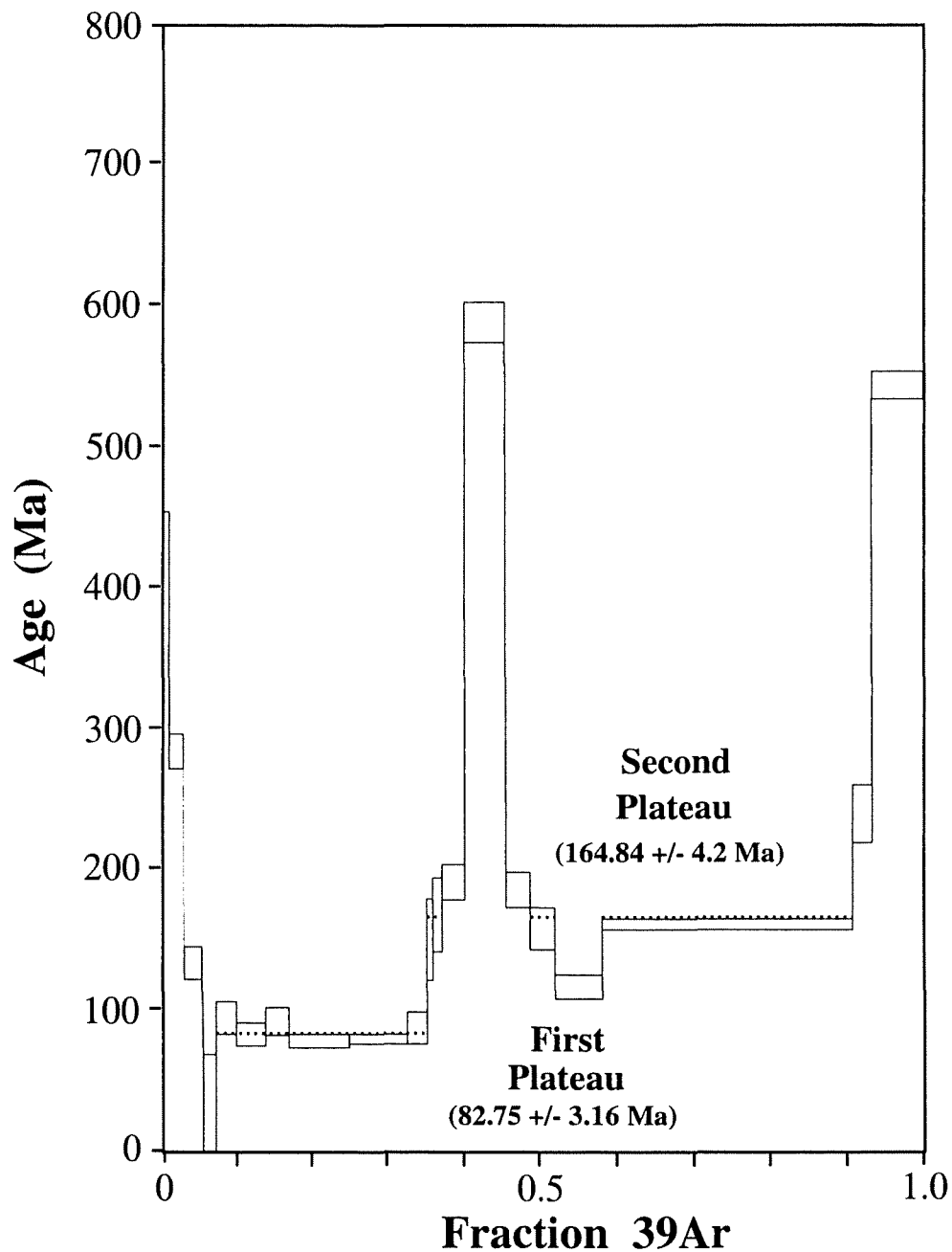


FIGURE 3: The step heating profile for hornblende from outcrop 6001. Two different separates were analysed, with the steps for the second analyses added on to the right of the steps from the first analyses. Most of the ^{39}Ar was released from the separate during the four steps used to calculate the second plateau age. This suggests that the second plateau age would have more geological significance than the first plateau age.

UPPER CRUSTAL ANALYSIS OF SNORCLE TRANSECT 2A, THROUGH THE BOWSER BASIN, NORTHERN B.C.: REFLECTION DATA IN COMBINATION WITH STRUCTURAL FIELD DATA.

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Recently acquired seismic reflection data (SNORCLE line 2A) through northwestern British Columbia and southern Yukon (Fig. 1), provides the opportunity for a comprehensive study of the north central Cordillera. Line 2A, which crosses a number of terrane boundaries including that between the Coast Belt and Stikinia allows the addressing of important questions relating to the accretion of terranes in the region. This study focuses on a section of the transect through the Bowser Basin where interpretation of the reflection data will be facilitated by detailed structural studies of select areas located within 10 km of the Lithoprobe transect. The portion of Line 2A studied is ~150 km long, transects the basin mainly along strike and relatively close to the present-day western margin of the Bowser Basin.

The Mesozoic Bowser Basin is a clastic sedimentary basin, in which the lithofacies assemblages and fossils record its evolution from the mid Jurassic to the mid Cretaceous (Evenchick, 2000). The basinal sedimentary sequence was subsequently deformed by the Skeena Fold Belt, considered to be active from the early Cretaceous to early Tertiary (Evenchick, 2000). Northwest trending folds characterize the fold belt; the scale of these structures involves wavelengths from hundreds of metres to kilometres and amplitudes of tens to hundreds of metres (Evenchick, 1991, 1996). Throughout the Bowser Basin, evidence of a separate phase of folding with a northeast trend has been identified (Evenchick, 1996, 1998). These folds may be related to other northeast trending structures to the north of the fold belt, that date between 135 and 106 Ma (Evenchick, 1998). Evenchick (1998, 2001) suggests that the northeast trending folds were formed during a period of sinistral convergence across the Cordilleran margin in the mid Cretaceous. An objective of this study is to provide structural data from select areas containing northeast and northwest trending folds, with the aim of further testing the hypothesis that the northeast trending folds are older and related to sinistral convergence.

Structural data have been collected from a number of areas in the western half of the basin, within 10 km of SNORCLE Line 2A. Regional mapping of the area (Evenchick, 1993, 1991b) identified distinct areas of differing structural trends. Within these areas several localities were mapped for this study and were chosen on the basis of their structural trends. Within the fold belt, areas of northeast trending folds tend to be distinct from their northwestern neighbours except in a small number of areas where interference between the two sets of

structures has been identified. For example, to the south of Todagin Creek (Fig. 1), a large scale interference structure exists, resulting from the interaction of both orientations of folding within competent, mainly conglomeratic rocks. Structural data from this area of fold interference are being compared to areas that show less obvious structural relationships, south of Burrage Creek and north of Ogilvie Creek (Fig. 1). These two areas show distinct zoning of fold trends with a dominant northwestern trend and discrete zones of northeast trending structures. Data analysis of these areas may illuminate details of the sequence of deformation that is proving hard to differentiate within the interference pattern, and was not resolvable at the previous regional scale of mapping. We anticipate that bedding – cleavage relationships and in particular the timing of cleavage formation with respect to bedding may shed further light on the origin of the northeast trending folds.

The structural analysis described above will be combined with the new reflection data of SNORCLE line 2A to better interpret the subsurface signature of the Skeena Fold Belt in this area. The reflection data will also provide information on the nature of the northern and western margins of the basin at depth. Further combined analysis of the seismic and structural data will allow the tectonic framework of the Skeena Fold Belt to be refined.

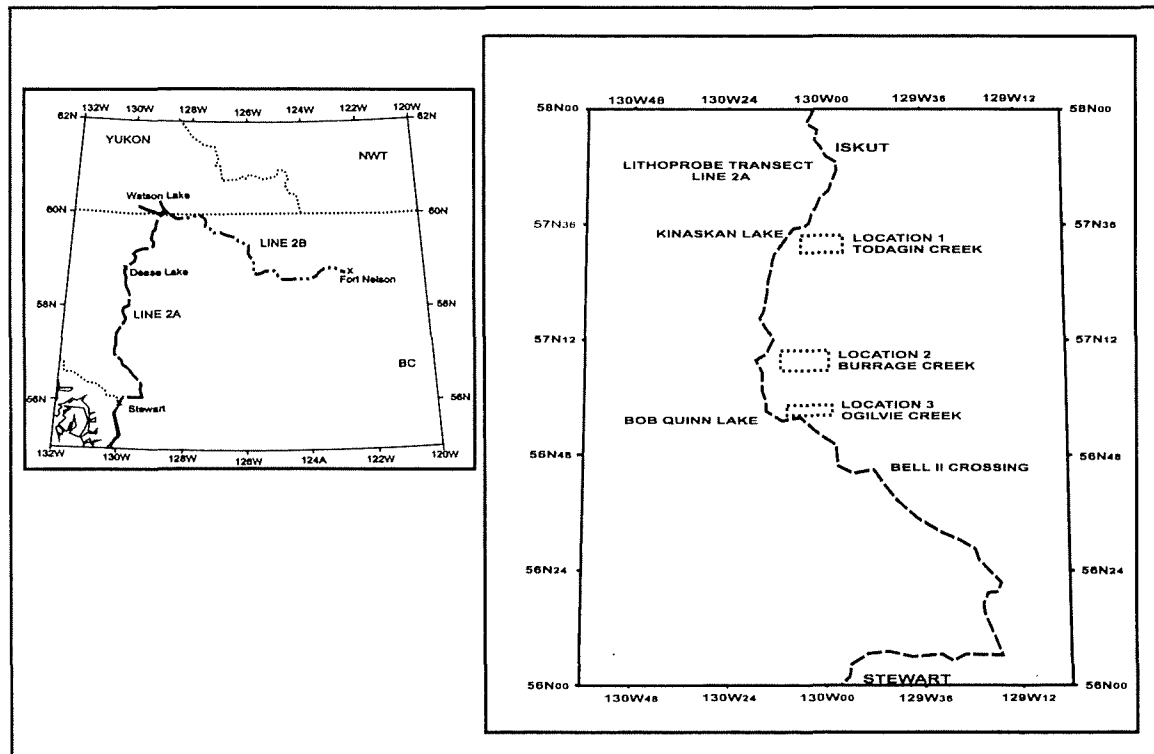


Fig. 1. Map giving location of Lithoprobe Transect 2A with respect to areas of structural analysis discussed in this article.

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CROSS-STRIKE POTENTIAL FIELD ANOMALIES IN THE CANADIAN CORDILLERA

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Abstract

Application of bandpass and directional filtering to potential field maps of western Canada has led to the discovery of regionally extensive anomalies that cross the northwest strike of Cordilleran structures. The most dramatic of these, the Steamboat-Fraser trend, is a north-south feature that projects from the Western Canada Sedimentary Basin in northeastern British Columbia, where it merges with north-south trends of the Precambrian Canadian Shield, southward to the northern limit of the Fraser River fault, a distance of about 600 km. Within the hinterland of the Cordillera, the trend appears to spatially correlate with the margins of Tertiary volcanics. In the Foreland belt the northeasternmost, and most cratonward, Tertiary volcanics are located on the northern projection of the trend. The trend may thus be interpreted as either late, post-orogenic dikes that are not exposed, or as pre-orogenic (pre-Mesozoic) features (e. g., faults) in the crust that were overridden by the Cordilleran thrusts and along which Tertiary volcanics were extruded along its southern portion. One major implication of the latter interpretation is that Precambrian Shield rocks would be present at depth beneath the southern Canadian Cordillera as far west as the Fraser River fault.

Introduction

The purpose of this paper is to describe the discovery of regional potential field anomalies, here called the Steamboat-Fraser trend, that cross the strike of the Canadian Cordillera and to discuss possible interpretations of them in the context of Cordilleran structure and tectonics. The trend has been delineated by application of bandpass and directional filtering to potential field maps (region outlined in Figure 1), an approach that provides a means of separating effects associated with known surface and subsurface trends and thus to determine the relative contributions of each.

A variety of studies, including LITHOPROBE deep reflection, refraction, electromagnetics, and various geological efforts, have provided important clues on the

nature of the crust and upper mantle beneath western Canada. Various interpretations have been presented, but many uncertainties mean that our understanding of the origin of the deep crust and upper mantle are continuously evolving.

West of the Foreland thrust and fold belt, which is a region dominated by thin, northeastward verging thrust sheets in supracrustal rocks, the hinterland of the Cordillera is interpreted to be constructed of various kinds of accreted terranes. However, the nature of the crust and upper mantle there remains elusive. Possibilities include: 1) they could be the foundation to accreted terranes that are interpreted from surface rocks, or 2) they could be North American in whole or in part with the implication that the rocks near the surface are either accreted terranes and very thin, or have not travelled far relative to the North American craton.

One method used to characterize the crust is to analyze potential field data (gravity and magnetics) and to relate the anomalies to cross sectional information provided by seismic and geological sections (e. g., Cook, 2000; Geiger and Cook, 2001). Thus, potential field results, which provide areally distributed data as maps, can complement and extend the cross sectional information from other results. In order to accomplish this effectively, however, it is usually necessary to analyze the maps with various filtering techniques to separate unwanted from desired signals.

Filtering the Potential Field Maps

Potential field maps are presented as contoured (or colored) values that have been determined by calculating the effect of an assumed Earth model, and then subtracting that result from the observations. The resultant “anomalies” are then interpreted to be the response of the measurement to local variations in mass (for gravity), or magnetic characteristics. The map images are conducive to applications of various filtering techniques to enhance or attenuate signals according to different characteristics. Two of the characteristics that are applied for this study are the signal wavelength (bandpass) and signal direction (azimuth).

Bandpass filtering allows signals to be separated according to their wavelength. This is significant for potential field maps because deep sources produce long wavelength anomalies, and shallow sources produce either long or short wavelength anomalies (depending on whether the source is wide or narrow, respectively). Thus, wavelength filtering can provide estimates of maximum source depth. Directional filtering can separate signals according to azimuth, thus allowing areal geological patterns to be related to identifiable anomaly azimuths. In general a directional filter will include a lowpass filter.

As with any filtering technique, bandpass and directional filters have limitations and

potential pitfalls. Perhaps the most serious is that as a filter includes increasingly narrow ranges of wavelengths, the resulting signal becomes increasingly sinusoidal, and discrete anomalies become difficult to delineate. For this reason, most bandpass filters include sufficiently large range of wavelengths (2 - 3 "octaves") to allow the signal to be dominated by the major anomalies. For the maps presented here, the bandpass portion of the filter includes nearly 3.5 octaves (14 km to 160 km). Similarly, the range of azimuths needs to be sufficiently small for the filtered signal to be recognizable (i. e., if the application filters out too many directions, it is difficult to distinguish any meaningful signals), and sufficiently large to be useful. For the maps presented here, the directional filters include about 20° of azimuth (e. g., Geiger and Cook, 2001).

Results

Regional data are presented in two maps: unfiltered isostatic gravity (Fig. 2a) and directional filtered isostatic gravity (065° azimuth relative to grid north; Fig. 2b) to allow comparison of the data at different stages of analysis. Detailed maps are presented for bandpass filtered isostatic gravity (Fig. 3a; 14-160 km wavelength), magnetics (Fig. 3b), and directional filtered isostatic gravity (Fig. 3c). Filtering results are presented only for the isostatic gravity because the gravity values represent responses from a wide range of depths. This contrasts with magnetic anomalies which represent responses from rocks whose temperatures are below the maximum for retaining magnetic properties (Curie temperature), typically about 550°C, which occur within the crust .

The unfiltered isostatic map illustrates three important regional characteristics: 1) The Mackenzie Mountains of the northern Canadian Cordillera have one of the largest (in area as well as magnitude) isostatic highs in North America (MMH in Fig. 2a); 2) Isostatic anomalies of the southern Canadian Cordillera are generally near zero with the exception of a high in the Foreland Belt (RMH), and 3) The Canadian Shield exhibits regionally low isostatic values.

A series of residual highs and lows generally follow the arcuate geometry of the Mackenzie Mountains and are, in many cases, associated with regionally extensive structural highs and lows, respectively, at depth (Cook, 2000). The correlation of many residual shorter wavelength features to intra-crustal structures implies that the long wavelength trends, and in particular MMH, are due to deep (lower crustal or upper mantle) mass variations.

The signal that remains after directional filtering represents features whose map orientations do not include NE anomalies such as those that are characteristic of the Canadian Shield (e. g., Great Slave Lake shear zone, GSLsz, in Fig. 2a). The directional

filtering has produced one of the most remarkable anomaly trends, here called the Steamboat-Fraser trend, in western Canada (SF in Fig. 2b). The Steamboat-Fraser trend extends from the foreland region of northeastern British Columbia to the northern limit of the Fraser River fault in southwestern British Columbia, a distance of over 600 km. The trend is remarkably straight, has an azimuth almost N-S, and a magnitude of about 20 mGals. Its width is typically about 30-50 km (Fig. 3 and 4); hence because linear (two dimensional) features such as a cylinder or rectangular prism have a maximum depth that is equal to the half-width (the distance from the center of the anomaly to one half of the value at the center), the source of the SF anomaly must be within the crust.

Interpretations

The most outstanding characteristic of the Steamboat-Fraser trend is that it crosses the regional northwest strike of the Cordilleran structures. Accordingly, it is not obviously associated with Cordilleran structures. There are, therefore, three possible interpretations for it: 1) It could be an artifact of the data or filtering process, 2) It could be a post-orogenic feature such as a late dike system, or 3) It could be a pre-orogenic feature such as a fault or dike system. We consider each of these.

Because any filtering technique can produce artifacts, it is necessary to address this possibility; however, this interpretation is considered unlikely because: a) the filter parameters included sufficient ranges of bandwidths and azimuths to allow major anomalies to be included and to minimize effects such as ringing and sideline development of the waveforms. Further, the observation that the same anomaly trend is visible on both the data with directional filtering (Fig. 3c), and with bandpass filtering (Fig. 3a), is convincing evidence that the SF trend is not created by the filtering process. Finally, the trend is visible on the unfiltered magnetic data (Fig. 3b) as well as the filtered isostatic gravity (Fig. 3c), and is thus not a characteristic of the gravity field alone.

The geological interpretation of the SF trend to a large extent depends upon whether the source is pre- or post-orogenic. If it is post-orogenic (e. g., Tertiary), a logical interpretation is that the source is a regionally extensive, linear intrusive, such as a dike. Although dikes of this orientation and extent are not observed at the surface, both the magnetic and filtered isostatic signatures of the Steamboat-Fraser trend correlate with limits of outcrops of Tertiary volcanic rocks west of the Foreland belt (Fig. 3d). Furthermore, along the northward projection of SF, small outcrops of coeval volcanic rocks are present within the Foreland belt (e. g., at about 54° 20' N. Lat, 122° 45' W. Long; open arrow in Fig. 3d), and these represent the most cratonward exposures of these volcanics in the Cordillera.

These observations can be interpreted to be consistent with a linear north-south striking conduit through which the volcanic rocks extruded (Fig. 4). South of about 54° N. Lat., the conduit broke to the surface and allowed massive amounts of volcanic rocks to be extruded. North of about 54° N. Lat., the conduit did not migrate to the surface, and the surface Cordilleran structures were not affected.

If the source of the SF trend is pre-orogenic, the implications for the crustal structure of the southern Canadian Cordillera are profound. Specifically, this feature would be the first recognizable pre-orogenic trend that could be followed across the Cordillera, albeit at an oblique angle, from east of the Foreland belt to the Fraser River fault. The implication of this geometry would be that the Precambrian crust that underlies the Western Canada Sedimentary basin and Foreland belt of the Cordillera is present beneath nearly the entire Cordillera to at least the Fraser River.

Although it is not possible to determine unequivocally whether SF is pre- or post-orogenic, the fact that there are no late (Tertiary) faults or dikes with the same orientation as the SF trend in the Foreland belt and Foreland basin of northeastern British Columbia lead us to favor the interpretation that the Steamboat-Fraser trend is, at least partially, pre-orogenic. We might speculate that a pre-existing fault or other zone of weakness was present in the middle and deep crust upon which the Cordillera was built and that this zone was the locus of magmatic migration toward the surface during the Tertiary. In the interior of the orogen some volcanics were able to reach the surface, whereas in the Foreland region they migrated only up to the middle crust.

Conclusions

Two-dimensional filtering of potential field maps in the Cordillera of western Canada has led to the discovery of a prominent trend of anomalies that cross the strike of the surface structures at an oblique angle. The trend is nearly 600 km long and can be followed from northeastern British Columbia, where its northern limit is in the foreland basin, southward across the Foreland belt, Omineca belt, to the northern limit of the Eocene Fraser River strike slip fault at the west side of the Intermontane Belt. The Steamboat-Fraser trend is the only trend found so far that crosses most of the Cordillera and is either a post-orogenic feature such as a dike, or a pre-orogenic feature such as a fault or dike. The observation that the Steamboat -Fraser trend follows the mapped edges of regional Tertiary volcanic flows and that minor Tertiary volcanics are located along the trend into the Foreland belt lead to the suggestion that SF may be a zone of weakness along which volcanics migrated prior to extrusion along its southern portion, but that did not reach the surface along its northern segment.

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Figure Captions

Figure 1. Region of study encompasses most of western Canada south of about 66° N (large shaded area). Small outlined area represents the area enlarged in Figure 4. GN is grid north and is the reference for directional filters.

Figure 2. Isostatic and filtered isostatic gravity maps. a) Isostatic gravity map of western Canada south of 66° N. Lat (Goodacre et al., 1987; Geological Survey of Canada, 1999a). Note the large regional high throughout the Mackenzie Mountains (MMH), the values near zero in the southern Cordillera, and the high in the Rocky Mountains (RMH). b) Directionally filtered isostatic gravity. The filter includes a lowpass and the filter azimuth is 065° relative to grid north, rather than true north. Note the prominent Steamboat - Fraser trend (SF) that crosses the strike of surface structures. GSLsz = Great Slave Lake shear zone. GN is grid north.

Figure 3. Comparison of a) bandpass filtered isostatic gravity data(14-160 km wavelength), b) magnetic data (Geological Survey of Canada, 1999b), c) directionally filtered gravity, and d) outcrops of Tertiary volcanic rocks overlain on the filtered gravity (c; Wheeler and McFeely, 1991). Dark arrows indicate the SF trend, and the open arrow in Figure 3d locates an outcrop of Tertiary volcanic rocks along the northern portion of the trend. In Figure 3c, the arrows point to locations of profiles A and B shown in Fig. 4.

Figure 4. Schematic model illustrating the possible linear zone along which dikes and volcanics migrated upward and were extruded in the south, but remained at depth in the north. Whether the zone was originally a pre-orogenic structure is not known. Also shown are two profiles, one in the north and one in the south, across the anomaly to illustrate the magnitude (about 15-20 mGals), and the width of the anomaly in different locations. Scale bar in the upper right is for the gravity profiles.

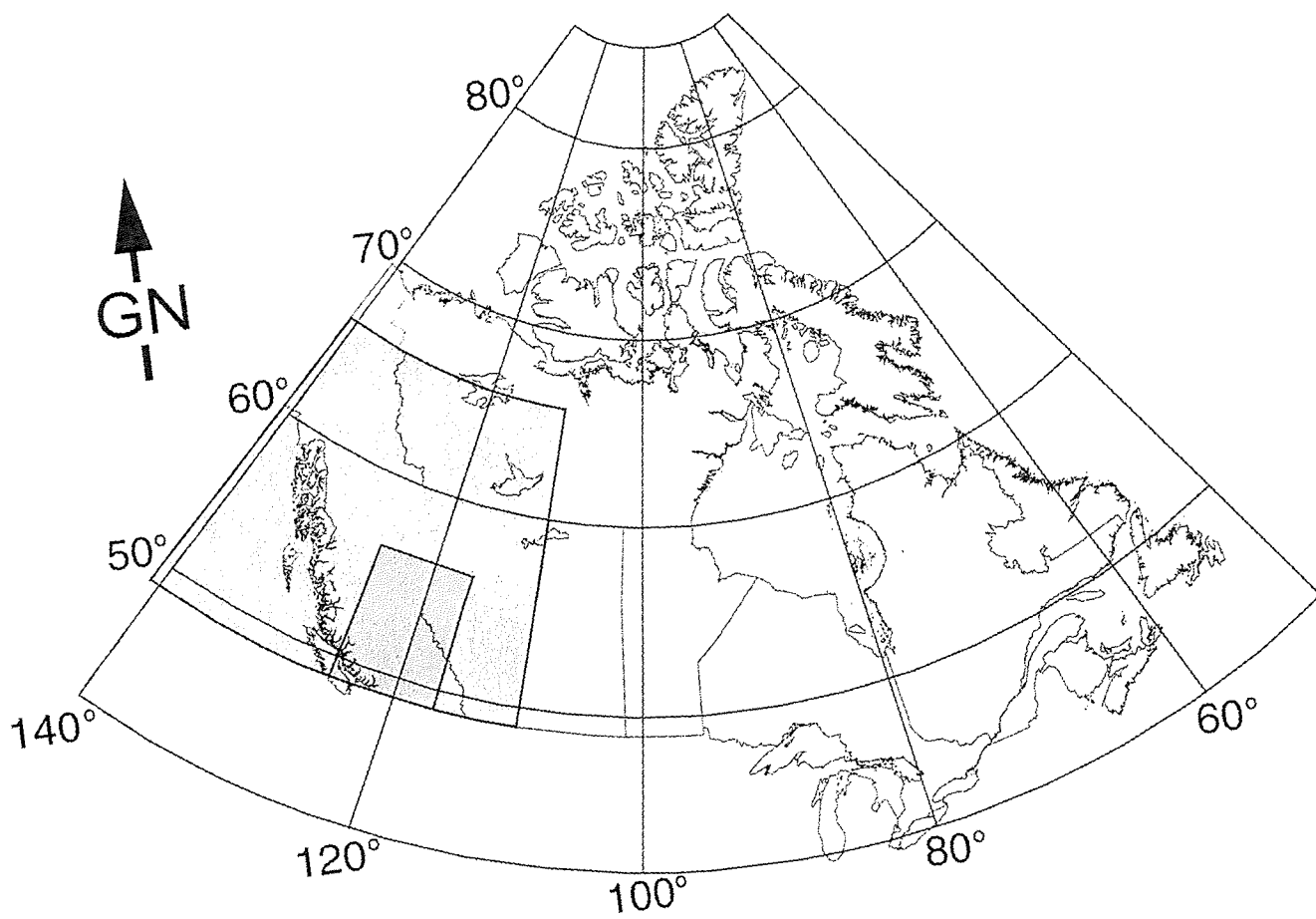


Figure 1

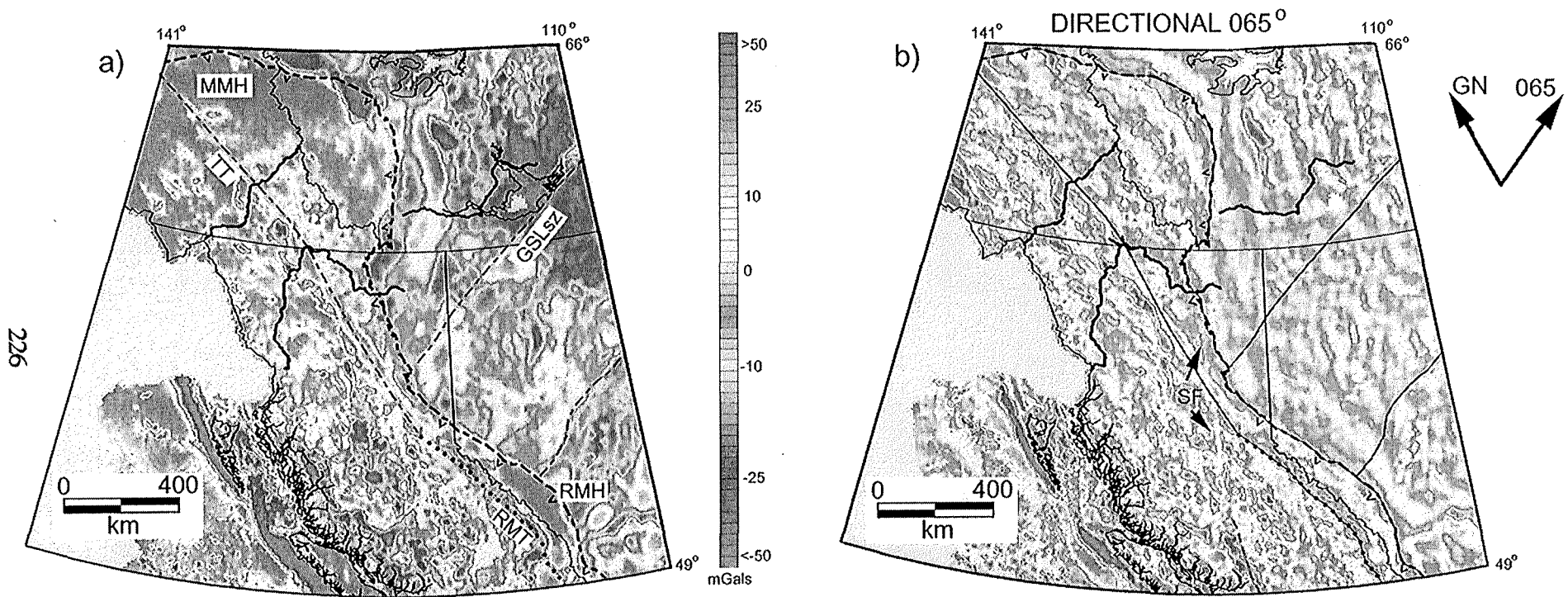
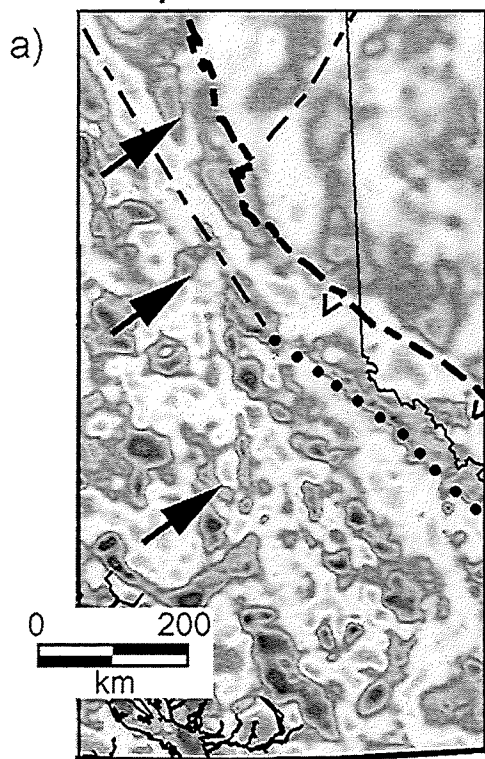
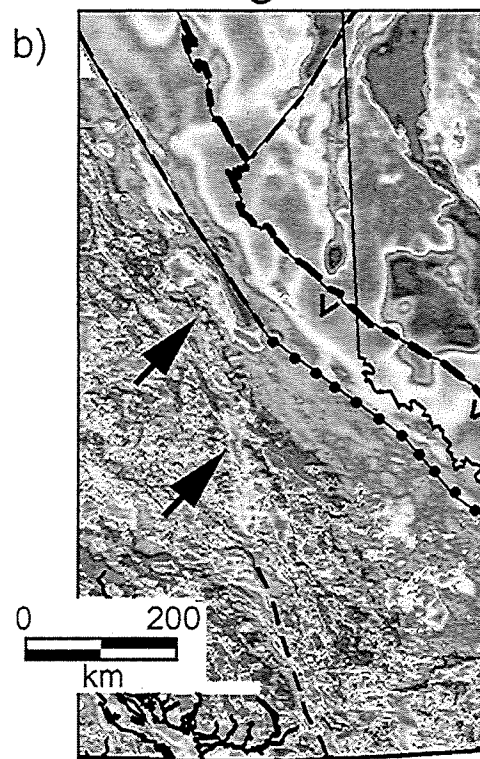


Figure 2

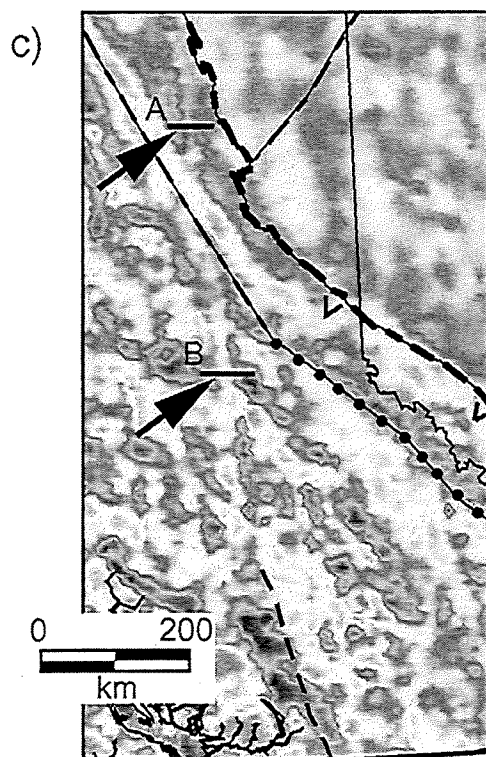
Bandpass 14-160 km



Magnetics



Directional 065°



Tertiary Volcanics

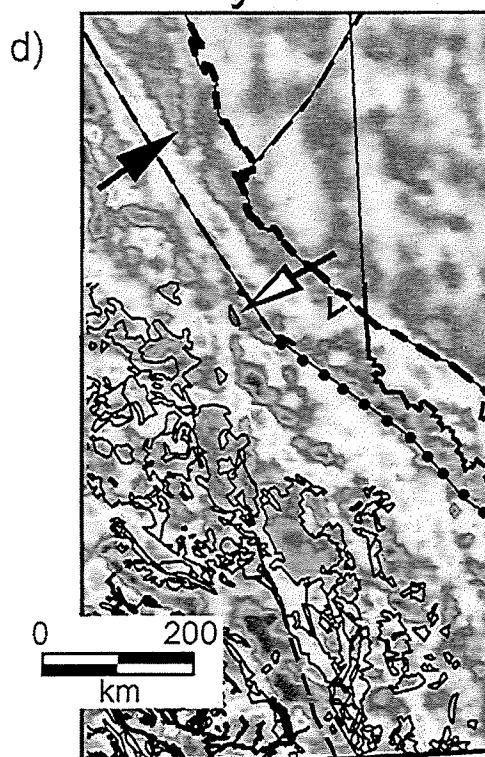


Figure 3

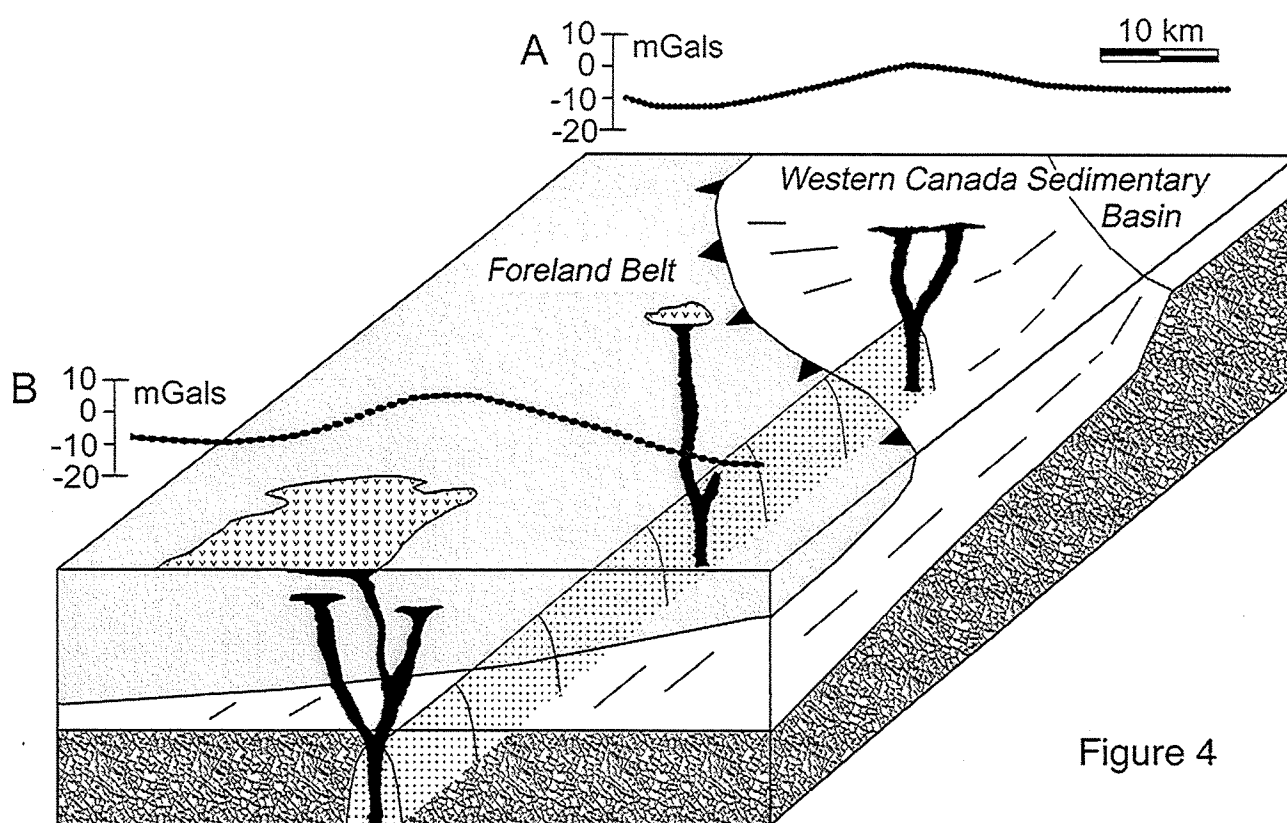


Figure 4

INVESTIGATION OF A SUB-SURFACE LATERAL RAMP IN THE SOUTHERN CANADIAN ROCKY MOUNTAINS OF SW ALBERTA BY 3-D SEISMIC INTERPRETATION: PRELIMINARY REPORT

David W. Deline and John M. Dixon

ABSTRACT

This study is part of the Fold-Fault Research Project's on-going investigation of tear-faults and lateral ramps. A lateral ramp is a ramp in a thrust surface that strikes parallel to the direction of transport. These structures have been found in the Canadian Rocky Mountain Foothills and the number of known examples is increasing as more seismic surveys are being performed. A 3-D seismic volume from the southern Canadian Cordillera in SW Alberta has been made available to FRP by one of our sponsor firms. A preliminary interpretation of this data set shows that it images a lateral ramp structure at depth. Current work is concentrated on the seismic interpretation of this volume. This will be followed up with analog centrifuge modelling using models constructed to simulate the mechanical stratigraphy of the region, in order to investigate factors which influence the nucleation of lateral ramps, and development of associated structures in the thrust sheet and footwall adjacent to lateral ramps.

INTRODUCTION

This study is part of the Fold-Fault Research Project's on-going multidisciplinary investigation of tear-faults and lateral ramps (Dixon *et al.* 1998, 1999; Halladay & Lawton 1999, 2000 (this volume)). The goals are to develop a better understanding of how lateral ramps form and the nature of associated structures in the thrust sheet and the footwall. The project will integrate the interpretation of an industry 3-D seismic survey made available to FRP by one of our sponsor firms, and new analog modelling in the Experimental Tectonics Lab at Queen's University.

A lateral ramp is a "ramp in the thrust surface that is parallel to the direction of transport of the thrust sheet" (McClay 1992; Figure 1). In the Canadian Rocky Mountain Foothills these structures commonly have ramp angles (the angle between the thrust fault and the bedding) of between 10 and 30 degrees, although some examples have been interpreted to be much steeper (Fermor 1999). If the ramp angle approaches 90° the lateral ramp should instead be called a tear-fault. Structural features commonly associated with lateral ramps are lateral folds and lateral culmination walls (Figure 1). These structures are due to the differential uplift of the thrust sheet owing to its transport over different portions of the ramp system (Dahlstrom 1970). As more seismic surveys penetrate the Rocky Mountain Foothills it is becoming apparent that lateral ramp structures are more common than previously thought. It is now known that certain thrust

sheets terminate against lateral ramps and in some cases provide structural traps that host oil and gas deposits.

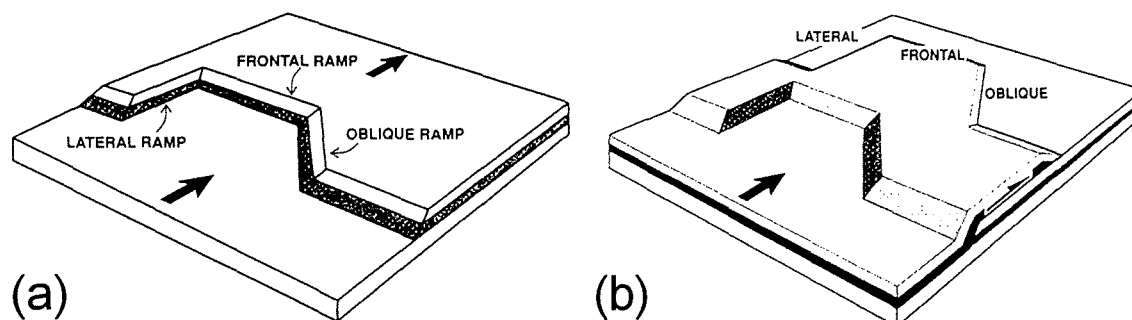


Figure 1. (a) Frontal, lateral and oblique ramps on the footwall of a thrust fault. (b) Frontal and oblique and lateral folds (culmination walls) resulting from the movement of a thrust sheet over the footwall ramps depicted in (a). The heavy arrows indicate the direction of thrust displacement. (from McClay 1992).

Some well-known structural features in the Rocky Mountains have been attributed to thrust displacement over tear-faults or lateral ramps. For example, the large eastward salient of the central Lewis thrust sheet is bounded to the north and south by lateral ramps. At North Kootenay Pass, structures in the Lewis sheet plunge steeply northward where the Lewis thrust fault cuts up-section to the north on an oblique ramp (Price 1964). To the south, at Marias Pass, the fault cuts up-section southward on a lateral ramp (Dahlstrom 1970).

It is more difficult but nonetheless important to recognize smaller-scale and sub-surface lateral ramps and tear-faults. For example, the Turner Valley thrust fault cuts up-section to the south on a lateral ramp through Paleozoic carbonates at the south end of the Turner Valley structure. Fermor (1999) provided a description of the three-dimensional nature and tectonic evolution of thrust sheets advancing over lateral ramps and discussed specific examples found in the Rocky Mountains Foothills. These examples include the Limestone Mountain culmination and the Marble Mountain “swing-back” of the Brazeau thrust sheet, and the southern extension of the Livingstone thrust sheet. F.R. Frey used 2-D and 3-D seismic data and several wells to draw a strike section across the Marble Mountain swing-back (published in Fermor 1999) in which the south edge of the Brazeau sheet is interpreted as being bounded by a tear-fault (Figure 2). Bégin *et al.* (1996) interpreted the same feature as a lateral ramp (Figure 3).

Given the potential for structures associated with lateral ramps and tear-faults to contain oil and gas, it is important to understand how and where such structures form, and to be able to recognize them in the sub-surface. This project will integrate two approaches: interpretation of the structural geometry of a sub-surface lateral ramp that has been imaged in a 3-D seismic volume and penetrated by several wells, and physical (centrifuge) modelling in the Experimental Tectonics Laboratory at Queen’s University

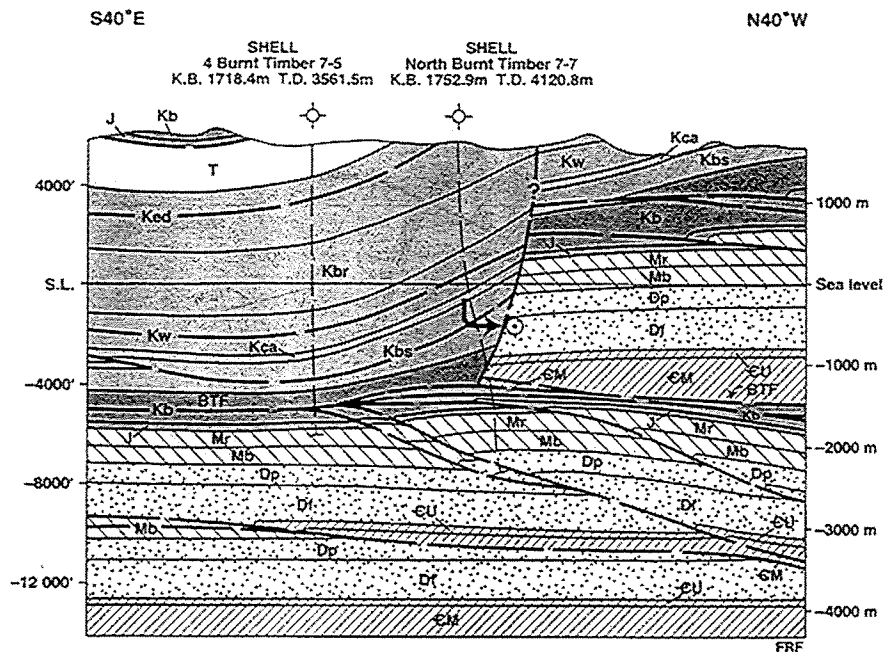


Figure 2. F.R. Frey's interpretation of the Marble Mountain "Swing-back" (as published in Fermor 1999). The heavy arrow points to the steep dip of the tear-fault. The transport direction is out of the page.

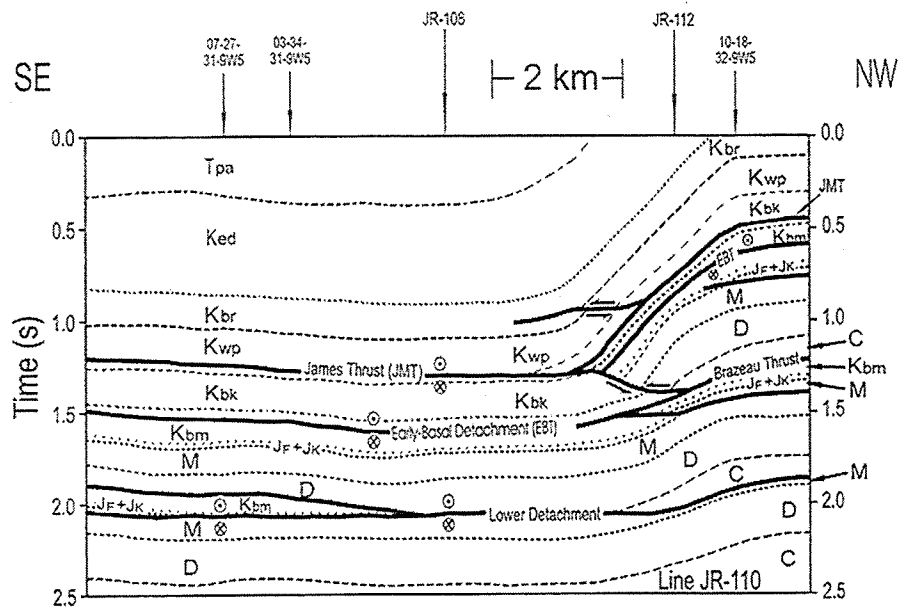


Figure 3. Longitudinal (strike) section across the Marble Mountain "Swing-back", interpreted by Bégin *et al.* (1996). Uplift at the NW end of the section is attributed to a lateral ramp. Transport direction is out of the page.

using models which accurately simulate the local mechanical stratigraphy. The solid-modelling and visualization software package, GOCAD®, will be used to compare the three-dimensional structure of the seismic interpretation and the analog models. Currently a preliminary interpretation of the seismic volume has taken place and some well log analysis has been carried out.

GEOLOGICAL SETTING OF THE STUDY AREA

The study area is located in the Rocky Mountain Foothills of southwestern Alberta (Figure 4). It is situated in the “Gap” map sheet (NTS 82G 16/W). Douglas (1950) produced a map as well as three cross-sections based on surface strike-and-dip measurements. Although the area was mapped more than fifty years ago, it still stands as the only current detailed map at 1:63 360 scale. The ongoing NATMAP project has not yet produced an update of the area at 1:50 000 scale.

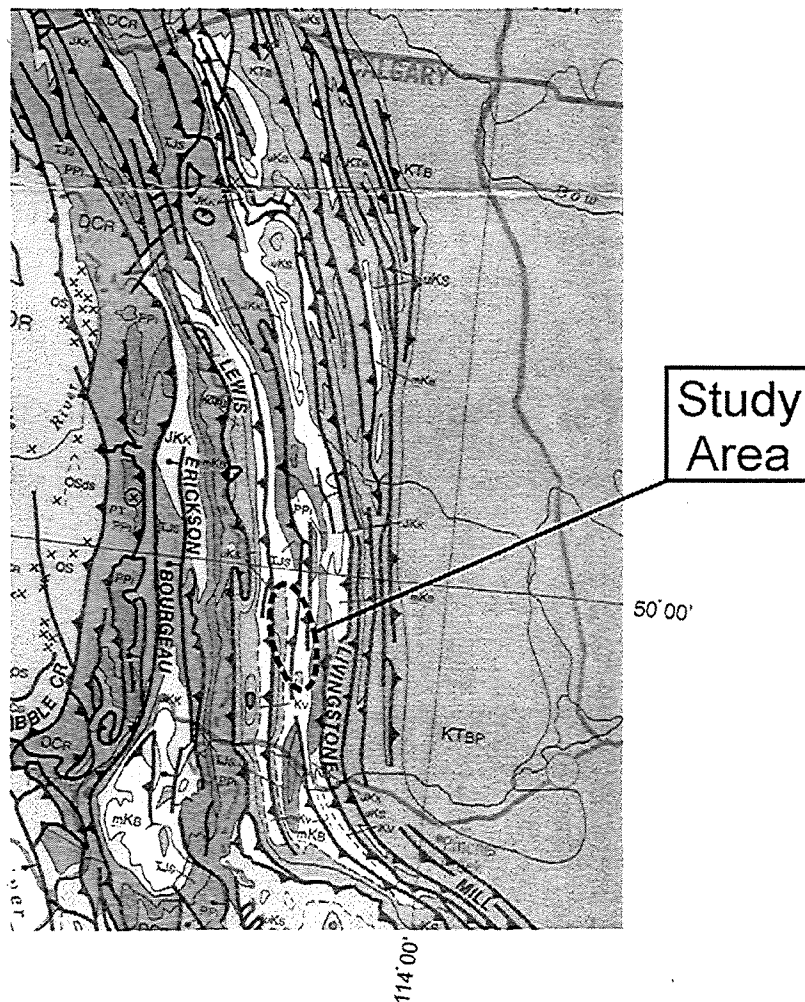


Figure 4. Location of the study area within the Canadian Rocky Mountain Foothills. Modified from Wheeler and McFeely 1991.

The 3-D seismic survey covers an area in which at least three thrust faults break the surface and one distinctive lateral ramp exists in the subsurface. The bedrock exposed at the surface includes units from the Lower Cretaceous Kootenay Formation through the Upper Cretaceous Belly River Formation. The Lower Cretaceous Blairmore Group is the most abundant map unit. East of the study area another large thrust brings rocks as old as the Mississippian Banff Formation to the surface.

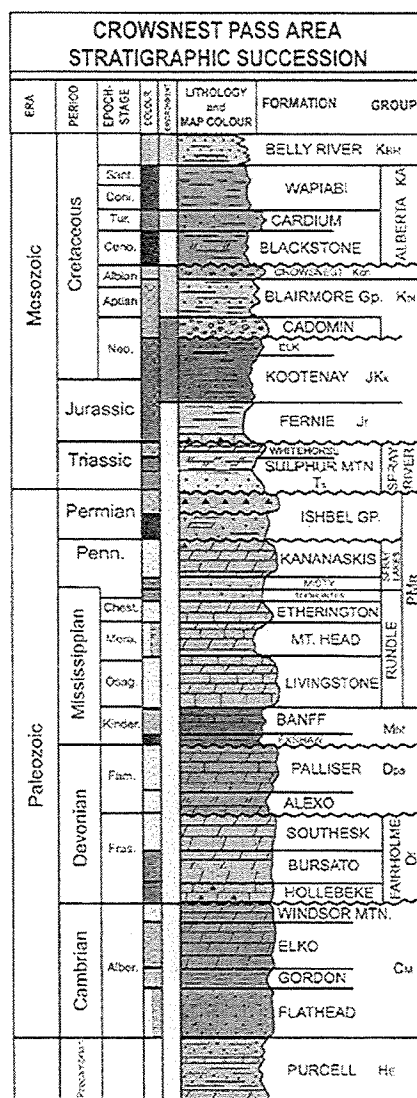


Figure 5. Stratigraphic section Crowsnest Pass area. Modified from Hamilton *et al.* 2000.

The stratigraphic column is illustrated in Figure 5. It consists of Cambrian sandstones, shales and dolostones unconformably overlain by Devonian, Mississippian and Pennsylvanian carbonates of the Fairholme Group, Palliser and Banff Formations, Rundle Group and Spray Lakes Group. The Spray Lakes Group is unconformably

overlain by the Mesozoic sequence of shales and clastics that include the Sulphur Mountain Formation through to the Cretaceous Belly River Formation.

The structural style of the area is dominated by thin-skin tectonics consisting of numerous imbricate thrust faults branching upwards from a common décollement. The deformation involves two major tectono-stratigraphic packages carried above an extensive basal detachment. The older package includes Paleozoic platform carbonates, which accumulated prior to the major collisional events that formed the Canadian Cordillera. The younger package comprises the Mesozoic detrital sequence of interbedded clastics and shaley rocks that were deposited as a result of tectonic uplift (Gabrielse *et al.* 1991). In this data set the footwall of the lateral ramp is made up of the near-basement autochthon which places this structure well below the Livingstone thrust. To the east a frontal ramp has developed at the same stratigraphic level.

CURRENT WORK

We are currently in the preliminary stages of interpreting the 3-D seismic volume from the study area. Some of the horizons that have been interpreted to date are shown in a dip line in Figure 6a and a strike line in Figure 6b. Data-quality issues and the locally discontinuous nature of reflectors have proven a challenge when following certain horizons. Figure 6a shows an example where the late time reflectors are affected by a velocity anomaly that has no expression in early time. This anomaly is present in all dip lines and completely obscures reflectors in the corresponding strike lines. Despite this anomaly it remains possible in most cases to correlate equivalent reflectors.

The seismic volume contains two prominent ramp features. A lateral ramp is visible in the longitudinal “strike” section (Z in Figure 6b) and an oblique frontal ramp can be found in the southeastern portion of the transverse “dip” sections (X in Figure 6a). The lateral ramp is oriented perpendicular to the local strike of the fold-thrust structures (*i.e.*, parallel to the inferred transport direction) and maintains a very low ramp angle to which a thrust sheet monocline conforms. The northern portion of the thrust surface follows a detachment flat while the southern portion ramps upward laterally and longitudinally. Therefore the structure is the footwall intersection between a trailing frontal ramp and a lateral ramp.

Further to the east, there is a frontal ramp (labelled X in Figure 6a) that is not related to the lateral ramp. This frontal ramp strikes at approximately 180°, that is, somewhat obliquely to the regional strike of the fold-thrust belt. The ramp angle is steeper than that of the previously described structure; however, due to its position in the seismic volume it is possible that the ramp angle is accentuated by the velocity anomaly.

No well data have yet been tied to the seismic volume. However, some initial work has been done to correlate the available dip-meter well logs.

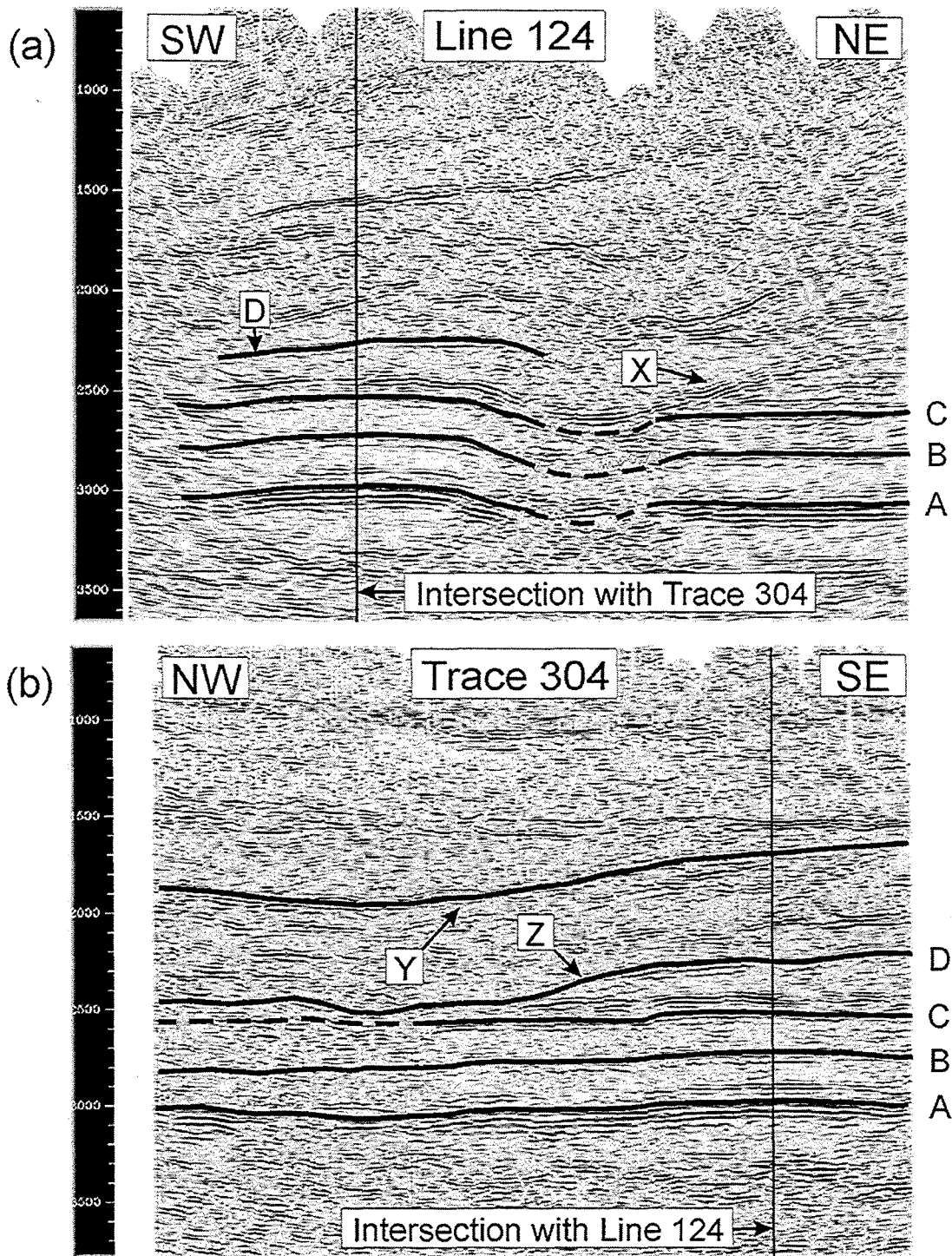


Figure 6. (a) Seismic Line 124 illustrating the autochthonous horizons A, B, & C as well as the lateral ramp forming horizon D and the oblique frontal ramp at position X. Note the velocity depression affecting the reflectors in the central portion of the line. (b) Seismic Trace 304 depicting the seismic expression of the lateral ramp structure (labelled Z). An upper reflector (Y) in the thrust sheet conforms to the lateral ramp.

PREVIOUS ANALOG-MODEL RESULTS

Dixon *et al.* (1998) presented results of an analog centrifuge modelling study of the structural effects of transport of a thrust sheet over a ramp system comprising two frontal-ramp segments linked by a tear-fault, a lateral ramp, and oblique ramps. The models were built with a generic mechanical stratigraphy and a pre-cut ramp system in the lowest stratigraphic unit (Figure 7). Two models of each configuration were built and deformed to allow sectioning in both dip and strike directions.

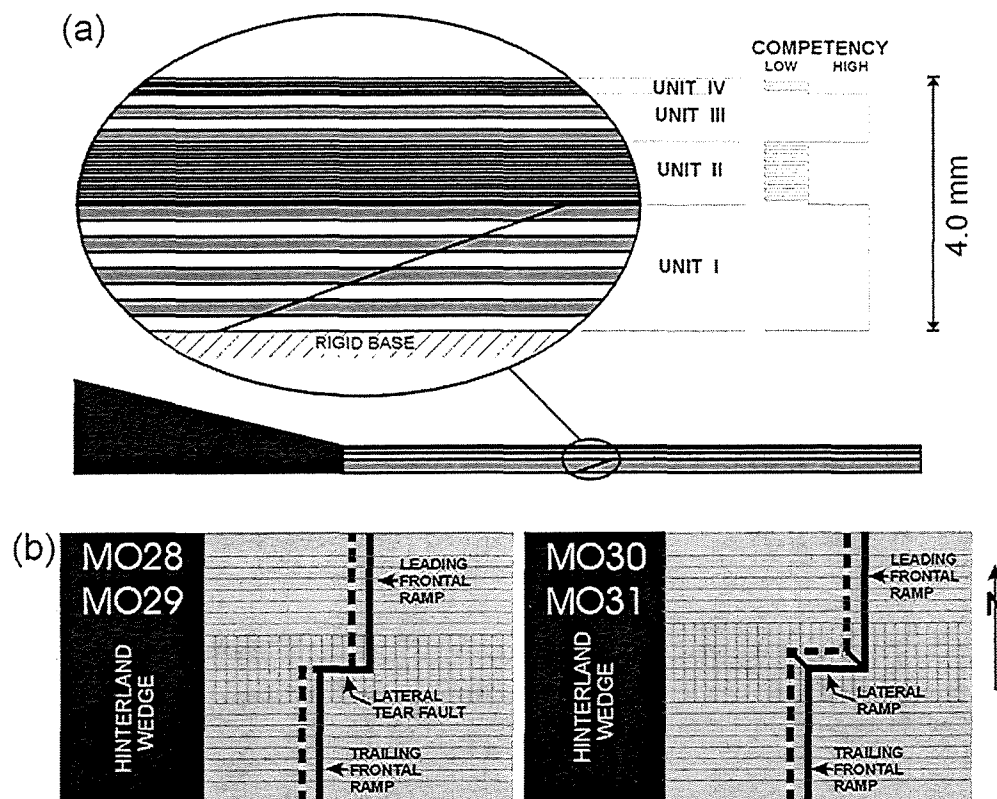


Figure 7. (a) Schematic transverse section of models MO28-MO31. Enlargement shows internal stratification and position of the trailing frontal ramp (a pre-cut discontinuity in unit I). Relative competency of units is shown by stratigraphic column at right. (b) Schematic plan views of the tear-fault and lateral ramp configurations, showing positions of ramp systems in unit I (top of ramp = heavy solid line; base of ramp = heavy dashed line). Light lines show position of transverse (W-E) and longitudinal (N-S) sections cut through the models after deformation. Thrust displacement is to the right (E). From Dixon *et al.* 1998.

The models revealed some unexpected structural features that developed in both the thrust sheet and footwall in the vicinity of the transverse link. One very prominent feature found in the lateral-ramp case (models MO30 and MO31, Figure 8) was out-of-plane transport and deformation located at the footwall intersection of the trailing and lateral ramps. As the thrust sheet moved up and over the intersection of the trailing frontal ramp and the lateral ramp, the structural culmination underwent gravitational

spreading which induced a fragment of the footwall panel to be displaced laterally, out of the plane of transport. The displaced fragment of the footwall can clearly be seen in section 40 of model MO31 (fwf in Figure 8a). It was also found that the lateral-ramp linkage between the two frontal-ramp segments was not stable. As deformation progressed, both the leading frontal ramp and the trailing frontal ramp began to propagate laterally through the model, and the thrust linkage evolved from a lateral-ramp linkage to a lap-joint displacement-transfer zone (Figure 8a). This propagation displays the instability of the lateral structure as it is effectively shut down and the displacement is preferentially transferred to the two frontal ramps.

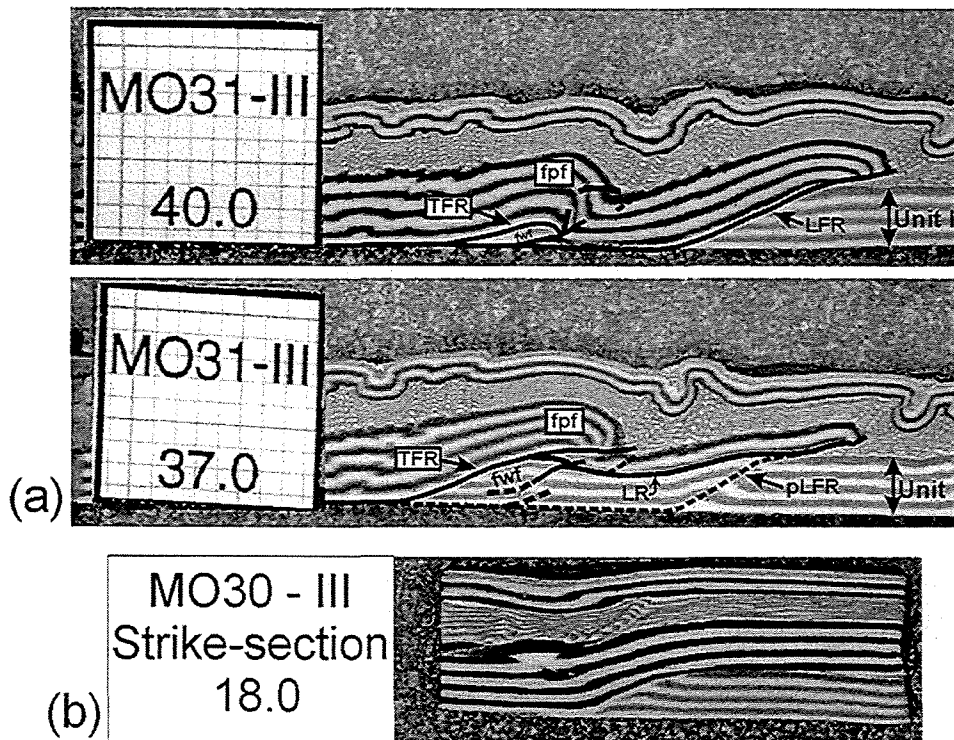


Figure 8. (a) Enlarged images of the central portions of transverse sections 37.0 and 40.0 through model MO31. The hanging-wall and footwall panels of Unit I are composed of different colours of plasticine (hanging-wall panel, dark grey and white; footwall panel, light grey and white). Black lines are interpreted structural boundaries: solid line = fault; dashed line = incipient fault or shear zone. LFR= leading frontal ramp; TFR = trailing frontal ramp; LR = lateral ramp; pLFR = propagating leading frontal ramp; fpf = fault-propagation fold; fwf = footwall fragment. (b) Longitudinal section from MO30 (twin to model MO31) showing the thrust sheet conforming to the lateral ramp in the footwall. From Dixon *et al.* 1999.

Dixon *et al.* (1999) continued the study of the MO series models with further investigation into the MO31 model utilizing the GOCAD visualization software package. Cross sections of the model were digitized and imported into GOCAD where a three dimensional model was built. Figure 9 demonstrates the deformation of the trailing footwall ramp intersection as well as the along-strike propagation of the leading frontal ramp into the footwall panel.

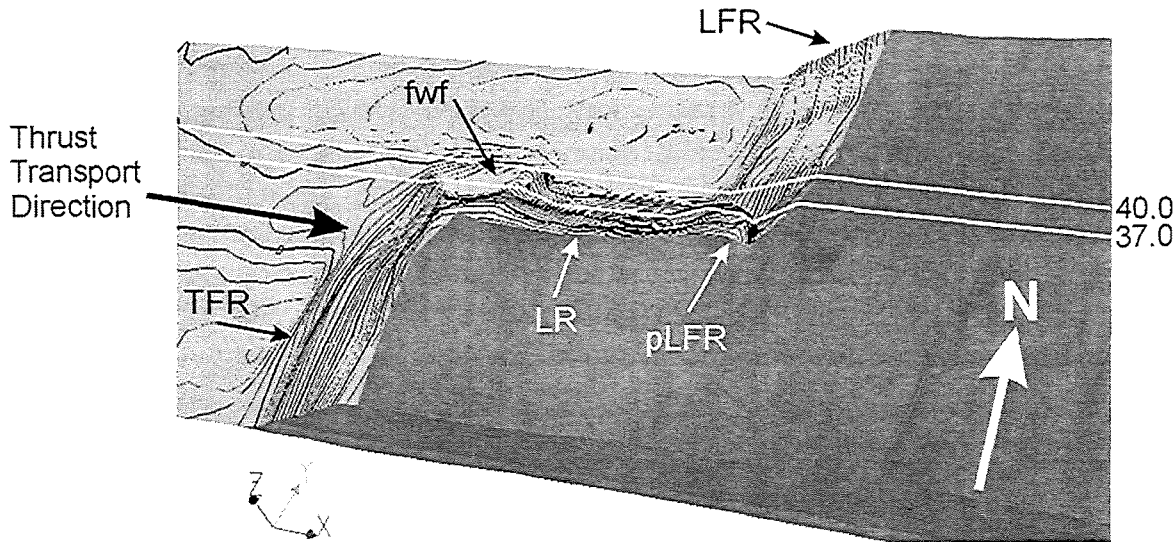


Figure 9. An oblique view of the thrust system in model MO31, rendered in a GOCAD model based on the sections shown in Figure 8a. View to the north (with foreland on the right). Lines show approximate positions of sections 37.0 and 40.0 (Figure 8a). The lateral ramp (LR) links trailing and leading frontal ramps (TFR and LFR, respectively). The leading frontal ramp is propagating laterally into the footwall as an overturned fold labelled pLFR. The corner of the footwall panel adjacent to the lateral ramp and the trailing frontal ramp (labelled fwf = footwall fragment) has been displaced both laterally and towards foreland due to footwall collapse. From Dixon *et al.* 1999.

These model investigations have provided insights into some structural complexities that may be expected in the vicinity of lateral ramps. We will be looking for evidence of these types of structures as we complete the interpretation of the 3-D seismic volume in the present study.

FUTURE WORK

The next few months will be devoted to completing and refining the seismic interpretation and tying it to the well data. Comparison of the post-stack data with the pre-stack migrated data will allow further revision of the seismic interpretation. The resemblance between the seismic sections presented above and photographed sections of analog models MO30 and MO31 (compare Figure 6 with Figure 8) demonstrate that one may use analog modelling to aid the investigation of these structural elements. The physical models of the MO series have provided insights into the types of structures that one may find associated with and in the vicinity of a lateral ramp. The interpretation of the seismic volume and the future structural investigation will attempt to verify their existence and tectonic role within this area.

Once the seismic interpretation has been completed, we will embark on a program of analog modelling designed to test the interpretation. The previous physical models of the lateral-ramp and tear-fault scenarios (Dixon *et al.* 1998, 1999) were constructed with a “generic” four-unit mechanical stratigraphy (Figure 7). In the new work we will construct

models which closely simulate the mechanical stratigraphy of the study area (Figure 5). It is hoped that the models will develop structures broadly representative of the large-scale features in the seismic interpretation, but it is also possible that more local features in the models may assist us in further refining the seismic interpretation. The model study may also help us to identify the circumstances that lead to development of the lateral ramp as a link between frontal-ramp segments.

We will use the GOCAD software package to build digital renderings of both the 3-D seismic interpretation and the physical models.

ACKNOWLEDGMENTS

This work constitutes part of the M.Sc. project of David Deline under the supervision of John M. Dixon and with the aid of R. Gerhard Pratt at Queen's University. We thank all the industrial sponsors of the Fold-Fault Research Project as well as the Natural Science and Engineering Research Council of Canada (NSERC) for their continued financial support. We are particularly grateful to NorthStar Energy Corporation for contributing the 3-D seismic data set and dipmeter well logs, and to Paul MacKay of Paul MacKay Geoconsultants Ltd. for providing the groundwork for the interpretation.

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The Bob Lake assemblage, amphibolite-grade rocks atop Proterozoic basement in the Nicola horst: the Ancient Margin gets wider beneath "Quesnellia"

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The Nicola Group, a Triassic volcanic assemblage with arc characteristics, has been included in the most inboard of the proposed accreted terranes of the southern Canadian Cordillera, the "Quesnel terrane" or "Quesnellia". Previous interpretations of crustal seismic profiles hypothesized that the low-grade arc crust of "Quesnellia" is more than 20 km thick around the Nicola horst, an inferred Cenozoic uplift in the Intermontane Belt approximately 20 to 60 km south of Kamloops.

Recently uncovered evidence of Proterozoic crust beneath a thin veneer of Triassic strata to the west of the Okanagan Valley (Erdmer et al. 2001) is incompatible with the interpretation of "Quesnellia" as an intra-oceanic arc. Accordingly, we set out to test the notion that the Nicola Group was deposited on cratonic basement, possibly the ancient margin of North America.

Evidence from the Bob Lake assemblage

Mid-crustal rocks, deformed and metamorphosed to amphibolite grade, are exposed in the Nicola horst and informally named the Bob Lake assemblage. This succession includes andesite to dacite porphyry, associated intermediate to felsic fragmental volcanic rocks, quartzite- and chert-clast conglomerate, pelitic schist, quartzite and other minor rock types. Sm-Nd analysis of the matrix of a sample of metaconglomerate yielded an ϵNd value of -4.5 (for mid-Jurassic time) and a model age of 1.44 Ga. Other samples of the

metasedimentary rocks yielded similar ϵNd values. These indicate the contribution of an evolved cratonic source during sedimentation.

Garnet, staurolite, andalusite, and minor kyanite, and sillimanite are present in metapelite of the Bob Lake assemblage, together with altered pseudomorphs of unidentified early porphyroblasts with rotated or discordant inclusion trails; porphyroblast-matrix relationships indicate at least two periods of metamorphic grain growth, a synkinematic one and a static one. The Bob Lake assemblage was intruded by the Clark Creek metatonalite (228.0 ± 0.6 , U-Pb zircon; Moore et al. 2000) which also intruded a nearby metadiorite-metagabbro pluton of assumed early Triassic age. Both plutons as well as other nearby, variably strained mafic plutons are likely subvolcanic elements of the arc represented by the Nicola Group rocks.

A strained felsic porphyritic intrusion that cuts black pelitic schist of the Bob Lake assemblage has yielded a U-Pb zircon date of 160.8 ± 0.6 Ma. It also yielded an ϵNd value of +3.0 and a model age of 0.77 Ga. The U-Pb zircon age is in agreement with a date for a previously sampled metatonalite or andesite porphyry which is assumed to be correlative (158.3 ± 0.6 Ma, U-Pb zircon, Moore et al. 2000). Similar dates within error were yielded by (1) a "cobble" of quartz-feldspar porphyry in metaconglomerate - interpreted as a fragment of extended dyke, and (2) zircon fractions from a whole rock sample of metaconglomerate. An

unidentified but highly distinctive black stain or coating, assumed to be graphite, occurs on some of the zircon grains from the ~160 Ma intrusion and on some of the detrital grains, suggesting a common origin. It is interpreted as a record of local interaction of the porphyry magma with the black schist during intrusion.

The ~160 Ma intrusion displays the north-westerly striking schistosity and westerly plunging stretching lineation affecting the Bob Lake assemblage as a whole. In the metapelite, the early metamorphic mineral growth is synkinematic with respect to the stretching fabrics. The late static metamorphism is attributed to thermal overprinting by local unstrained Paleocene plutons (Rocky Gulch granodiorite, 65.1 ± 0.9 Ma, U-Pb titanite; Frogmoore granodiorite, 64.5 ± 0.1 Ma, U-Pb zircon and titanite; Moore et al. 2000). Therefore, the ductile strain and mid-crustal metamorphism have an age between ~160 Ma and ~65 Ma. In contrast to this, the surrounding rocks of the Triassic Nicola Group (222.5 ± 1.4 Ma, U-Pb zircon; Moore et al. 2000) outside the bounding faults of the horst are unstrained and at low greenschist grade.

A cobble of fine-grained granite in metaconglomerate of the Bob Lake assemblage yielded a concordant U-Pb zircon date of 1038 ± 9 Ma. Sm-Nd analysis of the cobble yielded an ϵ_{Nd} value of -3.2 and a model age of 1.09 Ga. In the whole rock sample of metaconglomerate, one of the recovered zircon grains yielded a concordant date of 1030 ± 40 Ma. Another zircon grain was concordant at 434 ± 19 Ma. Other zircon fractions showing greater discordance were also present. We interpret the oldest dates as evidence that local Proterozoic crust was intruded by a felsic pluton near ~1040 Ma and was at the surface at the time of deposition of the Bob Lake assemblage. The interleaving of intermediate fragmental volcanic rocks and metaconglomerate with quartzite clasts in the Bob Lake assemblage records the juxtaposition

of strongly contrasting crustal sources, i.e., of possible arc affinity for the metavolcanic rocks and of cratonic or pericratonic affinity for the metasedimentary rocks. The depositional age of the Bob Lake assemblage is unknown. Direct dating of a sample of intermediate volcanic metaporphry is in progress. A minimum constraint is provided by the Clark Creek metatonalite (~228 Ma). We interpret the Nicola Group to overlie the Bob Lake assemblage stratigraphically, and to be between a few hundred metres and a few kilometres within the horst.

Implications for Cordilleran tectonics

The presence of Proterozoic cratonic basement and of pericratonic cover of pre-Nicola age, in a region that was interpreted to be thick arc crust of "Quesnellia", supports the hypothesis that pre-Cordilleran rocks of ancient margin affinity constitute most of the crustal lithosphere between the Okanagan Valley and the Fraser fault. That hypothesis was originally advanced on the basis of similar evidence from Spa Creek, approximately 20 km west of the Okanagan Valley (Erdmer et al., 2001). These data support the inference of a regional stratigraphic contact rather than the existence of an allochthon.

The interpretation of a depositional contact of "Quesnellia" with North American cratonic rocks does not preclude the development of locally thick Triassic arc successions as observed in parts of the Nicola Group or incipient oceanic rifts containing older strata, expected in the general setting of a heterogeneously stretched "margin".

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DEFLECTION OF THE LA BICHE ANTICLINE, YUKON AND NORTHWEST TERRITORIES.

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The La Biche Structure, located in the southern Franklin Mountains, involves Devonian and Mississippian strata at the surface and displays a marked change in trend and geometry of folds and related faults along strike (Figure 1). At 60°40'N, the structural trend deflects from NNW-SSE to approximately N-S. Fieldwork between June and August 2000 initiated a study of the La Biche Anticline which will integrate surface and subsurface data to better understand the structures, and their controls, in the study area. This change in structural trend correlates to an apparent dextral rotation exhibited in structures of the Liard and Kotaneelee Ranges along the Beaver River Structure (after Morrow and Potter 1998, later discussed by Morrow and Miles, 2000). The cause of this change in trend will be investigated and discussed. Following summer field work, it is noted that there is a change in dominant vergence direction of the structures across this deflection line. North of the deflection line, vergence of structures is to the east, whereas south of the deflection line, vergence is towards the southwest. The style of folding also changes from a box or kink fold geometry in the south, to a more broad and open fold style north of the deviation line.

Figure 2 shows the study area divided into four domains of consistent structural trend. Structures in Domain 1 plunge 7° toward 007°. Structures in Domain 2 plunge 12° toward 320° whereas in Domain 3 structures are trending 327°-147° with little or no plunge. The trend difference from Domain 1 to Domain 3 is 40°. Structures in Domain 4 plunge 6° toward 157° marking the beginning of another deflection.

Four possible causes of deflection of structures are briefly discussed and their applicability to the La Biche Anticline is considered. Deflection of surface structures can occur by 1) strike-slip basement faulting oblique to the surface trend occurring syn or post-Laramide deformation, 2) deflection across a facies boundary, 3) pre-existing topographic highs and lows in basement affecting structural propagation up-section, 4) pre-Laramide vertical motion of basement blocks creating deflections of overlying strata during compression.

Dextral strike-slip motion of oblique basement faults (as proposed by Morrow and Miles, 2000) would result in a consistent offset or rotation of surface structures along the fault and in overlying structures. Sigmoidal warping of structures may well occur if there is ductile shear or syn-Laramide compression, but is not expected if displacement along basement faults is post deformation, particularly in the more brittle surface stratigraphy. In the study area, an extension of the west-dipping limb (Domain 2, Figure 2) crosses the line of deviation without a change in orientation. It is also noted that the Fantasque

syncline to the west of the study area does not exhibit any deflection across the Beaver River Structure.

Deflection across a facies boundary oriented oblique to the compression direction, such as a carbonate bank to shale, would result in jogs in structures along the margin. Structures in the less competent units, such as shales, will propagate further into the foreland than the more competent carbonate unit (Fallas, 1994). An increase in displacement would therefore be expected north of the deflection line (the proposed shale-carbonate margin) if this is the case in the La Biche Range, where Paleozoic limestones may shale out towards the north. At this time, subsurface data relating to deeper Paleozoic strata are not available because of little economic exploration in the area, so their nature is therefore poorly understood.

Pre-existing basement topography can also deflect structures or control overlying geometries at the surface. As interpreted by Morrow and McLean (2000), the Pre-Phanerozoic unconformity is not simply a gentle west-dipping monocline, but rather an undulatory surface that may control overlying structures at the surface. Topographic ridges or highs in the basement may initiate folding or fault propagation. It is noted that the Liard Range and La Biche Range both follow aeromagnetic contours at some locations, suggesting that basement topography may have an influence on structures (this is assuming that the aeromagnetic patterns are at least partially related to basement highs and lows).

Vertical uplift of blocks such as the models discussed by Hodder et al. (1998) illustrate how a rigid basement block can deflect structures around it. These vertically moving blocks affect sedimentation and thus can be the cause of facies changes. Hodder et al. (1999) documented major differences in Paleozoic strata thickness across the Liard Line, and later interpreted that this was due to upward vertical movement of the southern block relative to the northern block. The eastward jog of structures north of the Liard Line, and the subsequent thickening of Paleozoic strata on the north side of this boundary, provide evidence to support vertically moving blocks having a role in deflecting structures. Because of evidence of vertically shifting blocks, and the proximity of the Liard Line to the study area, a vertical component of basement block movement should be considered at the deflection in the northern La Biche range.

Upon consideration of field observations, aeromagnetic data, and stratigraphic relationships, it is suggested that the apparent deflection is due to a combination of the aforementioned models, with vertically moving basement blocks playing a key role in influencing surface structures during compression.

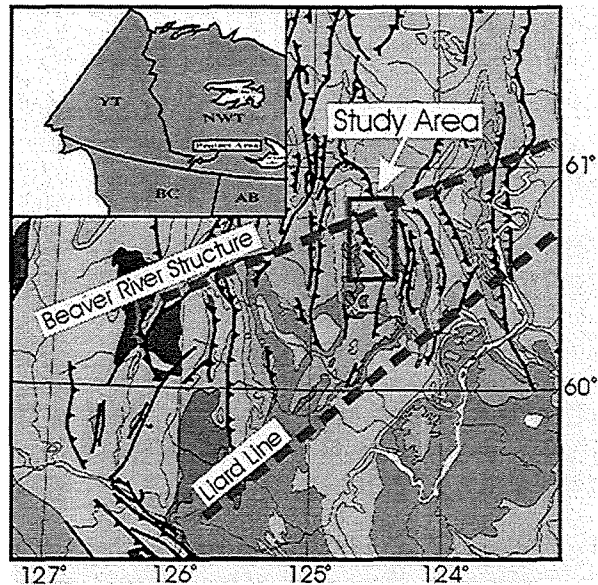


Figure 1. Location of study area and its relation to the Beaver River Structure (Morrow and Potter, 1998) and the Liard Line (Cecile et al. 1997). Map modified from Wheeler and McFeely (1991).

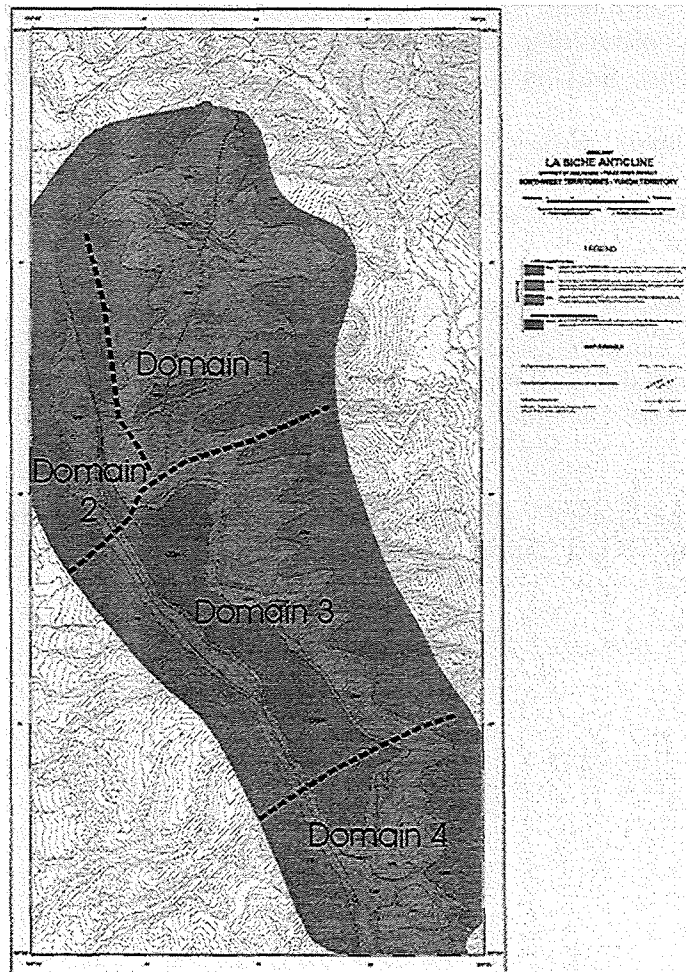


Figure 2. Structural domains in the La Biche Anticline study area. Approximate location of Beaver River Structure plotted as thick dashed line.

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U-Pb geochronology and geochemistry of Permian and Triassic intrusions and volcanics in southwestern British Columbia: Implications for tectonic setting and terrane correlations

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Previously unrecognized Late Permian granitoids and a suite of Middle-Late Triassic intrusions and volcanic rocks are exposed in fault panels between the Yalakom fault and the Jurassic-Tertiary Coast Plutonic Complex in SW B.C. The Late Permian, highly altered Tatlayoko granitoid (254 Ma) and the dioritic to tonalitic Mt Skinner plutonic complex (227-231 Ma; $n=2$) are nonconformably overlain by Upper Triassic clastics of the Cadwallader terrane, which are overlain by Lower to Middle Jurassic clastics and volcanoclastics correlated with the Methow terrane and Jura-Cretaceous clastic rocks of the Tyaughton-Methow basin. In a separate fault panel, a latest Permian to earliest Triassic aplite dyke (251 Ma) intrudes marine clastic rocks that may represent part of the Paleozoic basement of Cadwallader terrane. The Crazy Creek pluton (217-220 Ma; $n=2$) is thrust imbricated with, and locally nonconformably overlain by Upper Jurassic clastic sedimentary rocks of the Tyaughton-Methow basin. The Niut pluton (212, 219 Ma; $n=2$) and Sapeye Creek pluton (207 Ma) intrude Ladinian to early Norian mafic to intermediate volcanics correlated with the Pioneer Formation of Cadwallader terrane.

Geochemically, intrusive rocks define a low-K trend and are metaluminous to mildly peraluminous; trace element affinities are transitional between primitive arc tholeiites and calc alkaline rocks. Volcanics are more diverse, with ocean ridge basalts, arc tholeiites and calc alkaline rocks represented.

Age, geochemical and field relations of Triassic igneous rocks agree with regional relationships which suggest that the Cadwallader and Methow terranes were part of a single Middle Triassic to Middle Jurassic arc-basin system that developed above oceanic crust (Early Permian Shulaps ophiolite complex?). This arc system, together with the associated Bridge River subduction-accretion complex to the west, was overlapped by the Tyaughton-Methow clastic basin in late Middle Jurassic time. The Cadwallader-Methow arc system bears a strong resemblance to the Late Triassic Stuhini/Lewes River and Jurassic Hazelton magmatic arcs of Stikinia, which underlie much of the Intermontane belt to the north. The Cadwallader-Methow arc may be a southern continuation or a faulted fragment of the Early Mesozoic arcs of Stikinia. The Late Permian intrusions may correlate with a recently recognized discontinuous belt of Late Permian - Early Triassic volcanic and plutonic rocks situated adjacent to the Cache Creek terrane and characterized by primitive bimodal arc tholeiites. Permo-Triassic and Upper Triassic arc-related volcanic and plutonic rocks similar to those described here from Cadwallader terrane also characterize the Wallowa terrane of Oregon and Idaho.

GEOLOGICAL MAPPING IN THE KLINKIT AND SWIFT RIVER ASSEMBLAGES, NORTHERN BRITISH COLUMBIA

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Preliminary results of 1:25,000 geological mapping of the Klinkit and Swift River assemblages (see Harms and Stevens, 1996a and b) in map sheet 104-O/11 (Klinkit Lake) of northern British Columbia are reported.

Although local facies variations are common within the map area, a consistent sequence of units in the Klinkit assemblage can be identified and informally named as follows (base to top): Silvertip Quartzite - massive, thick-bedded pale quartz sandstone to arkose with two prominent horizons of well-bedded carbonate near the top; Screw Creek Limestone - thick (≤ 40 m) massive, fossiliferous limestone, late Carboniferous in age (GSC Report No. MJO-1997-7); Butsih Volcanics - Basaltic to andesitic, generally green, dominated by fine volcaniclastics and tuff, with aphanitic flows, in places pillowed, and microdiorites. Includes a marker horizon of red banded chert; Transitional Unit - very variable in character, may include carbonate with abundant silicification, tuff, chert, black argillite, and/or carbonate conglomerate; Teh Clastic Unit - interbedded black shale, argillite, siltstone, and quartz sandstone in variable proportions with finer grain sizes generally predominating. Sandstone has dark, glassy, spherical grains in a quartz matrix. Minor interbedded chert or fetid carbonate is characteristic. Interbedded conglomerates are a minor component. Mid-Late Triassic in age (GSC Report N. MJO-1997-7). No subdivision of the very heterogeneous Swift River assemblage has been attempted, but the assemblage contains the following rock types: black slate to silvery phyllite, banded and massive recrystallized chert, greywacke, and minor diorite. The Swift River assemblage has not been dated, and its age relative to the Klinkit assemblage is indeterminate. Geochemical and isotopic analyses of clastic rocks from the two assemblages have been previously reported (Creaser and Harms, 1998).

The Klinkit assemblage occurs in map-scale folds with overturned limbs. In the absence of marker-units, folds of this scale cannot be identified in the Swift River assemblage. Several sets of close to tight, upright, outcrop-scale folds and hand specimen-scale crenulations are abundant in the Swift River and phyllitic fabric is common in its more pelitic rocks. These structures characterize the Swift River assemblage and contrast to the comparatively less deformed Klinkit.

The Klinkit assemblage structurally overlies the Swift River assemblage in this map area. Outcrop-scale folds in the Swift River are discordant with the contact between the two assemblages and may be truncated by the contact. Different units of the Klinkit assemblage, ranging from the basal Teh quartzite to the youngest Silvertip clastics, overlie the contact in various localities in the map area. The contact between the Klinkit and Swift River assemblages may therefore be an angular unconformity, a fault, or a faulted unconformity. The contact has not been directly observed in this map area.

The Klinkit and Swift River assemblages are intruded by a variety of both Jurassic and Cretaceous granitic plutons (Gabrielse, 1970). The contact with the Jurassic Simpson Peak batholith is locally mylonitized.

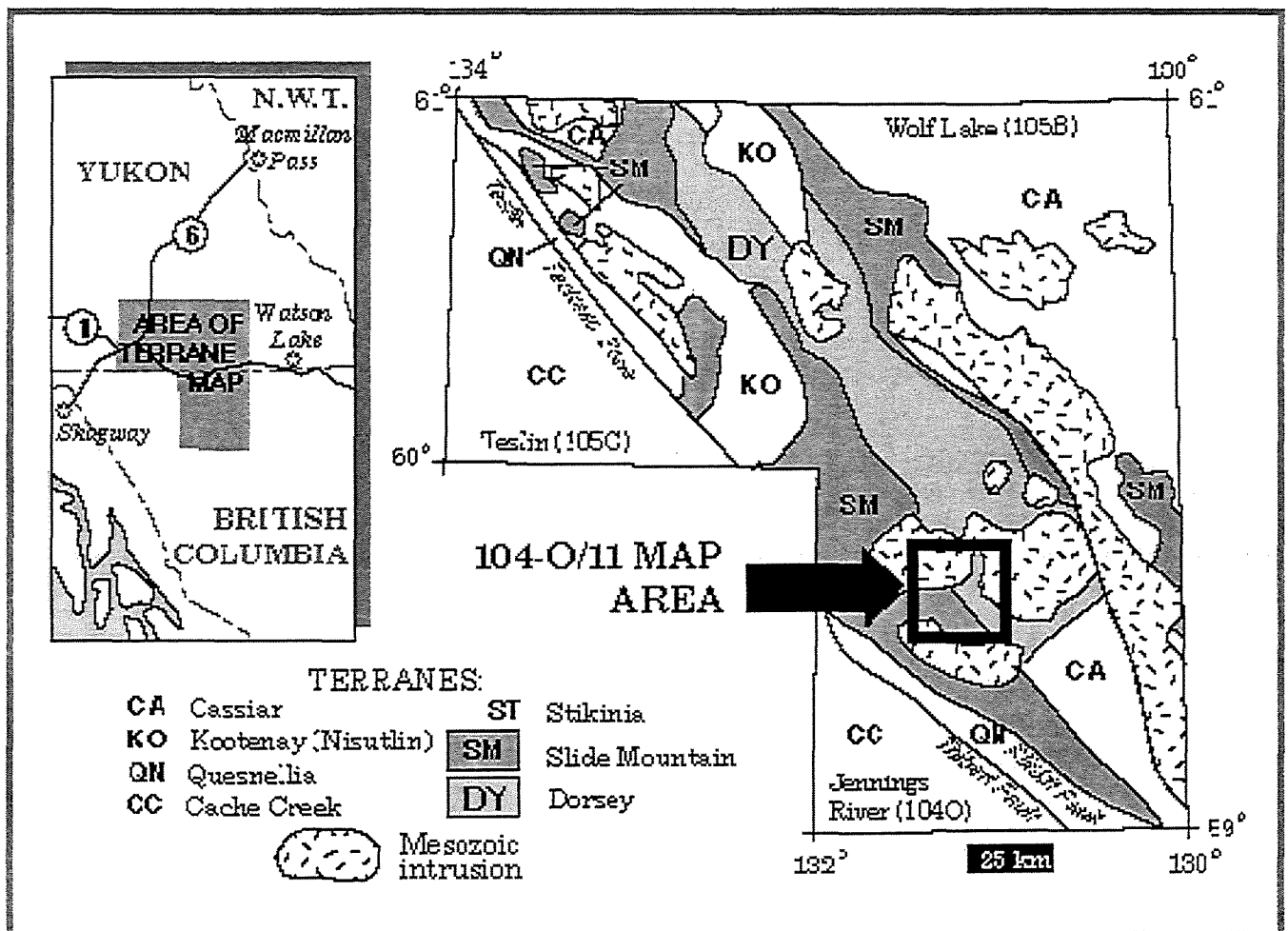
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Location of the 104-O/11 map area. Terrane map adapted from Wheeler and McFeeley (1991).

**New developments in compiling a balanced cross-section through
Pink Mountain Anticline, Northeast British Columbia:
Insights into the nature of the deformation front**

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As part of the Fold-Fault Research Project (University of Calgary) and the Central Foreland NATMAP Project (GSC Calgary), a M.Sc. thesis on the Pink Mountain Anticline was initiated to study its geometry and structural evolution in detail. The Pink Mountain Anticline (approximately 57° N latitude, 123° W longitude) is an anomalous topographic high immediately in front of the physiographic foothills. Due to limited surrounding surface rock exposure, the Pink Mountain Anticline was interpreted by previous authors to be a single upright anticline that trends and plunges to the northwest. We correlated the surface and subsurface configurations of the Pink Mountain structure using new geology maps, digital well logs, structural measurements, and seismic lines (seismic data provided by Shell Canada, Conoco Canada and Petro-Canada).

The well formation tops for the Pink Mountain area were revised based on the Triassic interpretations of Dixon (1999). The Devonian-Carboniferous and Cretaceous formation tops were adjusted to accommodate the new Triassic interpretations. Using the new tops as a guide, décollement zones within the Devonian-Carboniferous Besa River Formation were interpreted in the seismic lines in the Pink Mountain area. The Besa River Formation has been identified previously as the principal regional décollement horizon in the Halfway River (94B) map area, just south of the Pink Mountain area and in the foothills of the Trutch map area (94G) to the west of the study area.

This initial seismic analysis shows that the shallow subsurface structural deformation in the study area is mostly fault dominated. The Pink Mountain structure consists of a succession of thrust-bounded pop-ups that produced the topographic high in front of the Rocky Mountain foothills. The main pop-up structure contains the Pink Mountain Anticline plus subsidiary folds separated by thrusts and backthrusts east and west of Pink Mountain.

Recently, the best quality seismic section used in the initial analysis was pre-stack depth and time migrated using five wells and a velocity model as a guide. Initial processing results show the Pink Mountain Anticline to be a part of a series of detachment folds over a thrust fault that forms a décollement within the Besa River Formation. This reprocessed seismic section indicates shallow subsurface detachment folding is the dominant structural style at the edge of the Rocky Mountain Deformation Front in the study area.

Dixon, J., 1999. Isopach maps of Triassic units in the Trutch (94G) map area (Central Forelands NATMAP). Geological Survey of Canada Open File 3765, 21 p. 9 maps.

Late Proterozoic to Mesozoic structural evolution of the northern termination of the Purcell Mountains, southeastern British Columbia.

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Strata of the late Proterozoic Horsethief Creek Group (HTC), Lower Cambrian Hamill Group, and Lower Cambrian Donald Formation, near the south end of Kinbasket Lake (Blackwater Ridge and Esplanade Range), southeast British Columbia, comprise the northern termination of the Purcell Mountains, and the northern termination of Dogtooth Range structures. These strata are imbricated on thrusts, and deformed into relatively tight folds, that plunge to the southeast at angles of $\sim 20\text{--}45^\circ$, while the overall regional plunge is $5\text{--}15^\circ$. The local plunge variation may reflect underlying thrust ramps and local shear within the thrust panels. Strata in the eastern Esplanade Range form west dipping thrust panels of chlorite to garnet zone HTC meta-pelite and semi-pelite, with significant internal ductile deformation evidenced by tight folds, limb attenuation, and sheared zones. The easternmost thrust faults on Blackwater Ridge are themselves folded. These thrust faults and folds occur in the immediate hanging wall of the out-of-sequence, pre-mid Cretaceous Purcell thrust fault, and they are truncated by it. Post mid-Cretaceous open buckle folds, and kink folds are oblique to the general structural trend, and locally fold the Purcell fault. Steep northwest-southeast and east-west striking faults are likely of late Cretaceous to Tertiary age. These faults truncate compressional structures, and include a re-activated early Cambrian growth fault.

Lithofacies within the Lower Cambrian Hamill Group are directly correlative with the Jasper member and Poker member of the Lower Cambrian McNaughton Formation (Gog Gp.). The facies, their thickness relationships, and underlying unconformities are continuous with those of strata directly to the east, in the western Rocky Mountains. Furthermore, the lithostratigraphic relationships that outline the late Proterozoic-early Cambrian Kinbasket High, cross from the northern termination of the Purcell Mountains, directly to the east into the western Rocky Mountains. These observations corroborate previous findings, that demonstrate that any large scale post mid-Cretaceous right lateral motion focussed in the Southern Rocky Mountain Trench must decrease to less than ca. 5 kilometres by the latitude of the study area (ca. $51^\circ 45'$). This is incompatible with suggestions that post mid-Cretaceous right lateral strike slip faulting in the Southern Rocky Mountain Trench translated exotic terranes hundreds of kilometres northward (e.g. Chamberlain and Lambert, 1985; Pope and Sears, 1997).

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Preliminary report on ichthyolith studies from the Baldonnel and Pardonet formations, Trutch and Halfway River map-areas, northeastern British Columbia.

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Ichthyoliths (microscopic fish teeth and scales) are a new tool for the development of a tectono-stratigraphic framework for Triassic rocks and future hydrocarbon exploration in northeastern BC. Of significance are marker beds with abundant ichthyoliths that are widespread, correlative, and may be associated with high gamma ray signatures. The Baldonnel and Pardonet formations have been central to the study. Additional samples have been analyzed from the Liard Formation. The project has been supported through the LITHOPROBE SNORCLE Program and the GSC's Central Foreland NATMAP Project. Ichthyolith research progress is summarized below. The following results and interpretations are preliminary.

- **Collections & future plans.** Field collections in 1998, 1999 and 2000 have resulted in analysis of over 635 Triassic samples in the Trutch (94G) and Halfway River (94B) map-areas of northeastern BC. Pre-1998 collections have included over 170 samples. Plans for 2001 include completion of laboratory processing and microfossil picking (1999 & 2000 samples), compilation and integration of the ichthyolith (Johns, Barnes) and conodont (Orchard) data, and commencement of manuscript preparations.
- **Ichthyolith and conodont biostratigraphy.** NE BC Triassic ichthyolith diversity and abundance is greatest in the inner marine shelf to neritic environments (e.g. Baldonnel & Liard formations). Conodont diversity and abundance is greatest in the marine shelf to slope environments (e.g. Grayling, Toad, & Pardonet formations). Yet, both ichthyoliths and conodonts occur in each of the environments so that an integrated dual biostratigraphy can be developed. Combining ichthyolith and conodont studies provides a powerful biostratigraphic tool spanning a diversity of sedimentary environments in the Triassic of NE BC. Incorporating macrofossil (ammonoid, bivalve, brachiopod) data further enhances the biostratigraphic framework.
- **Pardonet Fm. ichthyoliths.** Ichthyolith distribution anomalies were first observed in the Pardonet Formation (shelf to slope environment). A typical Pardonet ichthyolith distribution has common to rare ichthyoliths in most carbonate samples. However, ichthyolith abundance dramatically increases at four levels in the Pardonet. Three of these levels (lower, lower middle, and middle Pardonet) are widespread (occurring at several section locations in the Peace River Embayment) and each may record the top of a shallowing-up cycle. To date, the fourth level only has been observed at Moose Lick Creek ridge (94G) near the base of the *Monotis* beds (upper Pardonet). This bed contains a new and abundant ichthyolith assemblage that lies within another new and different ichthyolith assemblage, and may represent a local lag deposit or a channel-cut.
- **Baldonnel Fm. ichthyoliths.** The Baldonnel Formation contains many different ichthyoliths and abundance levels. The Baldonnel represents a transition from the terrestrial to tidal environments of the Charlie Lake Formation to the outer marine shelf facies of the Pardonet Formation. The variety of ichthyolith distributions reflect the many changing environments during this transitional period. Sea level changes appear to be frequent in the lower Baldonnel/upper Charlie Lake and then go through at least two (possibly 3) cycles in the middle and upper Baldonnel.

- ***Marine environment & ichthyolith distribution.*** Ichthyoliths are diverse and frequently abundant in shallow water environments. Commonly ichthyolith diversity but not necessarily abundance declines in deeper waters. Widespread bone beds are present in the Baldonnel and Pardonet. These could have been produced either by improved oceanic conditions (upwelling, oxygenation) resulting in increased productivity and diversity or during periods of highstand with low sedimentation and deposition of lag deposits. Ichthyoliths may be concentrated 1) at the contact of two sedimentary facies (several examples in the Baldonnel) 2) in local lag deposits such as from wave action, currents, or erosion (possibly in the upper Baldonnel); 3) in a channel base (possibly the upper Pardonet); and 4) at highstand or during slow sedimentation resulting in condensed units (especially in the Pardonet & in some Baldonnel beds).
- ***Pattern of ichthyolith assemblages in a flooding and shallowing-up sequence.*** Ichthyolith diversity and abundance are commonly greatest at the top of a shallowing-up sequence. Several different assemblages occur when sea levels and ichthyolith/fish environments change. A final concentrated ichthyolith assemblage at a contact conglomerate or lag may contain ichthyoliths common to the lowered sea level condition, reworked ichthyoliths from older sediments, and ichthyoliths typical of the returning flood environment. In NE BC, the flood-rising to flood-maximum environment is commonly marked by a low diversity of ichthyoliths dominated by *Birgeria*, a few other bony fishes, and rare elasmobranchs. Conodont abundance and diversity are typically high. Flood-rising to flood-maximum events in a carbonate depositional system are commonly characterized by a basin-wide flooding sedimentary record, the return or increase of carbonate generation, deeper-water faunas, and slow deposition or condensed sections (discussed in Miall, 1997, p. 58-65, 81, 352-355, 376). In addition, sediments may contain abundant bone and phosphatic material and show a drowning unconformity as the transgression rises over a reef or part of the shelf passes below the photic zone (Miall, 1997, p. 17, 63). Baldonnel and Pardonet ichthyolith data from Pink Mountain and other locations include examples of flooding followed by a shallowing-up sequence.
- ***Baldonnel/Pardonet boundary.*** The top of the Baldonnel Formation in Trutch is marked by thick, commonly resistant (especially to the west), gray to light gray bioclastic limestone. The gamma ray signature is commonly low and the rocks contain a distinctive ichthyolith assemblage. The Pardonet begins with flaggy limestones to limy siltstones/shales. The bioclastic content of the limestones increases above the boundary, the rocks are commonly organic and have a fetid odour, bone beds are observed at certain levels, and the gamma ray signature is notably high. Ichthyolith-abundant beds of low diversity occur near the base of the Pardonet.
- ***Charlie Lake/Baldonnel boundary.*** The base of the Baldonnel Formation in Trutch could be interpreted as 1) the last major occurrence of Charlie Lake sandstones/limy sandstones (base of the upper two Baldonnel dark limestone units) or 2) the first marine flooding at the top of the Charlie Lake sandstones and evaporites. The former is the preferred mapping boundary because it is easier to recognize the base of the two dark gray limestone units that are 1) below the uppermost Baldonnel massive, bioclastic, light gray limestone unit and 2) above a sandstone to sandy limestone. The latter, is commonly in recessive strata. Ichthyolith assemblages can be recognized at or near both positions.
- ***Liard, Baldonnel & Pardonet ichthyoliths.*** Ichthyoliths are abundant and diverse especially near the top of the Liard Formation. Ichthyoliths from the lower Baldonnel show some affinities to those found in the Liard. Pardonet elasmobranchs are completely different from those in the Baldonnel and Liard.

- ***Gamma ray signatures & ichthyoliths.*** Some ichthyolith-abundant beds can be correlated with high gamma ray signatures especially in lag deposits, bone and phosphatic intervals. Ichthyolith-abundant marker beds are an important tool for surface to subsurface correlation.
- ***Unconformities.*** Ichthyolith-abundant beds near or at unconformities and facies boundaries at the top of the Liard and several within the Baldonnel may correlate with the Coplin, Boundary, Siphon, Laprise, and Jedney unconformities observed in the subsurface.
- ***Hydrocarbons.*** The changing environments in the Baldonnel could have important implications for oil and gas resources. Organic-rich source rocks may be generated at times of sea level flooding and/or in lagoonal facies. Associated sandstones, conglomerates, and dolomitic limestones are potentially important reservoir rocks.
- ***Ichthyolith thermal & other alteration.*** Organic material within ichthyoliths is known to change colour in response to rock/thermal heating, oxidation, and mineral staining. Review of samples throughout a section will help to determine variability. Some ichthyoliths in Baldonnel blocky to massive limestones in eastern Trutch are light in colour and may have been affected by oxidation. Some specimens from dark, flaggy limestones are darker perhaps resulting from 1) hydrocarbon or mineral staining or 2) fluid flow within these layers. Uncorrected ichthyolith colour values fall within the conodont colour alteration index of 1 to 4 (50-300°C) from samples in the eastern Trutch map-area.
- ***Ichthyolith rare earth & trace elements.*** Initial analyses (1990, 1992) of rare earth and trace elements within ichthyoliths from Liard, Baldonnel, and Pardonet Formation samples indicate element similarities between two levels in the Liard and Baldonnel and element differences in the Pardonet. Ichthyolith concentrations of strontium, yttrium, and uranium are significantly elevated in the Pardonet sample. Future analyses are planned to learn more about sediment source and offshore marine inputs (especially in the Baldonnel), and to apply regional chemostratigraphy.
- ***Sampling strategy.*** Extensive sampling has been important to assess the changing environmental conditions, to recognize ichthyolith marker beds, and to develop a refined biostratigraphy.

Summary.

A large database of Triassic ichthyoliths from over 800 samples has been assembled, mainly from recent fieldwork in the Trutch and Halfway River map-areas. The work is leading to a refined ichthyolith biostratigraphy that will be integrated with the existing conodont and macrofossil biozonations. The pattern of ichthyolith paleoecology is becoming clear and can be related to the diverse lithofacies in the region. Ichthyoliths also are concentrated at certain horizons (lag deposits, highstands, etc.) that are useful for surface and subsurface correlation, especially since many are characterized by high gamma ray signatures. Hence, they can be useful in determining sequence boundaries and regional unconformities. Regional thermal alteration patterns from the ichthyoliths and conodonts suggests that many of the strata could lie in the oil to dry gas windows. Additional work is planned on ichthyolith geochemistry to aid in regional chemostratigraphy and to better interpret the depositional environments.

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This abstract was presented (talk and poster) at the Central Foreland NATMAP Project 2000 fall workshop.

GEOLOGY OF THE FORT LIARD REGION, YUKON AND NORTHWEST TERRITORIES: INITIAL RESULTS OF NEW MAPPING BY THE CENTRAL FORELAND NATMAP PROJECT

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Abstract:

Regional mapping at a scale of 1:50,000, in parts of the Fort Liard and La Biche River (NTS 95B, 95C) map areas, was resumed during 2000 by the Geological Survey of Canada and the Yukon Geology Program as part of the Central Foreland NATMAP program. The Fort Liard and La Biche River map areas lie at the southern termination of the Mackenzie Mountains at the territorial border with British Columbia, and are part of the Foreland fold and thrust belt. Topography in the area is structurally controlled, with anticlines underlying the ranges and synclines underlying the valleys (Douglas, 1976; Douglas and Norris, 1976). The eastern half of the La Biche River area and the west half of the Fort Liard area are known to hold important gas reserves in structural plays, as exemplified by the Beaver River, Kotaneelee, Pointed Mountain, and La Biche fields. Mapping by the GSC focused on the Mount Martin (95C/1), Fisherman Lake (95B/5), and Mount Flett (95B/12) map areas. This continues work begun in 1997 with new mapping in Babiche Mountain (95C/8) and Chinkeh Creek (95C/9) (Currie *et al.*, 1999a,b).

In the outer foothills near Fort Liard, the exposed strata range from the Late Devonian to Carboniferous Besa River Formation to the Late Cretaceous Wapiti Formation. Much of the area is underlain by Carboniferous and Permian strata of the Prophet, Flett, Mattson and Fantasque formations (Currie *et al.*, 1998). With the exception of local exposures of Triassic strata, the upper Paleozoic succession is unconformably overlain by Lower Cretaceous rocks of the Chinkeh Formation and younger units (Fallas and Lane, 2001). Facies changes and sub-Cretaceous erosion are major controls on the distribution of these units within the project area. Although the regional distribution of map units on existing reconnaissance maps is broadly appropriate, our new mapping already has resulted in substantial revisions and updates.

The structural style in the eastern foothills comprises a series of en-echelon folds with a box fold geometry controlled largely by the competence of the Mattson Formation. The sinuous appearance in map trace of major structures is largely due to en echelon linkages of smaller culminations (Currie *et al.*, 1999a; Fallas, 2001). Nearly vertical bedding dips on the flanks of several structures may have led previous mappers to interpret thrust faults in some of these areas. In at least some cases, these interpreted faults are not required by the distribution of the map units. Where mapped, the thrust faults appear to sole into a décollement horizon in the Devonian to Carboniferous Besa River Formation.

Newly released total-field aeromagnetic anomaly data for the Fort Liard - La Biche region are dominated by large elliptical anomalies commonly greater than 20 km in diameter. These probably reflect "basement" features at depth. A prominent northeast trending anomaly coincides with the Liard Transfer Fault, an inferred basement feature that has been interpreted to control to the dramatic eastward deflection of the Cordilleran deformation front. Only in the western part of the project area do the anomalies reflect surface and near-surface features. Small elliptical magnetic highs correspond to mapped (and unmapped) Cretaceous and/or Tertiary intrusions and plugs. Locally, steep linear gradients approximate large mapped thrust faults.

A map of the second vertical derivative of the total-field aeromagnetic anomaly data reveals features in the shallow crust superimposed upon the major crustal features seen in the total-field data. The shallow features are most impressive in the central part of the project area where the Liard, Kotaneelee, and La Biche anticlines are clearly outlined. Farther east the Bovie Fault and its northern extension are well imaged. These data show, in part, the relationship between shallow structures and deeper crustal features, and support the hypothesis that deeper crustal features influence surface structures in the area.

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REGIONAL ELECTRICAL CONDUCTIVITY OF THE NORTHERN CORDILLERA: CORRIDORS 2 and 3

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The Slave-Northern Cordillera Lithospheric Evolution (SNORCLE) transect addresses LITHOPROBES's theme of crustal evolution through an integrated program of geological and geophysical studies across this geologically diverse region and through four billion years of Earth's history. The application of the magnetotelluric (MT) method to the study of the northern Cordillera has three principal objectives. The first concerns characterization of the regional conductivity properties of the northern Cordillera in relation to the major tectonic units. The second thrust involves upper features: determination of the depth extent and internal variation within the lithosphere beneath the Cordillera the fate of the vast amount of Pacific lithosphere overridden by North America during the past. The third specific goal is determination of the subsurface geometry and character of the Tintina strike-slip fault zone for comparison with the reflection data and a better understanding of reasons for conductivity enhancement at faults.

In this paper we present a study of the regional crustal structure along Corridors 2 and 3 of the SNORCLE-Northern Cordillera. The major discontinuities are imaged qualitatively through the use of magnetotelluric data. In addition, a preliminary interpretation of the data at terrane scale will be presented.

ENIGMATIC STRATA OF POSSIBLE NEOPROTEROZOIC AGE, POOL CREEK MAP AREA (NTS 95C/5), YUKON TERRITORY

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During the 1950s and 1970s, regional mapping by the Geological Survey of Canada in the La Biche map area (NTS 95C) of the southeastern Yukon Territory delineated a belt of unfossiliferous meta-sedimentary rocks in the western part of the map area (Douglas and Norris, 1959; Douglas, 1976). This unit, dominated by green argillite, was inferred to be of Proterozoic age and to lie unconformably beneath Palaeozoic strata. As part of the Central Foreland NATMAP Project, these rocks were recently re-examined during 1:50 000 scale mapping carried out in the Pool Creek map area (NTS 95C/5) by the Yukon Geology Program. This work, in combination with stratigraphic studies by the Geological Survey of Canada, indicates that this succession is more stratigraphically complex and potentially of greater stratigraphic significance than was previously realised (Allen and Pigage, 2000; Allen *et al.*, 2001). Study of these strata is complicated by poor exposure and by metamorphism related to emplacement of the Pool Creek Syenite (early Eocene). In addition to producing assemblages of metamorphic garnet, epidote, tremolite, and chlorite, contact metamorphism caused pervasive hornfelsing that welded most bedding-plane contacts.

The putatively Proterozoic succession has been subdivided into three informal, map-scale units by Allen and Pigage (2000; see also Allen *et al.*, 2001). Map Unit 1 forms the basal part of the succession, is at least 550 m thick, and consists of very fine-grained, massive to faintly laminated, white to light-grey quartz arenite. This is gradationally overlain by Map Unit 2, which consists of approximately 60 m of black, platy to fissile, silty shale. Beds of quartz arenite like those in Map Unit 1 occur interbedded with the shale in the basal part of Map Unit 2. Because these units are generally poorly exposed, the upper contact of Map Unit 2 and basal contact of Map Unit 1 have not been observed.

Map Unit 3 comprises the upper part of the succession and is up to 500 m thick. It is sufficiently well exposed to permit sedimentological analysis and can be subdivided into three facies associations. Facies Association 1 is dominated by green argillite and siltstone, with minor, very thin beds of sandstone. By contrast, Facies Association 2 is significantly sandier, consisting of interbedded green siltstone, argillite, and thin-bedded sandstone. The sandstone beds are massive to normally graded with rare parallel lamination and current ripples. Facies Association 3 consists of grey-green to green, poorly sorted to massive breccia. Breccia composition varies significantly from bed to

bed, with some being dominated by intraclasts of siltstone and argillite and others dominated by basalt clasts. Clasts of limestone are relatively common in breccias in the upper part of the unit. In addition to being unsorted, the breccia beds locally display upper-surface relief, suggesting deposition from debris flows. A single, red-weathering breccia unit occurs roughly 300 m above the base of Map Unit 3 and is up to 20 m thick. The significance of its distinctive colour is not known at present. Slump folding is common throughout the map unit, indicating deposition on a slope, while the sandstone beds most closely resemble distal tempestites. Deposition thus may have taken place at or near the slope-shelf break.

Douglas and Norris (1959) tentatively correlated these strata with the Proterozoic Apekkuni Formation of southern Alberta on the basis of lithologic similarity. However, the succession is broadly similar in general aspect to a number of Neoproterozoic and Cambrian formations that are more typical of northwestern Canada. A goal of fieldwork in 2001 will be to delineate the relationship of the succession in the Pool Creek map area with units mapped immediately to the west in the Coal River map area (NTS 95D; Gabrielse and Blusson, 1968). In the Coal River area, the basal Palaeozoic succession is constrained by trilobites and archaeocyathans (Handfield, 1971) and published descriptions of map units suggest some similarities with the more enigmatic succession in the Pool Creek Area (Gabrielse and Blusson, 1968). In addition, several occurrences of basic volcanic rocks have been delineated in the eastern part of the Coal River map area and represent potential sources for the volcanic clasts that dominate many breccia beds in Map Unit 3.

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Contrasting Arc and Non-Arc Mafic Volcanic Rocks in the Fyre Lake Unit, Yukon-Tanana Terrane, Finlayson Lake Region, Yukon, Canada.

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The Fyre Lake unit (FLU) of the Finlayson Lake volcanic-hosted massive sulphide (VHMS) district consists of a heterogeneous assemblage of variably deformed and metamorphosed, Devonian to early Mississippian subaqueous, and lesser subaerial, mafic and felsic volcanic and volcanoclastic rocks; mafic, ultramafic and felsic intrusive rocks; and lesser clastic sedimentary rocks (Fig. 1; Murphy, 1998; Murphy and Piercey, 1999, 2000; Piercey and Murphy, 2000). Mafic volcanic rocks from Fyre Lake unit have geochemical signatures indicating derivation from extremely heterogeneous mantle sources. These mafic volcanic rocks can be broken into two broad subdivisions including those from arc mantle sources characterized by negative Nb anomalies relative to Th and La on primitive mantle normalized plots (Fig. 2a-d) and those from non-arc mantle sources characterized by weakly negative to positive Nb anomalies relative to Th and La on primitive mantle normalized plots (Fig. 2e-g). Furthermore, within these subdivisions the mafic volcanic rocks show evidence for variable contributions from depleted versus enriched mantle sources (Fig. 3). Within the arc subdivision suites ranging from most depleted to enriched include: boninite (BON)- low Ti tholeiite island arc tholeiite (LOTI)-LREE-enriched island arc tholeiite (IAT)-calc-alkaline basalt (CAB) (Fig. 2a-d). In the non-arc subdivision suites ranging from most depleted to enriched include: back-arc basin basalt (BABB)- transitional mid-ocean ridge basalt (T-MORB)- Nb-enriched ocean island-like basalt (NEB) (Fig. 2e-d). We suggest that such a plethora of volcanic suites and the association on arc and non-arc volcanics represents a complex evolving Devonian-Mississippian arc-back-arc system that may be due to: 1) arc magmatism punctuated by back-arc initiation and subsequent back arc evolution due to non-arc asthenospheric underplating of arc lithosphere (Fig. 4; e.g., Cameron, 1985; Crawford et al., 1989); or 2) arc magmatism punctuated by back-arc initiation due to intersection of a spreading ridge with an evolving arc and the interaction between non-arc asthenosphere and arc lithosphere (Fig. 4; e.g., Falloon and Crawford, 1991; Monzier et al., 1993; Meffre et al., 1996). Other models may also be appropriate (e.g., arc initiation, slab window formation; Crawford et al., 1989; Pearce et al., 1992) but combined with the above require further testing.

The Fyre Lake VHMS deposit is hosted by primitive arc (boninitic) volcanics, lies within an interpreted synvolcanic basin (Murphy and Piercey, 2000), and appears to be associated with a transition from arc to non-arc volcanism (Piercey et al., 1999; Seibert and Hunt, 1999). These features and in particular shifts in magmatism from arc to non-arc are prospective locations to find new VHMS mineralization and are common in mafic associated VHMS districts (e.g., Flin Flon-Snow Lake; Cyprus; Newfoundland ophiolite-hosted systems; Bailes and Galley, 1999; Swinden, 1991).

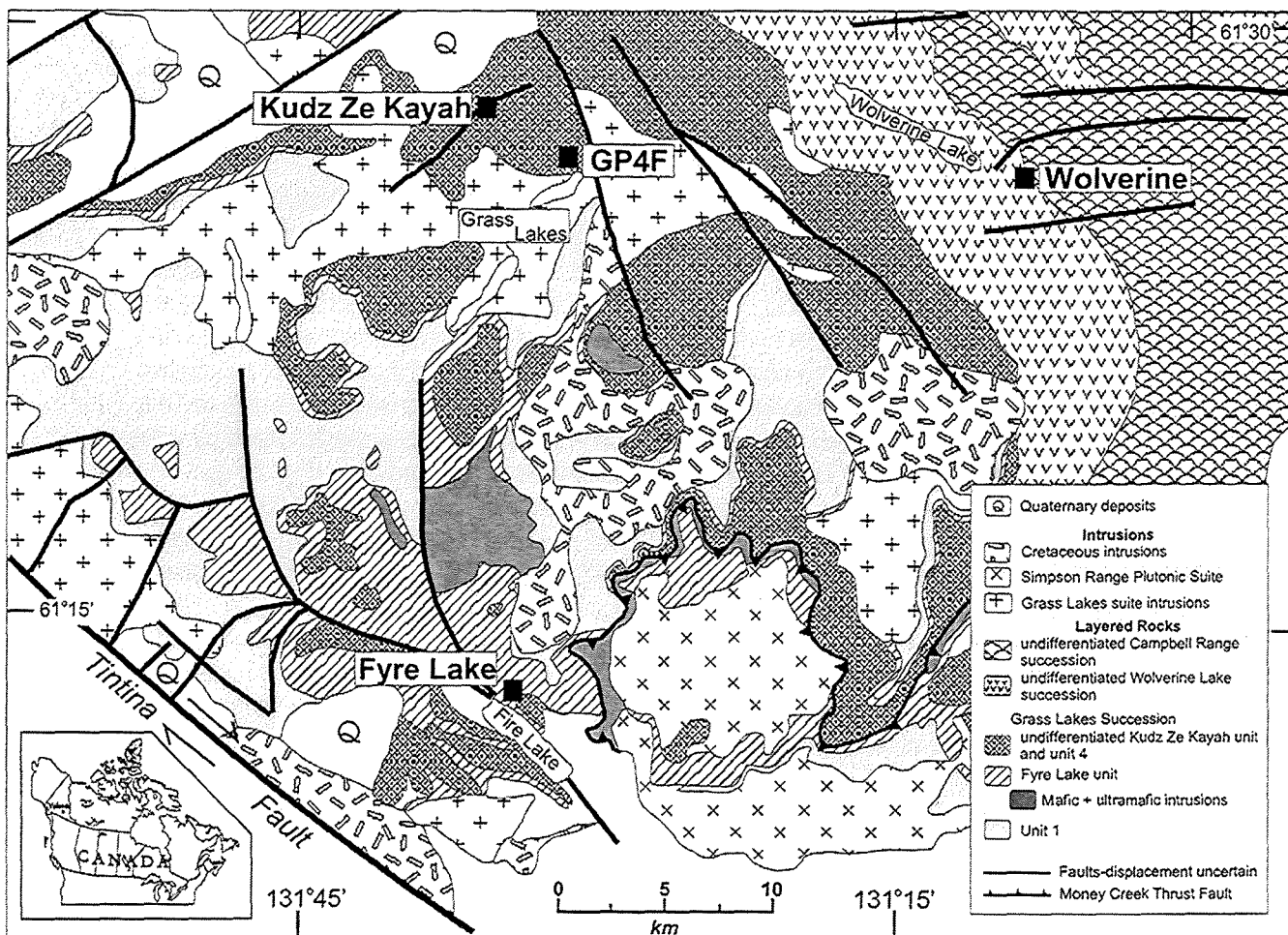


Figure 1. Geological setting of the Fyre Lake unit in relationship to other mid-Paleozoic rocks of the Yukon-Tanana Terrane in the Finlayson Lake region, Yukon. Map has been simplified from Murphy and Piercey (2000). Shown are volcanic-hosted massive sulphide (VHMS) deposits of the Finlayson Lake district.

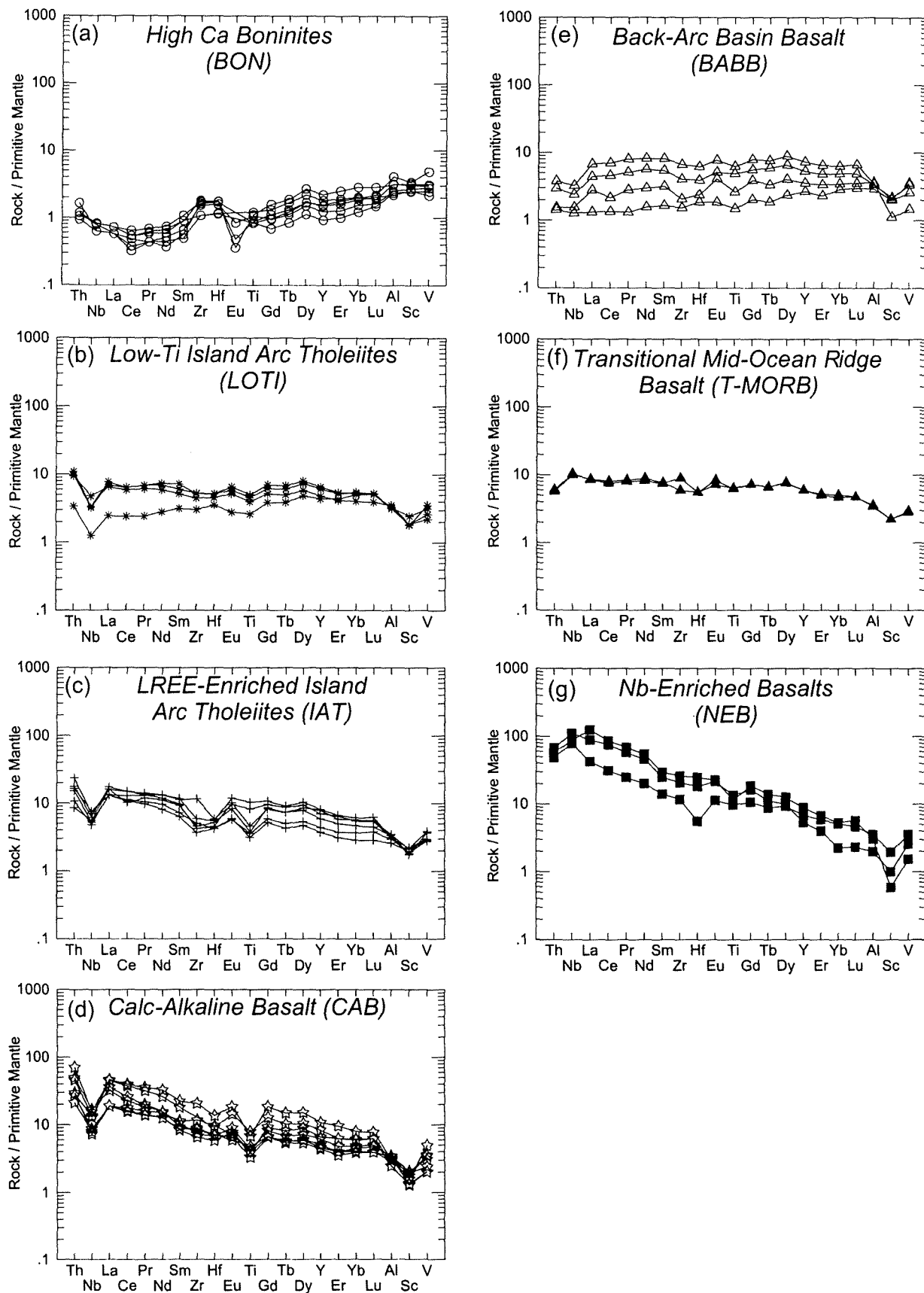


Fig. 2. In a-d it is notable that all samples exhibit a strong negative Nb anomaly relative Th and La and depletions in high field strength elements (HFSE; e.g., Ti, Hf, Zr, Nb, Ta). **These negative Nb anomalies and HFSE depletions are diagnostic features of arc volcanic rocks.** From a-d there is also a progression from extremely depleted mantle sources (i.e., boninites) to progressively more enriched mantle sources (e.g., calc-alkaline basalts) as indicated by the progressive increase in LREE contents (La, Ce) to HREE contents (Yb, Lu) across these suites. In e-g it is notable that all of the samples exhibit weak (e.g., BABB) to no negative Nb anomalies and low HFSE (e.g., BABB) to very enriched HFSE (NEB). **The lack of pronounced negative Nb anomalies suggest these rocks were derived from non-arc mantle sources, or those weakly affected by subducted slab metasomatic flux.** As in the arc rocks the rocks from e-g reflect a progression from derivation from depleted through enriched mantle sources.

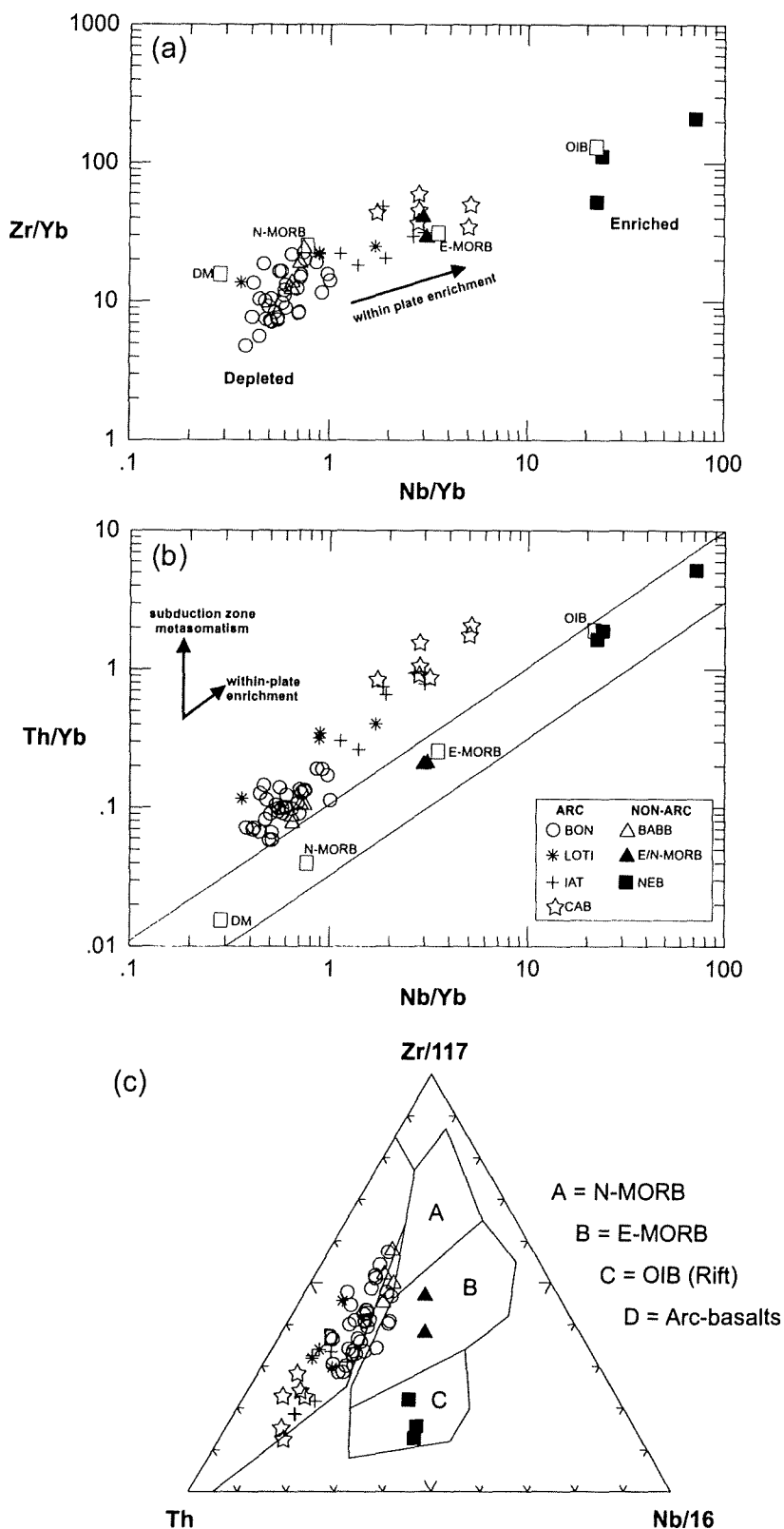
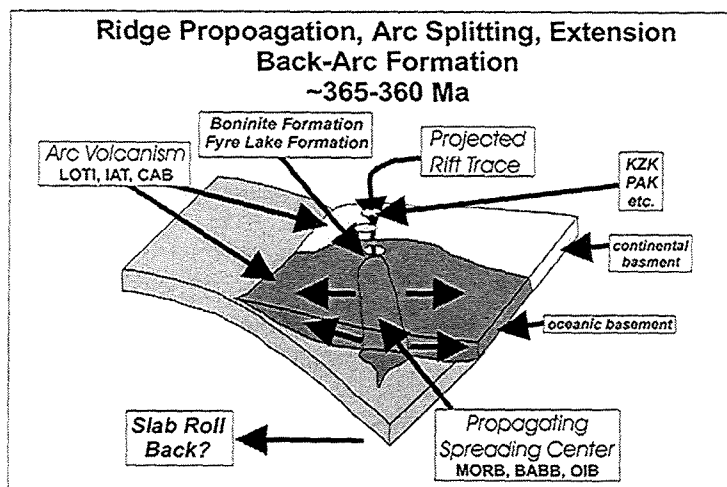
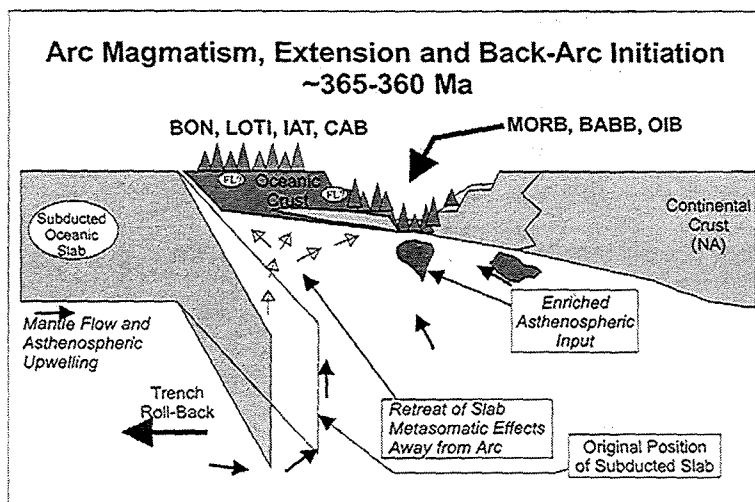


Figure 3. In (a) the plot of Zr/Yb - Nb/Yb reflects the similar incompatibilities of Zr and Nb in mantle melts. The increasing Zr and Nb reflects increasing contributions from enriched mantle source regions illustrating that both the arc and non-arc rocks from the Fyre Lake unit exhibit various contributions from enriched versus depleted mantle sources. Shown on this diagram to illustrate the progression from depleted to enriched sources are the depleted mantle (DM; Staudigel et al., 1998), N-MORB, E-MORB and OIB (Sun and McDonough, 1989) reservoirs. The Th/Yb - Nb/Yb is somewhat similar as both Th and Nb have similar incompatibility in mantle melts. Notably, however, Th is often metasomatically added to arc mantle source regions leading to negative Nb anomalies characteristic of arc magmas. In this diagram rocks with very little subducted slab influence (non-arc) lie within the mantle array as defined by the DM-N-MORB-E-MORB-OIB array. Samples of arc parentage or those influenced by subducted slab flux (BABB) lie on a different trajectory with higher Th at a given Nb content than non-arc rocks. Diagrams modified after Pearce and Peate (1995). In the Th - Zr - Nb diagram (Wood, 1980) the non-arc samples lie primarily in the field for OIB and E-MORB with BABB drifting towards the arc fields reflecting the weak negative Nb anomalies of these rocks. Notably, for a given Zr and Nb content the arc rocks have higher Th contents consistent with derivation from arc mantle sources.



Model # 1. Ridge Propagation Model

In this model arc volcanism is occurring within the Fyre Lake unit and then a spreading ridge intersects the arc and begins to split the arc. During the arc volcanic stage there was the eruption of LOTI, IAT, and CAB. With ridge propagation into the arc there would have been upwelling of hot non-arc asthenosphere that could drive melting of the refractory mantle source residues left after LOTI, IAT, and CAB extraction resulting in the formation of the boninites. Continued extension and melting of the asthenosphere would result in the non-arc volcanic rocks and interaction and contamination by sub-arc lithosphere could yield rocks with hybrid arc and non-arc characteristics (e.g., BABB). Notably, the Fyre Lake deposit is preferentially associated with boninitic rocks (Sebert and Hunt, 1999) and the Fyre Lake deposit likely formed during the arc breakup and ridge intersection. With the exception of lack of continental crust this tectonic situation is very similar to modern day Tonga Forearc and the Hunter Fracture Zone, North Fiji Basin.



Model # 2. Arc-Rifting and Back-Arc Initiation Model

In this model arc volcanism is occurring within the Fyre Lake unit and then trench roll back inducing back-arc spreading. In this case during the arc stages there would be the eruption of LOTI, IAT and CAB. Rifting of the arc and back-arc initiation would result in asthenospheric underplating of the arc with non-arc asthenospheric mantle that could have driven partial melting of the refractory mantle wedge, left after LOTI, IAT and CAB extraction, to form boninites. Similar to the previous model continued extension and melting of the asthenosphere would result in the non-arc volcanic rocks and interaction and contamination by sub-arc lithosphere could yield rocks with hybrid arc and non-arc characteristics (e.g., BABB). In this model the Fyre Lake deposit would form just as back-arc initiation begins and the initial stages of arc rifting on the arc to back-arc magmatic transition; similar scenarios have been observed in many other world class VHMS camps such as Flin Flon - Snow Lake, and ophiolite-hosted deposits in Cyprus.

Figure 4. Potential petro-tectonic models to explain the diversity of arc and non-arc volcanic rocks within the Fyre Lake unit of the Yukon-Tanana Terrane in the Finlayson Lake region.

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ANATOMY OF AN ANCIENT PASSIVE MARGIN TURBIDITE SYSTEM: THE WINDERMERE SUPERGROUP (WESTERN CANADA)

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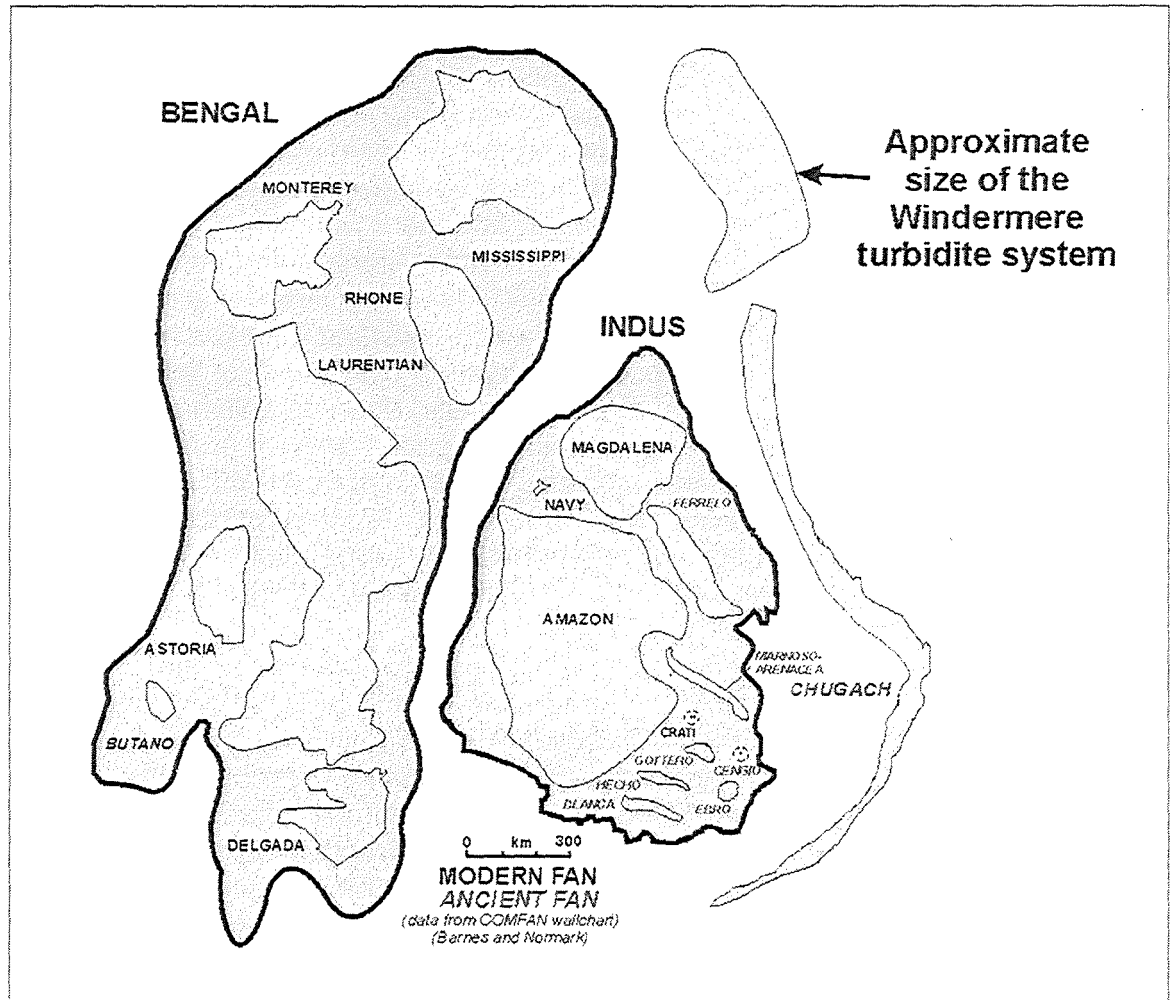
ABSTRACT

The Windermere Supergroup is a regionally widespread succession of predominantly deep marine sedimentary rocks deposited on the ancestral continental margin of western Canada following its separation from Australia ca. 700 Ma. A vertical succession from sand-rich basin plain turbidites to mud-rich slope facies occurs over a ca. 7 km thick stratigraphic interval and records the progradation of the continental margin attendant with diminished rates of thermally-driven subsidence. Although unfossiliferous and deformed in the Mesozoic, careful stratigraphic mapping, utilizing a lithologically distinct condensed interval and two carbonate-rich lowstand intervals have established regional continuity of this depositional system over ca. 4° of latitude. The continuity of the depositional system has also been established using geochronology of detrital zircons which reveals a uniform provenance throughout the southeastern Canadian Cordillera. Taken in concert with facies patterns and paleocurrent data, the Windermere turbidites are interpreted as a longitudinal turbidite complex that flowed west-northwest, subparallel to the base of the continental margin and was fed from a distinct source area in the present southeastern Cordillera.

Basin plain facies are dominated by stacked, acyclic lobate units of thick-bedded, coarse sand to pebble grade arkose and minor intervening "classic" turbidites. In contrast, the slope facies are dominated by thin-bedded "classic" turbidites that enclose channeled coarse sand units and two deep-water carbonate intervals, the latter resedimented from the adjacent shelf during falling sea level. The contrast between slope and basin plain facies is attributed to sediment bypass such that the basin plain was the repository for much of the sand-grade material. Local evidence of syndepositional structural control is found in the slope facies and is provided by thick, sand-rich stratigraphic intervals deposited in northwest-trending structural troughs adjacent to coeval mud-rich intervals deposited on structural highs.

As pointed out by numerous workers, most ancient turbidite systems are a fraction of the size of modern counterparts. In addition, most on-land examples of turbidite systems were deposited in small rift basins (North Sea) or convergent margin basins. Although the Windermere turbidite system has been deformed, the establishment of stratigraphic and provenance continuity, in conjunction with published estimates of palinspastic restorations, suggests that this system was deposited over an area of at least 160,000 km², comparable in size to modern fans such as the Amazon. Furthermore, its tectonic setting, along the paleoPacific passive margin of western Canada, is analogous to modern offshore examples. This makes the Windermere a candidate for one of the largest intact passive margin turbidite sequences preserved in the rock record and may contain important lessons for understanding modern deep water targets

along continental margins. From the perspective of the Neoproterozoic Earth, the presence of a thick sequence of well-preserved deep water sedimentary rocks offers the opportunity to learn more about the nature of Neoproterozoic sea water and geochemical sequestration.



Comparison of the size of the reconstructed Windermere turbidite system with selected modern and ancient examples. Note that the Windermere is comparable to modern passive margin fans such as the Amazon and Laurentian and dwarfs ancient examples that have been used to develop most facies models and exploration concepts.

U-PB GEOCHRONOLOGY OF SUBPHANEROZOIC BASEMENT OF THE SOUTHWEST NORTHWEST TERRITORIES: A PRELIMINARY REPORT

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Basement domains in the southwest NWT are assigned to Wopmay Orogen, based largely on the extrapolation of the magnetic signatures of the exposed Wopmay Orogen more than 400 km south into the covered basement west of Great Slave Lake. In order to ground truth this correlation and provide age constraints for the recently recorded SNORCLE seismic reflection line, we examined the composition and age of drill core and cuttings in this region from over 100 wells known to penetrate basement. From this set 10 samples of core and cuttings were selected for U-Pb geochronology using the SHRIMP II ion microprobe at GSC Ottawa. Aeromagnetic data were used as the base for geophysical subdivision and samples were collected from each of the major domains which, comprise from east to west, the informal Hay River terrane, Great Bear Magmatic Zone, Hottah terrane, Fort Simpson terrane and Nahanni terrane.

The Hay River terrane lies along the southwest shore of Great Slave Lake and to our surprise gave an age of 1838 Ma, making it the youngest rock in the region. Wells in Great Bear Magmatic Zone returned cuttings of rhyolite and intrusive rock that gave U-Pb ages of 1873-1862 Ma - nearly identical to dates from the older portion (pre-folding plutons) of the exposed magmatic rocks near Great Bear Lake. In contrast to the exposed shield, Hottah terrane (>1914 Ma) has a different range of distinctly younger ages in the subsurface, ranging from 1875-1854 Ma, very similar to ages considered typical of GBMZ. Collectively the Hottah terrane is cut by a suite of northwest-trending positive linear aeromagnetic anomalies that resemble a dyke swarm. The magnetic linears are characteristic of the Hottah terrane throughout the southern NWT and northeastern British Columbia. Fortunately, several drill core and cuttings of fine to medium grained diabase confirm the dyke interpretation. The Fort Simpson terrane comprises plutonic rocks dated at 1845 -1853 Ma. The Nahanni terrane coincides closely with the zero-edge of a westward thickening wedge of Precambrian sedimentary rocks and thus, from the petroleum perspective, represents a fundamentally different kind of subPhanerozoic basement.

The geochronology offers some new ideas and poses some new questions on the interpretation of the SNORCLE reflection line and tectonic models for southern Wopmay orogen. It confirms that non-reflective sequences seen above reflective "Hottah terrane" on the SNORCLE line are magmatic arc rocks. This sequence is referred to informally as the 'P-56 basin' and represents a thick (15 km), seismically transparent region along the boundary zone of the Hottah and Fort Simpson terranes. The geochronology indicates that that a broad region (300 km across strike) of southern Wopmay orogen was affected by magmatism more or less synchronously (1872-1845 Ma). This suggests either a single subduction system or a series of subduction zones. The latter option contrasts with the SNORCLE reflection profile that shows a single, east-dipping subduction zone. This may

be reconciled if the P-56 basin and associated magmatism in southern Hottah Terrane is interpreted as the result of ridge subduction (suggested by Paul Hoffman in 1989) and thus represents anomalous near-trench magmatism, similar to the Tertiary plutonic rocks seen in the Chugach terrane of Alaska. Such a scenario allows for a single east-dipping subduction zone to generate the Great Bear Magmatic Zone coeval with ridge subduction to generate near-trench, but coeval, magmatism of the Hottah Terrane and P-56 basin.

PROVENANCE OF THE BELT BASIN: A PALEOGEOGRAPHIC SNAPSHOT OF PRE-RODINIA WESTERN NORTH AMERICA

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The Mesoproterozoic (1.47-1.37 Ga) Belt Basin in northwest United States and southern Canada is a thick (~16 km) assemblage of predominantly fine-grained siliciclastic strata with a much-debated tectonic setting. We have dated over 400 detrital zircon grains (U-Pb SHRIMP) from formations that cover the entire stratigraphic thickness of the Belt as well as from geographically diverse sections along the east, southern and western sides of the preserved basin. The new data offer an enhanced view of the mysterious western craton, the source area for a substantial portion of the Belt that has been removed by post-Belt tectonic processes.

The results indicate that three general sources have contributed detrital zircons to the Belt Basin. An eastern source (proximal Canadian Shield) supplied clastic material found in the sandstones in the Grinnell, Siyeh, Creston formations. These samples are characterized by a suite of detrital zircon ages in the intervals of 1.78-1.90 Ga and > 2.6 Ga, with subordinate grains in the interval 2.2-2.4 Ga. In Montana, a southern source for the Lahood, Newland and part of the Aldridge-Prichard formations is compatible with derivation from the Wyoming Province (2.5-2.7 Ga and 3.2-3.4 Ga), but these samples also contain a substantial proportion of young Proterozoic grains (1.76 Ga –1.9 Ga), derived possibly from the Great Falls Tectonic Zone. A western source appears in the upper Prichard and Aldridge formation quartzite facies and persists into Missoula Group deposition (over 7 km of stratigraphy). The western source is inferred from detrital zircons that range in age from 1510-1610 Ma, distinctly non-North American ages. Significantly, these samples also contain zircons with ages that nearly overlap with those for intraBelt magmatic rocks (1467-1435 Ma), suggesting that some Belt sedimentation may have been contemporaneous with magmatism occurring in the western craton.

The new data confirm that the western craton was similar to present-day eastern Australia in terms of the range of ages, supporting a pre-Rodinia (pre 1.0 Ga) juxtaposition of eastern Australia and western North America. Based on rapid rates of sedimentation implied by geochronology of intraBelt magmatic rocks (15 km in circa 70 m.y.) , the apparent presence of syndepositional magmatism and persistent reactivation in the western craton, we envision that an oblique extensional setting with a magmatically and tectonically active source area, broadly speaking a back-arc basin, may be a plausible tectonic setting for the Belt basin. Cenozoic Tethyan basins such as the Black Sea and Caspian Sea may be suitable analogues. If so it would imply that the western margin of North America circa 1450 Ma, although locally extensional, was predominantly a convergent margin.

The Cunningham Formation: A Ramp-to-Platform Carbonate in the Terminal Proterozoic of Western Canada

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The Neoproterozoic was a time of immense change in global biogeochemical cycles, climate and tectonics. The geochemistry of marine sedimentary carbonates provide one of the best means of recognizing and interpreting variations in seawater chemistry. It is within this backdrop of extreme global change that the Cunningham Formation of the Cariboo Mountains (British Columbia) was deposited. The Cunningham is a thick section (~800m) of regionally extensive carbonate that provides important but unstudied potential to better understand the Neoproterozoic. In particular it appears to be one of the youngest well-preserved carbonate units of the Windermere Supergroup and thus overlaps globally with a time of diversification of metazoan life forms.

The Cunningham Formation is found in the northern and western Cariboo Mountains where it is well-exposed in cliff and cirque sections above treeline. It overlies the Isaac Formation, a mixed siliciclastic and limestone unit deposited in a slope setting. It is overlain by storm-dominated shallow marine mixed clastic carbonate rocks of the Yankee Belle Formation. The Cunningham has been split into three main facies belts along a dip-section that extends for nearly 40 km from northeast to southwest; the outer platform in the west, the middle platform and the inner platform in the east (Ferguson 1994). The outer platform is by far the thickest, composed of cliffs of massive-weathering well-bedded limestone. The lower part of the outer platform comprises current-laminated, thin-bedded limestones that change upwards to a thick-bedded grainstone-packstone dominated facies containing common dolomicritic rip-up clasts. The grainstone-packstone facies comprises massive to commonly cross-bedded oolitic and intraclastic calcarenites to rudites that record deposition in a very high-energy environment. This zone is interpreted as the outer edge of the platform, a high-energy shoal environment, comparable to the windward margin of the modern Bahamian platform. The middle platform is composed of varied facies that include laminated limestone and ooid-pisoid packstone as well as interbedded siliciclastic mudstone and laminated quartz sandstone. This portion of the Cunningham is distinguished by the presence of siliciclastic material and by its cyclic sedimentation patterns interpreted to be shoaling upward cycles. The siliciclastic facies are interpreted a part of a coastal lagoon that developed by behind (landward) of the ooid shoal complex of the outer platform. The inner platform also contains cyclic sediments but the belts in this zone are thinner and extremely dolomitized in comparison to the central platform. The facies in this zone consist of massive and quartzose carbonate. This is the only area in the Cunningham to contain stromatolites, which are found in thick dolomitized beds. The inner platform is interpreted as "near shore" due to common siliciclastic units interbedded in the carbonates, the presence of stromatolites and because of the extensive dolomitization.

The restriction of the stromatolites to relatively quiet water lagoonal and nearshore facies suggest that substrate mobility of the outer platform inhibited stromatolite colonization.

The Cunningham Formation has yet to be studied in detail with respect to facies and depositional history. In addition, no geochemical or isotopic data are available for this unit. The Cunningham thus offers an excellent opportunity to reconstruct an extensive dip-profile of a Neoproterozoic carbonate unit. The presence of both limestone and dolostone in off- and nearshore facies will provide an excellent opportunity to assess the effects of dolomitization on isotope geochemistry. Eventually, isotopic profiles through vertical sections in the Cunningham will offer the possibility of global correlation as well as understanding seawater chemistry in western Canada at this time interval.

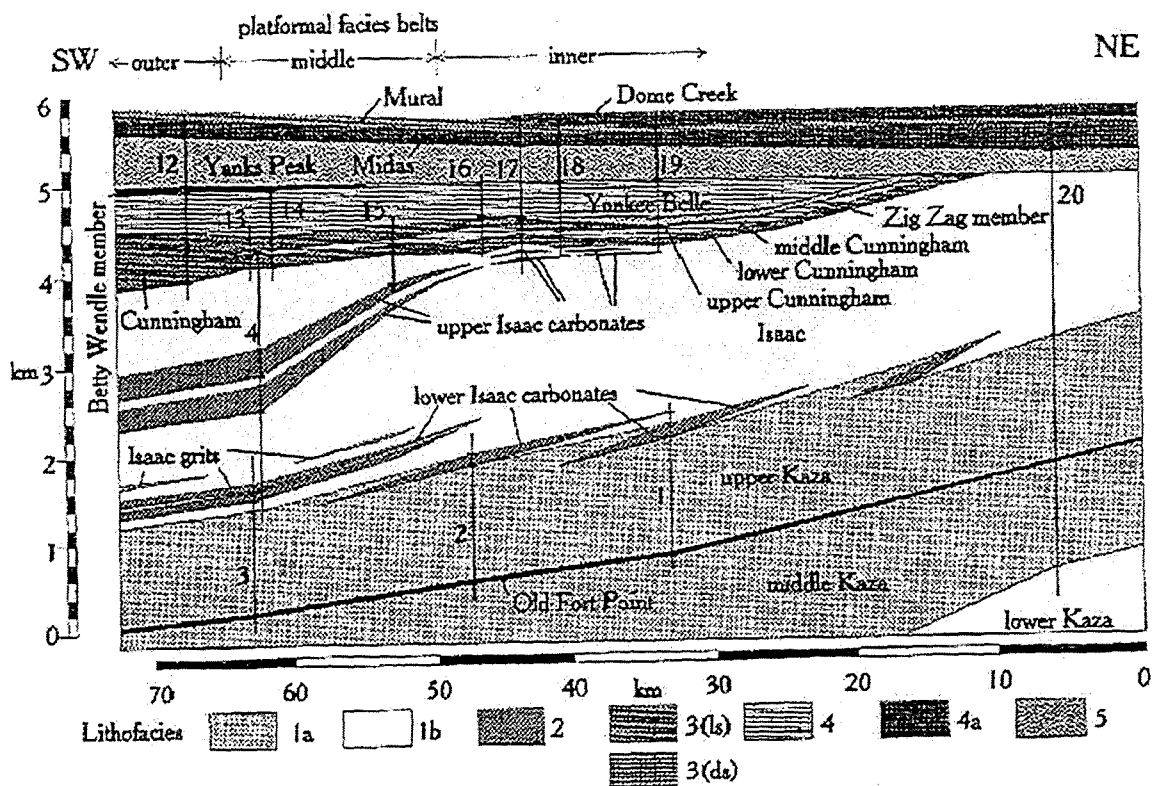


Figure 1. A stratigraphic transect of the northern Cariboo Mountains and westernmost Rocky Mountains. NOTE: The three platformal facies belts in the Cunningham Formation. Lithofacies and sublithofacies are: 1a) sandstone-dominated turbidite, 1b) pelite-dominated turbidite, 2) deep water carbonate, 3) shallow-water carbonate which is further divided into limestone (ls) and dolostone (ds), 4) shallow-water siliciclastic and carbonate, 4a) shallow-water siliciclastic (mostly mudstone) and 5) quartzite. Vertical exaggeration is 6.4X. (Figure from Ferguson 1994)

Protolith and tectonic significance of amphibolite facies gneisses of the Yukon-Tanana terrane in the Thistle Creek area (Stewart River map sheet), Yukon Territory

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Introduction

Geological mapping during the summer of 2000 near Thistle Creek represents the first of a four year program in the Stewart River area under the auspices of the Ancient Pacific Margin NATMAP project. The NATMAP project seeks to understand the composition, relationships, and metallogeny of poorly understood terranes caught between the ancestral North American margin and those known with more certainty to be tectonically accreted (Thompson et al. 2000). The Stewart River component focuses on the stratigraphic, structural and tectonic history, and the economic framework of the Yukon-Tanana terrane, which comprises complexly deformed meta-igneous and meta-sedimentary rocks of mostly Paleozoic(?) age. A multiparameter geophysical survey headed by Rob Shives (GSC Ottawa) was flown in a northeast-trending swath across the Stewart River area under the auspices of the Targeted Geoscience Initiative, and will be integrated with other aspects of the project. Preliminary results of the gamma ray and aeromagnetic surveys indicate that many geological contacts can be delineated with clarity. These surveys will not only be of great use for improving bedrock mapping, because the area has very poor quality bedrock exposure, but also for finding potential targets for mineral exploration.

Local Geology

Yukon-Tanana rocks in the Thistle Creek area comprise polydeformed and metamorphosed amphibolite facies gneisses and schists intruded by younger plutonic rocks. These rocks are discussed briefly herein, and are described more fully in Ryan and Gordey (2001a, 2001b). The gneisses 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 veins. Regional correlations of plutonic suites indicate that the transposition event was post-Carboniferous, and pre-Jurassic (eg. Mortensen 1992). Although stratigraphy is obscured by the transposition, as well as by later, open folds and faults, the area is clearly dominated by two fault-bounded tectonostratigraphic assemblages: 1) polyphase grey orthogneiss in garnet-amphibolite schist/gneiss, interpreted herein as a metavolcano-plutonic complex; 2) interstratified garnet-amphibolite schist/gneiss and metasedimentary schist and paragneiss derived from psammite, semipelite and quartz-arenite, collectively interpreted as a metavolcano-sedimentary succession. Geochronological investigations are underway (M. Villeneuve, GSC Ottawa) to place constraints on stratigraphic younging, provenance, ages of plutonism and timing of metamorphism and deformation.

Amphibolites, which are the most widespread rock type, have been intensely tectonized. Heterogeneous compositional layering and local vestiges of primary textures such as breccia clasts or pillow selvages indicate derivation from mafic volcanic to volcanoclastic rocks. Quartz-sericite schist, likely derived from felsic volcanic rocks or hypabyssal intrusions, are locally interlayered with the amphibolites. The interfingered metafelsites and amphibolites are consistent with bimodal volcanism, possibly in an arc setting.

Grey and white banded quartzites are generally in fault contact with the gneisses and schists, and their stratigraphic relationships are equivocal. An occurrence of metaconglomerate

along the Yukon River exhibits quartz and tonalite clasts in a matrix of quartzofeldspathic schist. The conglomerate grades locally to impure quartz arenite, and is adjacent to an occurrence of grey quartzite. Detrital zircons are being analyzed to constrain the age of the conglomerate and its provenance.

Three bodies of strongly deformed potassic feldspar augen granite gneiss may correlate with a suite of Devonian-Mississippian augen granites dated regionally at ca. 360 Ma (Mortensen, 1992). Young, leucocratic granitic plutons and/or dykes cross-cut the gneisses; some carry a weak foliation in their margins, and others exhibit weak boudinage. A pluton and several smaller satellite bodies and dykes in the central part of the map area consist of syenogranite with large potassic feldspar phenocrysts. They are undeformed, and likely Cretaceous or younger. Rare, crosscutting, quartz-potassic feldspar porphyritic rhyolite to rhyodacite stocks are probably of Tertiary age.

Deformation and metamorphism

Transposition foliation

The regional foliation (S_T) in the area is characterized by high-strain transposition of layering in gneisses and schists, with abundant intrafolial isoclinal folds that are commonly rootless. Prior to regional F_3 folding, S_T had a shallowly dipping orientation. Primary compositional layering (S_0) in metasedimentary rocks, unit contacts (e.g., dyke margins), and a pre-existing foliation (S_1) can be traced around closures of generally recumbent to shallowly inclined, close to isoclinal, long-wavelength folds (F_2). Associated with the folds is an intensely developed regional extension lineation (L_2) that is parallel to the F_2 axes. Surprisingly, even at these high strains, there is little development of an axial planar fabric; rather, S_T is characterized by the complete reorientation of, and strain intensification of, S_0 and S_1 . Although S_T is largely defined by pre-existing layering, its present geometrical surface is considered a second generation feature. Large garnet porphyroblasts preserve an S_1 foliation as inclusion trails at a high angle to enveloping S_T fabric, indicating significant rotation of the porphyroblasts during F_2 deformation. Locally, amphibolites show contradictory fabric-porphyroblast relationships. An apparent older set of amphiboles helps define the L_2 lineation in some layers, whereas in other layers a second population, made up of rosettes, radiate within, and locally across, the foliation plane. It is unclear if the rocks have undergone polymetamorphism, or that fabric development was diachronous. Further petrography will help clarify the relative timing of porphyroblast growth and fabric development across the area.

The intensity of strain within the regional foliation locally grades to mylonite; the amount of displacement along the mylonitic bands, however, is unknown. Spacing between local mafic boudins in the gneisses indicates extension on the order of 1000%. Some F_2 folds are doubly closing, which may indicate a sheath-like geometry, further attesting to the high strain in these gneisses. Despite the high strain associated with the F_2 event, vergence (direction of rotation of the short limb) is systematic. In the north-trending units on the western side of the map area, the vergence of F_2 folds is generally to the west, and their axes and L_2 plunge shallowly north and south. In the east-trending units to the south and east, F_2 folds verge to the north, and plunge west-southwest or east-northeast parallel to L_2 . We interpret that the S_T and L_2 had a consistent orientation across the area prior to faulting and F_3 folding. Parallelism between the regional extension lineation and the transposition fold axes, at a striking obliquity to direction of rotation of the fold limbs, precludes the use of F_2 fold vergence as a means of determining transport direction in these rocks. This structural style is most consistent with having developed in a

triclinic noncoaxial flow regime, where the transport direction was probably parallel to the extension lineation. Our future mapping should help clarify whether the transposition foliation is part of a regional-scale shear zone, or merely characterizes the type of flow active in the middle crust during and after orogenesis.

F₃ folds

F₃ folds are open, moderately inclined (with shallow to steep varieties), shallowly plunging structures. They have weak axial-planar fabric where developed in schistose layers, and have no associated extension lineation. Some localities show a moderately inclined, to upright spaced S₃ cleavage (spaced 1-3 cm), overprinting F₂ isoclinal. Tighter examples of F₃ folds can be geometrically difficult to distinguish from more open F₂ folds. F₃ structures postdate regional metamorphism, as the S₃ cleavage everywhere wraps the peak metamorphic assemblage. F₃ deformation of the amphibolite facies rocks occurred at relatively low metamorphic grade, because kinked muscovite and biotite grains show little post-F₃ diffusion-assisted recovery.

Faults

The Thistle Creek area, as well as the greater Stewart River area, is transected by abundant prominent physiographic and aeromagnetic lineaments, some of which correspond to faults in the underlying bedrock. Locating faults is hampered by poor exposure, particularly in valleys. Faults are more readily apparent where there is striking lithological or structural changes.

An east-trending fault that runs through the upper part of Blueberry and Lulu creeks not only forms a prominent topographic and aeromagnetic lineament, but also clearly separates rocks of the volcano-plutonic complex from the volcano-sedimentary succession. A granite that is truncated by the fault shows brecciation, with small east-west-trending, vertically dipping cataclastic shear bands, and intense chlorite alteration, adjacent to the fault. The age of the granite is unknown, and thus the age of the fault is not constrained. The low-grade mineral assemblage associated with the structure indicates that it formed at a relatively shallow crustal level, possibly quite late in the tectonic history.

Another prominent topographic and aeromagnetic lineament trends northwestward from the southeast corner of the Stewart River map area, along the valleys of Walhalla and Scroggie creeks, and the Stewart, Yukon and Sixty Mile rivers, for a combined strike length of > 90 km. Future mapping will reveal if there is significant geological offset along this feature. It is noteworthy that this lineament lies immediately along strike from the presently mapped northwestern extent of the Teslin Fault.

Discussion

The high strain and amphibolite facies metamorphism associated with transposition deformation and gneissosity development in the Yukon-Tanana rocks at Thistle Creek are typical of mid-crustal orogenic deformation. The timing and regional extent of this orogenic episode is poorly understood, and partly obscured by Jurassic and younger plutonism, volcanism, sedimentation and faulting. The orogenic episode probably postdates the late Permian, because ca. 260 Ma rocks in the Klondike region (Mortensen 1992) show similar deformational character to the Thistle Creek area.

Despite the S_T transposition deformation and metamorphism, the stratigraphy at Thistle Creek is seen to comprise a volcano-plutonic, volcano-sedimentary succession, which is most consistent with an arc setting. Restoration of ca. 430 km of Cretaceous-Tertiary dextral offset

along Tintina Fault places the Stewart River area close to the Finlayson Lake belt, an area of significant VMS potential (Murphy 1998). It is possible that the Yukon-Tanana rocks at Thistle Creek have stratigraphic links with the Finlayson Lake district, but have been deformed and metamorphosed at a deeper crustal level.

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Mapping of Superior Province Mantle Using Garnet Geochemistry

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The Superior Province of the Canadian Shield is host to two well characterized kimberlite clusters, the Temiskaming and Kirkland Lake kimberlite fields. The Kirkland Lake field lies to the north-east of the town of Kirkland Lake, approximately 75 km from the Quebec-Ontario boarder. The Temiskaming field is 200 km south of Kirkland Lake, on the north-west shore of Lake Temiskaming at Cobalt, Ontario. The kimberlites intrude Precambrian granite-greenstone sequences of the western Abitibi Subprovince, in the Lake Temiskaming Structural Zone (L.T.S.Z.). Uranium-Pb dates on kimberlitic perovskites from both fields yield eruption ages of 155 Ma (Heaman and Kjasgaard, 2000). The four kimberlites contain abundant mantle xenoliths and xenocrysts, the chemistries of which have been used to reconstruct mantle stratigraphy and structure during the Jurassic.

Calculated pressures and temperatures of equilibration of mantle-derived garnet peridotite xenoliths recovered from the Kirkland Lake kimberlites reveal a 40 mWm⁻² conductive steady-state sub-cratonic palaeo-geothermal gradient (Vicker, 1997). Garnet xenocrysts from the C14 and AM47 kimberlite pipes (Kirkland Lake) and Bucke pipe (Temiskaming), as well as the Lake Ellen kimberlite (in the Superior Province in Michigan, U.S.A.), were analyzed for 25 trace elements using an LA-ICPMS. The majority of the garnets derive from low-Ca garnet lherzolite, with a small number of harzburgite affinity. Equilibration temperatures were estimated using the Ni-in-garnet geothermometer of Canil (1999a) then projected to the local geothermal gradient to approximate relative depth, assuming the garnets equilibrated along the xenolith palaeogeotherm. Pipes AM47 and C14 sampled material over a depth interval of 125 – 165 km (corresponding to equilibration temperatures of 940°C – 1140°C), whereas the

Bucke and Lake Ellen pipes contain garnets from 125 – 212 km (940°C – 1275°C) depth. Garnet trace element concentrations from the C14, AM47 and Lake Ellen pipes are similarly constant over the entire depth range. We suggest “gaps” in the depth data are representative of non-peridotitic lithologies. Results from the Bucke pipe have order of magnitude differences in trace element concentrations over the entire depth interval sampled, and do not show the same depleted layers as the other pipes. The correlation between increased trace element concentration and depth indicates a strong temperature effect on equilibrium trace element distribution.

The overall low TiO₂, Y and Zr concentrations of the garnets from all the pipes corresponds to the Archon setting of Griffin *et al.* (1999a) (Archon refers to a segment of the crust that has not been affected by a tectonothermal event for >2.5 Ga). The narrow range of trace element concentrations from AM47, C14 and Lake Ellen indicate the mantle below the Superior has a broadly consistent composition, with occasional non-peridotitic intervals. The scatter in the Bucke data may suggest that the mantle below the Temiskaming field had not re-equilibrated after the most recent thermotectonic activity at the time of kimberlite sampling and eruption.

Very little high-resolution deep-seismic geophysical work has been done on the Superior Province (see *C.J.E.S. Special Volume 37*, 2000; Rondenay *et al.*, 2000; Ellis *et al.*, 1996). However, Griffin *et al.* (1999b) and Canil (1999b) have shown that trace element chemistries of mantle xenocrysts correlate closely with deep-seismic work in the Slave Province (Northwest Territories). Trace element variation is a robust indicator of mantle structure and lithology in the absence of geophysical data, and may be useful in augmenting existing geophysical profiles elsewhere.

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Prestack time migration of SNORCLE line 1: an improved seismic image of the Slave Province

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Seismic reflection data of SNORCLE line 1 in the vicinity of Yellowknife have been reprocessed to 15 s with Equivalent Offset Migration (EOM). EOM is an efficient prestack time migration algorithm described by Bancroft et al (1998), which is now available for use with IT&A processing software. EOM offers the following advantages over conventional processing: (1) dip moveout effects are reduced, which allows very steeply dipping events to be imaged in the near surface; (2) EOM is highly sensitive to velocity, causing multiples and mode conversions to be suppressed; (3) velocities are specified as imaging velocities at the migrated output locations, which reduces problems with differing velocities for crossing events common to conventional processing. Furthermore, EOM is a 3D algorithm, which helps to reduce the following problems common to crooked line processing: (1) mispositioning of reflections due to 2D migration of a crooked line which crosses 3D structures; (2) loss of stack fold at corners of the slalom line; (3) smearing due to stacking of traces which have common midpoints that lie in bins with large cross-line dimensions. However, since the SNORCLE data was acquired as a 2D data set, sampling in the cross-line direction is poor. This causes some migration noise which is particularly evident in the near surface.

Overall reflection patterns of the EOM result are similar to those of conventional processing. This observation increases our confidence that the traditional processing flow of NMO, stack and poststack migration yields reasonable results, and that the reflections imaged represent primary arrivals rather than mode conversions or multiples. In more detail, the EOM result has better lateral resolution, and the near surface image has been improved. The Y1 reflection of Cook et al. (1999) can be followed much closer to the surface. Reflection Y3 of Cook et al. (1999) is now much more laterally continuous, and cutoff relationships with several events below have been imaged. Very steep dips to the west of Yellowknife in the Defeat pluton may represent a foliation fabric, which dips 65° to 85° to the west.

These results show that EOM is a viable method for prestack migration of transect-scale crustal seismic reflection data. Improved imaging in the near surface will help in tying reflections to mapped surface geology. For future work, EOM will serve as a platform for surface-consistent residual statics estimation and rough topography migration.

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Stochastic Modelling of the Crust-Mantle Boundary

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Deep crustal reflection seismic profiles from different tectonic regimes reveal differing reflectivity characteristics of the crust-mantle boundary or the reflection Moho. For example, the variations in reflectivity can be interpreted as compositional differences in rock properties that define the rocks in the transitional zone. In the absence of rock exposures from this transition zone, it would be useful to quantify the reflectivity characteristics with forward modelling.

We used the finite-difference scheme second-order in both time and space in the heterogeneous formulation (Kelly et al., 1976) to generate synthetic seismograms for models which correspond to compositional differences in the crust-mantle boundary. To exhaust all possible scenarios, we took a stochastic approach to include statistical distributions such as exponential, Gaussian and van Karman to represent the velocity fluctuations. For simplicity, we assumed that the composition is made up of distinctly high-velocity and low-velocity rocks. All numerical examples involve P-SV waves propagating in a 2-D medium. Values of the numerical parameters in the finite-difference calculations were chosen to minimize numerical dispersion and insure stability (Aki and Richards, 1980).

In this presentation, we will provide illustrations of the deep crustal reflection profiles that summarize the different reflectivity types of the crust-mantle boundary, and will introduce the velocity models considered for the finite-difference calculations. The results of our work will be considered with observations made from real data to make inferences about the role that compositional differences play in the reflectivities of the crust-mantle boundary.

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**Lithospheric structure across the craton-Cordilleran
transition of northeastern British Columbia:
A final look at results for Line 21 of SNoRE '97**

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The lithospheric structure of the transition from the craton to the Cordillera in northeastern British Columbia is interpreted from inversion of seismic refraction/wide-angle reflection data along a 460-km profile, and from 3-d inversion and 2.5-d forward modelling of Bouguer gravity data. The seismic profile extends westward from the sediment-covered edge of cratonic North America across the Foreland and Omineca morphogeological belts to the eastern boundary of accreted terranes beyond the Tintina Fault. Across the ancient cratonic margin, the resultant models reveal a westward thickening package of low upper crustal velocities (6.2 km/s and less) and low densities to almost 20 km depth below the Western Canada Sedimentary Basin (WCSB), overlying a west-facing ramp of higher velocities and densities in the middle and lower crust. These features are inferred to represent passive margin sediments deposited on the ancient rifted margin during the mid-to-late Proterozoic and early Paleozoic. A wedge-shaped high velocity (7.3 km/s) crustal layer at the base of the crust beneath the edge of cratonic North America is interpreted to be the result of magmatic underplating during rifting. In the Cordilleran Foreland Belt, high velocities (6.4 km/s) in the upper five kilometers of the crust indicate rocks upthrust from the middle crust. A narrow trench of low velocities in the near-surface, which is imaged ~20 km to the west of the inferred location of the Tintina Fault, is interpreted to represent the actual location of the fault or a major splay. From east to west, the Moho decreases in depth from ~40 km to ~34 km below the rifted margin of ancestral North America, then defines a small root at ~38 km depth below the high topography and upper crustal velocities of the eastern Foreland Belt, and gradually shallows to ~34 km beneath the Omineca belt. An enigmatic laterally heterogeneous upper mantle has anomalously high velocities (up to 8.3 km/s) beneath the Foreland Belt flanked by regions of low velocities (7.7-7.8 km/s). Results indicate that the location of the Cordilleran deformation front west of the ramped cratonic margin directly affected the tectonic evolution of the region.

**Preliminary results from SNORCLE magnetotelluric data:
Johnson's Crossing to Upper Laird**

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INTRODUCTION

The LITHOPROBE SNORCLE (Slave-NORTHERN Cordillera Lithoprobe Evolution) Transect profile 200 (Corridor 2) is located in northern British Columbia and southern Yukon Territory. A total of 85 magnetotelluric (MT) stations along the Alaska Highway and highway 37 (BC) (Fig. 1) were deployed that cross over a very diverse region that has experienced substantial deformation and tectonism. This paper will focus on 16 MT stations (Group 1) that were measured from Johnson's Crossing to Upper Laird (Fig. 1). The primary goal of the MT study is to define the regional conductivity structure in relation to the major tectonic units.

REGIONAL GEOLOGY

The tectonic framework of the Cordillera is described in terms of morphogeological belts that make up tectonostratigraphic rock assemblages and component terranes. The belts vary in thickness, from 3 km in the west to 50 km in the east (Gabrielse et al., 1991). The morphogeological belts reflect the sum of geological processes that have occurred from middle Proterozoic to Cenozoic time. Each belt has different physiography, metamorphism, volcanism, plutonism, and structural styles, to which were in part inherited from their original terranes (Gabrielse et al., 1991).

The Cordillera can be subdivided into two distinct parts: the northeast part that encompasses the North American continent and southwest part that contains the accreted terranes. The nature and evolution of the terranes and its boundaries are complex and still controversial (Wheeler et al., 1991; Gabrielse and Yorath, 1992; and Gordey and Makepeace, 1999). Separating the two parts is the Tintina Fault, a northwest trending transcurrent fault that has had a minimum of 450 km dextral displacement since the Cretaceous (Lowe et al., 1994).

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.

Most rock forming minerals are poor conductors. High conductivity in rocks is from ionic and/or electronic conduction. The former occurs when ions are free to move through electronically conductive solids, such as sulphides, graphites, or iron oxides. The bulk electrical resistivity depends on both the amount of ionic and electronic conduction, as well as on their geometric distribution within the rock. This provides a method that can give unique information on the geodynamic processes than from other geophysical methods.

GEOELECTRIC STRIKE DIRECTIONS

Strike direction varies with frequency and is only consistent in surrounding areas. From initial examination of the data, it was apparent that a single regional strike at all frequencies would not be appropriate for the analysis. Instead, grouping sites based on best fits was carried out to determine the strike.

Five major groups were defined based on Groom-Bailey strike decompositions (Fig. 1). Group 1 MT stations were deployed on several accreted terranes, west of the Tintina fault. From west to east the accreted terranes are Kootenay, Slide Mountain, Dorsey, Slide Mountain (again), and Cache Creek.

GROUP 1: STRIKE ANALYSIS

Further strike analysis was performed via multi-site decompositions (McNeice and Jones, 2001) and plotting rose diagrams (Fig. 2). The strike varied as a function of period, with shorter periods (near-surface response) having a strike of N50°W and longer periods (upper and lower crust response) having a strike of N60°W. However, a comparison of the phase and apparent resistivity plots as a function of rotation revealed that near surface structures are similar. Therefore, a strike of N65°W was adopted for the group.

For the 2D case the EM fields will separate into TE (transverse electric) and TM (transverse magnetic) modes. The TE mode involves electric current flowing parallel to the structure. This is now defined as parallel to N65°W. The TM mode involves electric current flowing perpendicular to the strike.

GROUP 1: INDUCTION VECTORS AND MT PHASE PSUEDOSECTIONS

Induction vectors are derived from the transfer function between the vertical component of the magnetic field and the horizontal components. For uniform EM source fields the magnitude of the tipper and induction vector is close to zero, except near lateral conductivity contrasts. The induction vectors point toward regions of enhanced conductivity.

Induction vectors were consistent with the strike determinations (Fig. 3). At shorter periods, corresponding to signal penetration into the upper crust, the induction vectors have irregular orientations, indicating that the response is affected by local

geology. At middle periods, corresponding to signal penetration in the middle/lower crust, the response is more uniform, indicating that all sites are responding to a larger-scale structure at this depth.

Pseudosections are contour plots, with the vertical axis corresponding to increasing period and the horizontal axis to profile location. Frequencies increasing over 45° indicate areas with increasing conductivity. From examination of the phase determinant (Fig. 4) and phase (Fig. 5) pseudosections there could be a conductor in the middle of the profile at depths in the middle and lower crust. The phase determinant plot has high phases >70 degrees at sites 279-281 for periods >1 second. Similarly, the phase pseudosection has high phases for both the TE and TM modes in the same region. The enhanced conductivity region is located beneath the Dorsey terrane.

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Terranes and Tectonic Elements of Northwestern North America

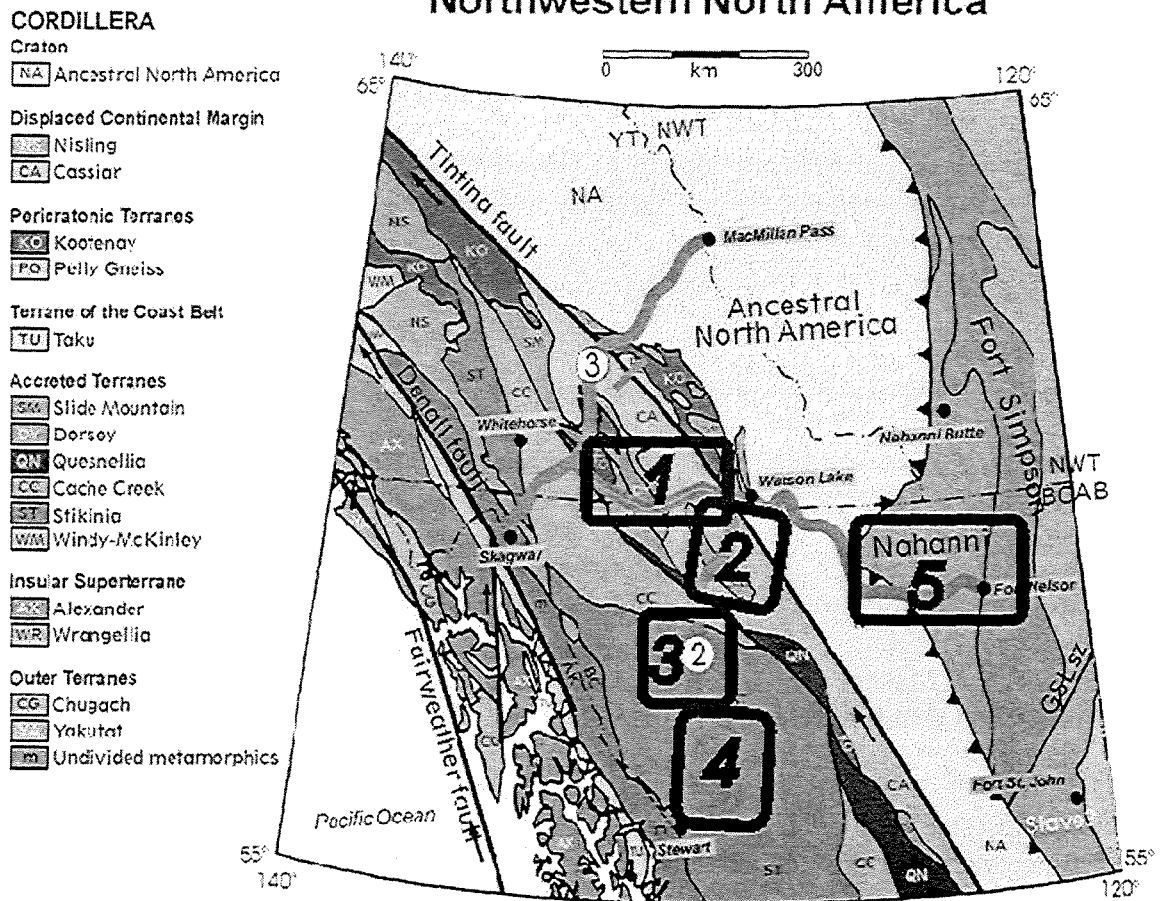


Figure 1. Terrane map of northwestern North America that shows the location of SNORCLE corridors 2 and 3. The boxes show groups of sites based on strike analysis. Faults are heavy dark lines.

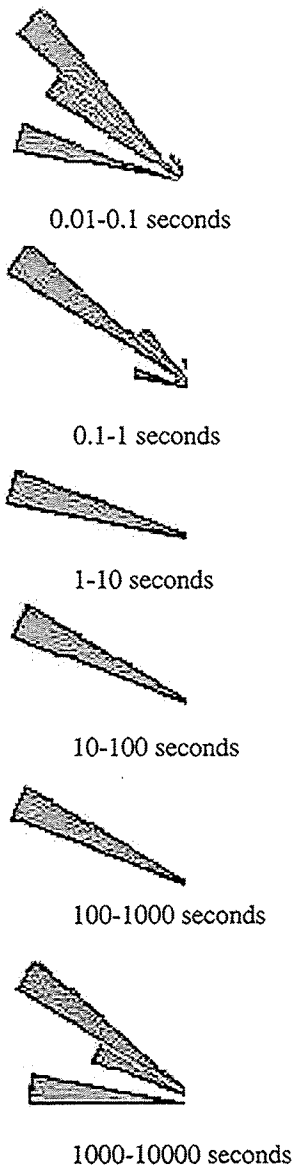


Figure 2. Rose diagrams for Group 1. The results are weighted average error, indicating the direction of strike.

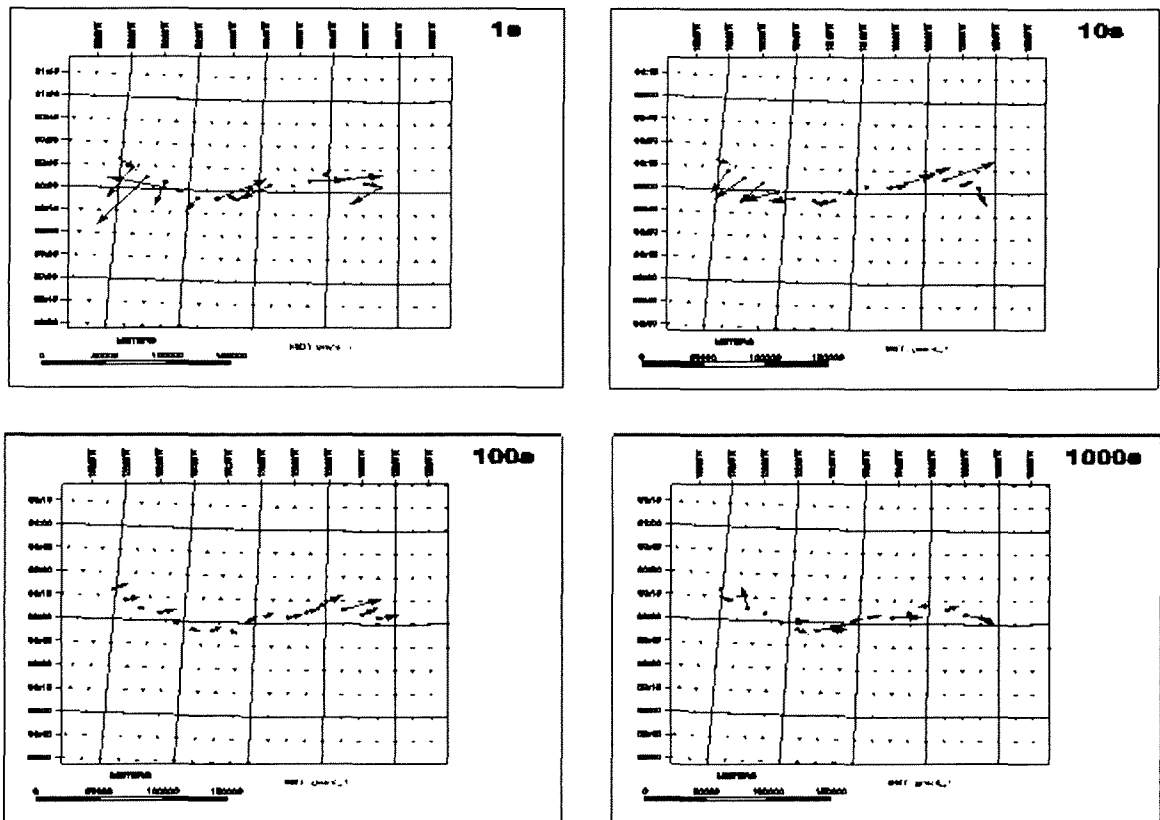


Figure 3. Induction vector plots for periods 1, 10, 100 seconds, and 1000 seconds.

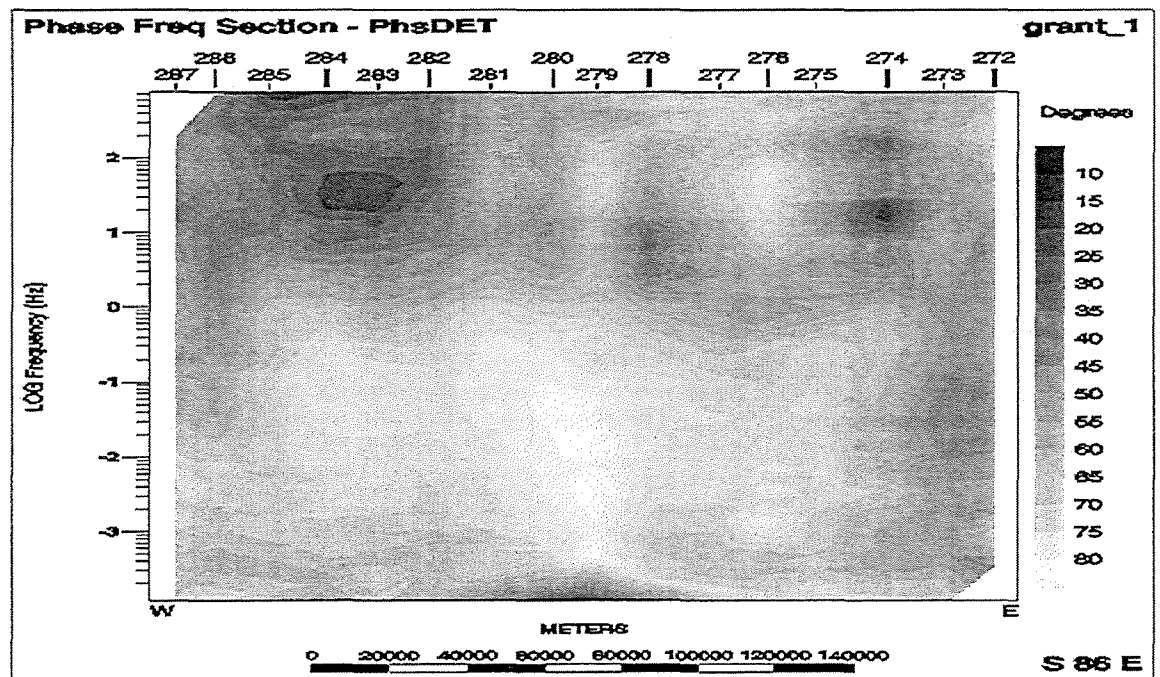


Figure 4. Phase determinant pseudosection for MT stations in group 1.

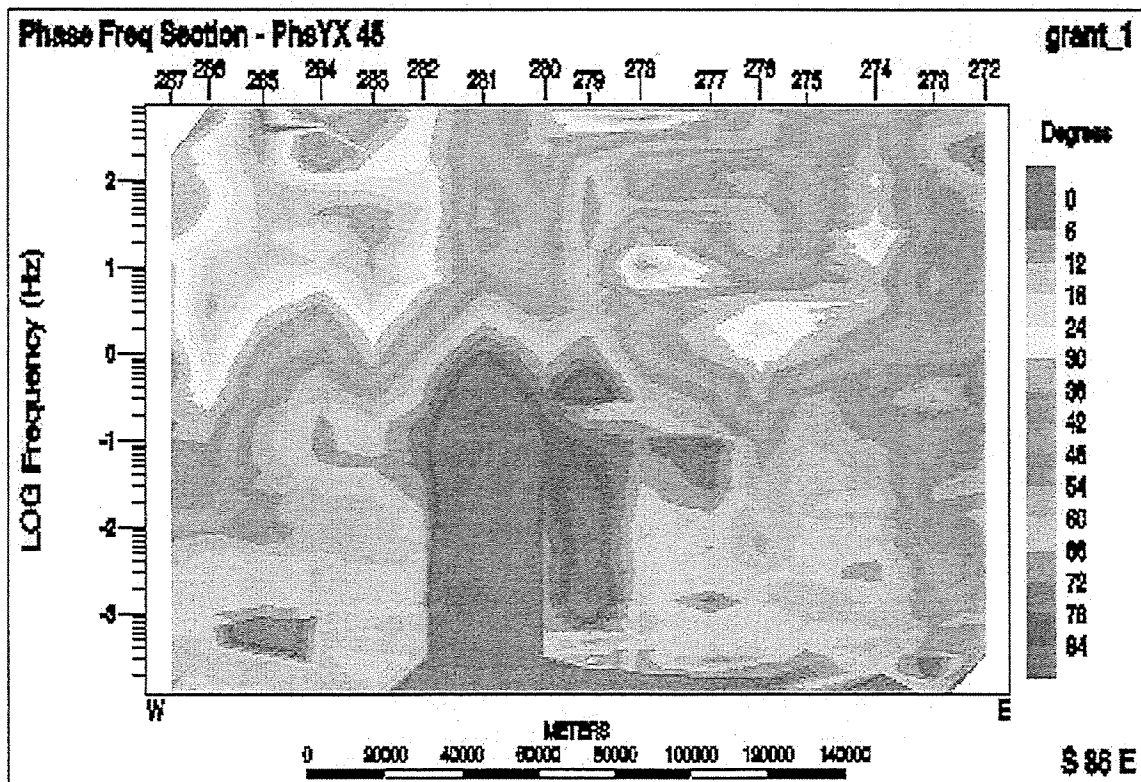
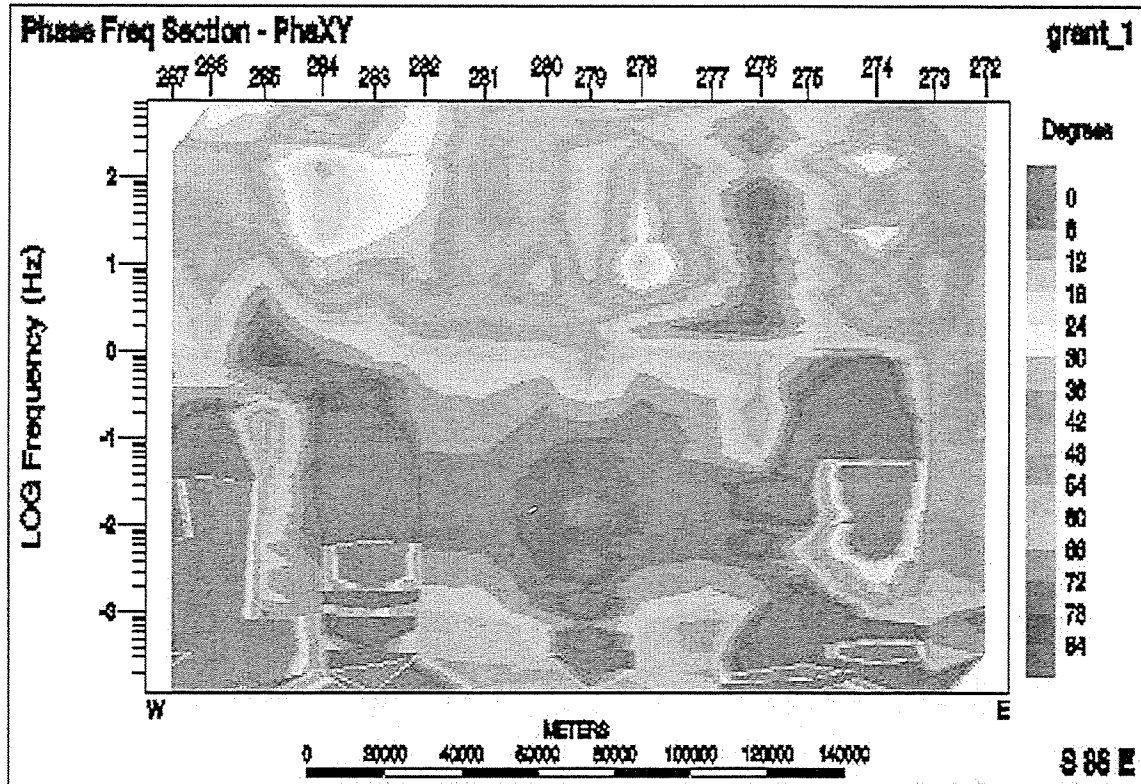


Figure 5. Phase pseudosections for MT stations in group 1.

