Precambrian crust beneath the Mesozoic northern Canadian Cordillera discovered by Lithoprobe seismic reflection profiling

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Received 13 May 2002; revised 9 April 2003; accepted 1 August 2003; published 30 March 2004.

[1] The Cordillera in northern Canada is underlain by westward tapering layers that can be followed from outcrops of Proterozoic strata in the Foreland belt to the lowermost crust of the orogenic interior, a distance of as much as 500 km across strike. They are interpreted as stratified Proterozoic rocks, including ~1.8–0.7 Ga supracrustal rocks and their basement. The layering was discovered on two new deep seismic reflection profiles in the Yukon (Line 3; ~650 km) and northern British Columbia (Line 2; ~1245 km in two segments) that were acquired as part of the Lithoprobe Slave-Northern Cordillera Lithospheric Evolution (SNORCLE) transect. In the Mackenzie Mountains of the eastern Yukon, the layering in Line 3 is visible between 5.0 and 12.0 s (~15 to 36 km depth). It is followed southwestward for nearly 650 km (~500 km across strike) and thins to less than 1.0 s (~3.0–3.5 km thickness) near the Moho at the Yukon-Alaska international boundary. In the northern Rocky Mountains of British Columbia, the upper part of the layering on Line 2 correlates with outcrops of Proterozoic (1.76–1.0 Ga) strata in the Muskwa anticlinorium. At this location, the layering is at least 15 km thick and is followed westward then southward into the middle and lower crust for ~700 km (~300 km across strike). It disappears as a thin taper at the base of the crust ~150 km east of the coast of the Alaskan panhandle. The only significant disruption in the layering occurs at the Tintina fault zone, a late to postorogenic strike-slip fault with up to 800 km of displacement, which appears as a vertical zone of little reflectivity that disrupts the continuity of the deep layering on both profiles (~300 km apart). The base of the layered reflection zone coincides with the Moho, which exhibits variable character and undulates in a series of broad arches with widths of ~150 km. In general, the mantle appears to have few reflections. However, at the southwest end of Line 3 near the Alaska-British Columbia border, a reflection dips eastward from ~14.0 s to ~21.0 s (~45 to 73 km depth) beneath exposed Eocene magmatic rocks. It is interpreted as a relict subduction surface of the Kula plate. Our interpretation of Proterozoic layered rocks beneath most of the northern Cordillera suggests a much different crustal structure than previously considered: (1) Ancient North American crust comprising up to 25 km of metamorphosed Proterozoic to Paleozoic sediments plus 5–10 km of pre-1.8 Ga crystalline basement projects westward beneath most of the northern Canadian Cordillera. (2) The lateral (500 km by at least 1000 km) and vertical (up to 25 km) extent of the Proterozoic layers and their internal deformation are consistent with a long-lived margin for northwestern North America with alternating episodes of extension and contraction. (3) The detachments that carry deformed rocks of the Mackenzie Mountains and northern Rocky Mountains are largely confined to the upper crustal region above the layering. (4) Accreted terranes include thin klippen that were thrust over North American pericratonic strata (e.g., Yukon-Tanana), and terranes such as Nisling and Stikinia that thicken westward as the underlying Proterozoic layers taper and disappear. (5) The ages of exposed rocks are not necessarily indicative of the ages of underlying crust, a frequent observation in Lithoprobe interpretations, so that estimates of crustal growth based on surface geology may not be representative. *INDEX TERMS:* 7205 Seismology: Continental crust (1242); 8102 Tectonophysics: Continental contractional orogenic belts; 8110 Tectonophysics: Continental tectonics—general (0905); *KEYWORDS:* crustal structure, seismic reflection, northern Cordillera. **Citation:** Cook, F. A., R. M. Clowes, D. B. Snyder, A. J. van der Velden, K. W. Hall, P. Erdmer, and C. A. Evenchick (2004), Precambrian crust

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0278-7407/04/2002TC001412$12.00

beneath the Mesozoic northern Canadian Cordillera discovered by Lithoprobe seismic reflection profiling, *Tectonics*, *23*, TC2010, doi:10.1029/2002TC001412.

1. Introduction

[2] Application of seismic reflection profiling to the study of the lithosphere in northwestern Canada has led to the discovery of regionally extensive middle and lower crustal layering that can be traced both directly and indirectly to outcrops of Proterozoic strata [*Snyder et al.*, 2002]. The data were recorded for Lithoprobe in three lines (Figures 1 and 2): Line 2b (~570 km) from Fort Nelson, British Columbia northwestward to ~60 km west of Watson Lake, Yukon; Line 2a (~675 km) from ~40 km north of Watson Lake, Yukon southward to Stewart, British Columbia; and Line 3 (~650 km) along the South Canol road from Macmillan Pass, Yukon to White Pass at the British Columbia-Alaska international boundary east of Skagway. Thus one profile crosses nearly the entire Cordillera in northern British Columbia (Lines 2b and 2a) and the other crosses the Cordillera in the Yukon from the central Mackenzie Mountains to the Coast Mountains (Line 3).

[3] The profiles are ~300–400 km apart and provide a rare opportunity to compare and contrast cross sections of the orogen in regions which have common regional structures as well as some important differences. When combined with more than 700 km that were recorded previously in the northwestern Canadian Shield [*Cook et al.*, 1999], these data form the seismic reflection component of the Lithoprobe Slave-Northern Cordillera Lithospheric Evolution (SNORCLE) transect. This transect extends from some of the oldest rocks in the world, those of the Slave Craton (>2.6–4.0 Ga), to the late Tertiary Coast Mountains (Figure 1). The new profiles cross the following major tectonic features: the western part of the Western Canada Sedimentary Basin, the Foreland belt of the northern Rocky Mountains and western Mackenzie Mountains, the strike-slip Tintina-Northern Rocky Mountain Trench fault system, accreted terranes of the orogenic hinterland, the Late Cretaceous Skeena fold belt, and the Cretaceous-Tertiary magmatic belt of the Coast Mountains. We begin by describing the regional tectonic setting.

2. Regional Setting

[4] The Cordillera in northern Canada developed on the western margin of North America during lithospheric plate interactions from the Devonian through the early Tertiary. Surface rocks consist of North American marginal strata that range from Paleoproterozoic to Paleozoic in age, pericratonic and allochthonous terranes that were accreted to western North America during Mesozoic-Tertiary orogenesis, and synorogenic to post orogenic igneous and sedimentary rocks [*Gabrielse et al.*, 1991]. The main collisional phase of the Cordillera resulted in east-northeast contraction [e.g., *Gabrielse et al.*, 1991] and the development of a series of late northwest striking dextral strike-slip faults, the most prominent of which is the Tintina–Northern Rocky Mountain trench fault system which displaced earlier structures (Figure 1) [*Gabrielse*, 1985]. As Proterozoic layered strata were probably deposited on crystalline arc rocks of the 2.1–1.8 Ga Wopmay Orogen, we first describe the completion of Wopmay orogenesis.

2.1. Proterozoic Wopmay Orogen (Circa 2.1–1.8 Ga)

[5] The Wopmay Orogen is a north striking Paleoproterozoic orogen that developed as a consequence of terrane accretion and arc magmatism on the west margin of the Archean Slave craton between about 2.1–1.8 Ga [*Hoffman*, 1980; *Hildebrand et al.*, 1987]. At ~1.84 Ga the Nahanni-Fort Simpson terrane accreted to the western margin of North America that earlier had been extended westward by Paleoproterozoic accretion of the Hottah terrane to the Archean Slave Craton at ~1.88 Ga [*Hildebrand et al.*, 1987]. The Nahanni-Fort Simpson collision involved east dipping subduction and tectonic delamination of the Fort Simpson crust [*Cook et al.*, 1999]. Fort Simpson magmatic arc rocks (~1.85 Ga) have been identified in drill cuttings from northeastern British Columbia and the southern Northwest Territories [*Villeneuve et al.*, 1991]. The drill hole locations coincide with regional north-south striking magnetic and gravity anomalies (Fort Simpson anomalies [*Hoffman*, 1987]) that appear to delineate a series of en echelon structural culminations (Fort Simpson structural trend [*Cook et al.*, 1992]). Westward thickening layers observed on deep seismic reflection data west of the Fort Simpson anomalies are interpreted as primarily Proterozoic basinal strata (Fort Simpson Basin [*Cook and van der Velden*, 1993; *Cook et al.*, 1999]) that are in turn overlain by Early Paleozoic rocks of the Western Canada Sedimentary Basin.

2.2. Post-Wopmay Proterozoic Supracrustal Rocks (Circa 1.8–0.6 Ga)

[6] Proterozoic supracrustal rocks deposited in the interval 1.8–0.6 Ga are correlated from the Canadian Shield to Alaska, and from the Arctic islands to the subsurface of the Western Canada Sedimentary Basin. Strata deposited within this time interval have been subdivided into three major sequences [*Young et al.*, 1979; *Young*, 1984; *Thorkelson et al.*, 1998, 2001): A (>1.72 Ga–1.1 Ga); B (~1.1 Ga–0.77 Ga); and C (~0.77 Ga–0.55 Ga). Although later studies have illustrated complexities in the stratigraphic and structural history for this time interval [*Thorkelson et al.*, 1998], the threefold subdivision provides a useful framework for analyzing large-scale profiles and relating gross features over much of the region.

[7] In exposed rocks of the northwestern Canadian Shield, strata were deposited on older rocks of the Wopmay Orogen and thicken westward to ~7 km prior to disappearing beneath Phanerozoic strata of the Interior platform [*Baragar and Donaldson*, 1973; *Ross and Kerans*, 1989]. Strata that correlate with the youngest of these layers are exposed in the Mackenzie Mountains as the Mackenzie Mountains Supergroup (Sequence B) and Windermere Supergroup (Sequence C), which together are ~7 km thick. Farther west, coeval rocks are underlain by almost 14 km of Wernecke Supergroup strata (Sequence A) in the Wernecke

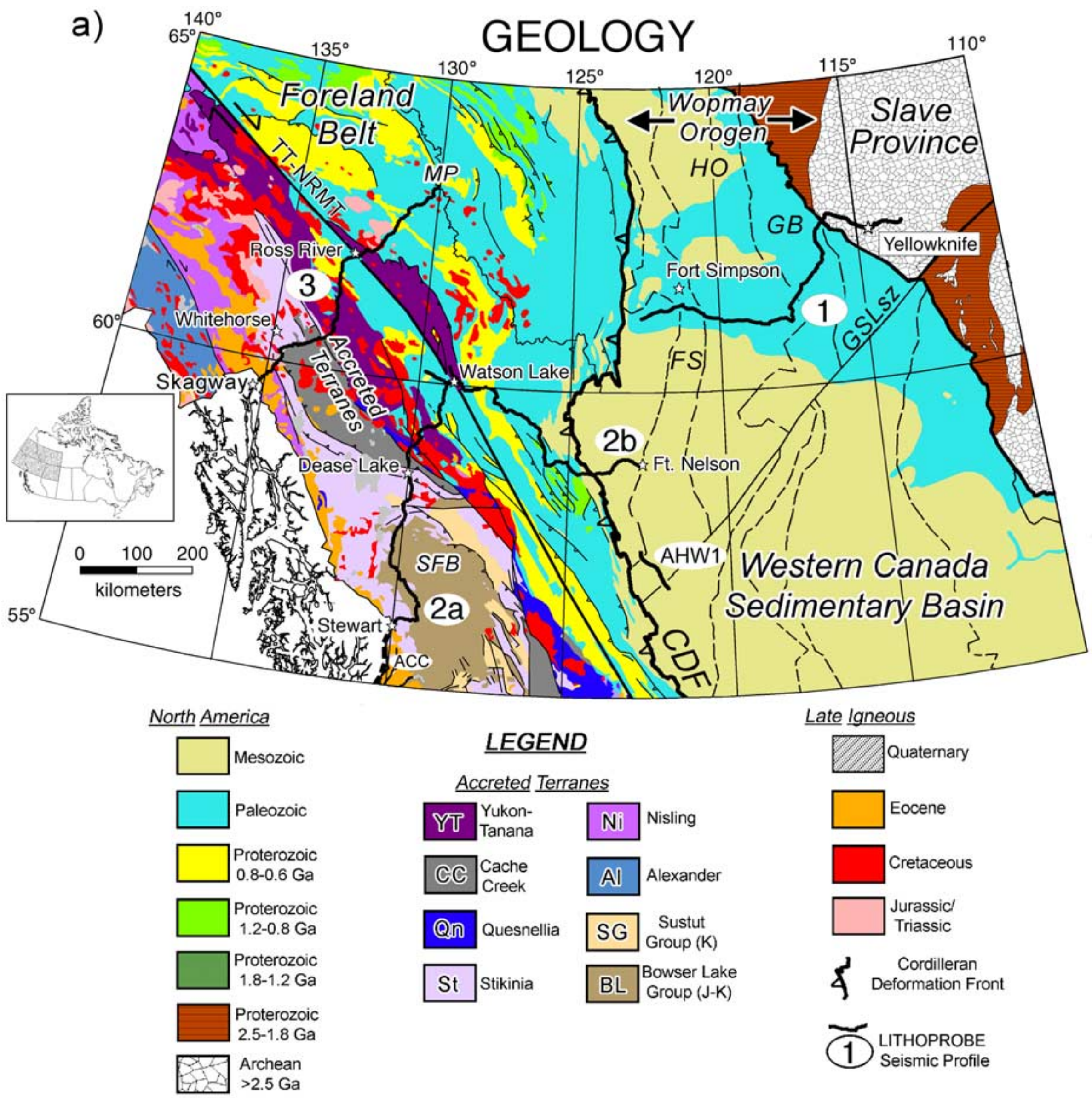


Figure 1. (a) Geological map of northwestern Canada (modified from *Wheeler and McFeely* [1991]; *Geological Survey of Canada*, 1999c]) showing the locations of the Lithoprobe SNORCLE profiles 1, 2a, 2b, and 3. Colors and patterns used here are consistent with other figures in the paper. Dashed lines in the Western Canada Sedimentary Basin represent Proterozoic domains identified on the basis of potential field data. North of the Great Slave Lake shear zone (GSLsz) these domains are part of the Wopmay Orogen and include the Great Bear magmatic arc (GB), the Hottah terrane (HO) and the Fort Simpson arc (FS). The lines south of GSLsz are Precambrian domains not discussed here. CDF, Cordilleran deformation front; AHW-1, Alaska Highway reflection profile [*Cook and van der Velden*, 1993]; MP, MacMillan Pass; SFB, Skeena Fold Belt, ACC, ACCRETE refraction profile of *Morozov et al.* [1999]. (b) Enlargement of the geological map in the vicinity of Lines 2a and 2b. Numbers along the roads identify station (also vibrator point, or VP) numbers (one station every 50 m). KF, Kechika fault; KS, King Salmon fault; NA, Nahlin fault; TT-NRMT, Tintina-Northern Rocky Mountain Trench fault system; TH, Thibert fault. (c) Enlargement of the geological map in the vicinity of Line 3. Numbers along the roads and TT-NRMT same as Figure 1b. MP, MacMillan Pass; PF, Plateau fault; TZ, Teslin zone; WP, White Pass.

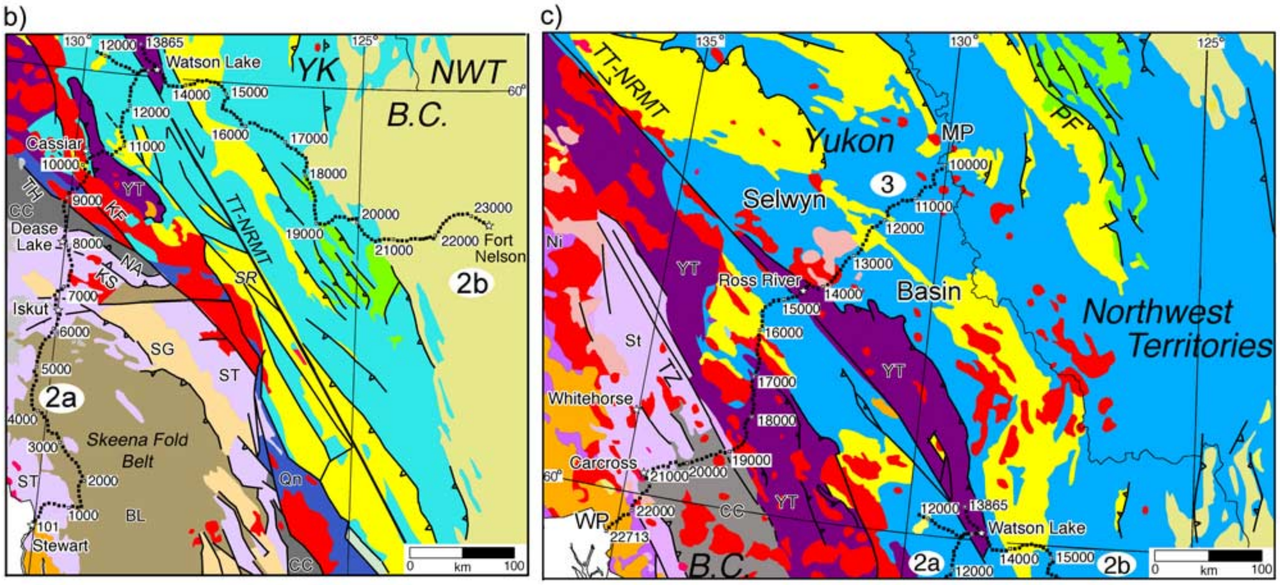


Figure 1. (continued)

and Ogilvie Mountains of the western Yukon [*Delaney*, 1981; *Cecile and Cook*, 1981]. From all available information, the composite thickness of these strata in the northwestern Mackenzie Mountains is ~20–21 km [*Young et al.*, 1979]. A suggestion we make here is that many of these rocks underlie much of the northern Cordillera, although the thicknesses of each sequence may vary along and across depositional strike [see also *Cecile and Cook*, 1981].

[8] At least three episodes of contractional deformation affected these strata: at ~1.7–1.6 Ga (Racklan orogeny); between 1.2 Ga and 1.0 Ga [*Hildebrand and Baragar*, 1991]; and between 0.9 Ga and 0.75 Ga [*Taylor and Stott*, 1973; *Thorkelson et al.*, 1998]. Some of the structures associated with these events are visible on regional seismic reflection profiles [e.g., *Cook*, 1988a; *Cook and Maclean*, 1995], although detailed correlations of seismic reflections to known strata are difficult at the present time.

2.3. Late Proterozoic–Paleozoic (550 Ma–250 Ma)

[9] During the Late Proterozoic and Early Paleozoic, a complex west facing Atlantic-type margin overlapped Proterozoic supracrustal rocks and their depositional basement [e.g., *Gabrielse et al.*, 1991]. Passive margin sedimentation continued until the late Devonian, when the source region of the detritus shifted to the west as a result of regional uplift and orogenic activity [e.g., *Gordey et al.*, 1987]. Evidence of Late Devonian tectonism in the northern Canadian Cordillera includes deformed Paleozoic rocks in structural windows within some of the accreted terranes, and local uplifts that predate Mississippian and younger sedimentation [*Bell*, 1973; *Coflin et al.*, 1990; *Gabrielse et al.*, 1991].

[10] Paleozoic sedimentation, volcanism and deformation affected rocks that form much of the western Cordillera [e.g., *Monger*, 1977a, 1977b; *Monger et al.*, 1991; *Evenchick*, 1991a]. These rocks were horizontally shortened during the Mesozoic.

2.4. Mesozoic (250–65 Ma)

[11] Most of the exposed structures of the northeastern Cordillera were formed during Mesozoic contraction and Late Mesozoic–Early Cenozoic transcurrent motion. Northeast verging thrust faults and folds of the Foreland belt formed when terranes, including Cache Creek, Yukon–Tanana and Stikinia [*Gabrielse et al.*, 1991], converged with North America. Whether these terranes were originally far from the margin, as suggested by interpretations of faunal assemblages [e.g., *Gabrielse et al.*, 1991] and paleomagnetic data [e.g., *Irving et al.*, 1996], or whether they were relatively close to the margin, as suggested for some of them by stratigraphic studies [e.g., *Nelson and Mihalynuk*, 1993; *Mihalynuk et al.*, 1994], remains unclear. During the latest Cordilleran activity, the accreted terranes were also involved in the deformation in a style similar to the Foreland belt, with northwest striking contractional structures forming across the width of the Cordillera [e.g., *Evenchick*, 1991a]. Convergence was dextral, normal, and sinistral during this period [*Engebretson et al.*, 1985] with the sinistral convergence ending in the mid-Cretaceous [*Kelley*, 1993].

2.5. Cenozoic

[12] Postorogenic activity in the Cenozoic consisted primarily of strike-slip displacement along the Tintina–Northern Rocky Mountain Trench fault system which began in the late Mesozoic, and volcanism that continued into the Holocene. The amount of translation observed on the Tintina–Northern Rocky Mountain Trench and related structures is uncertain; estimates range up to 750–800 km

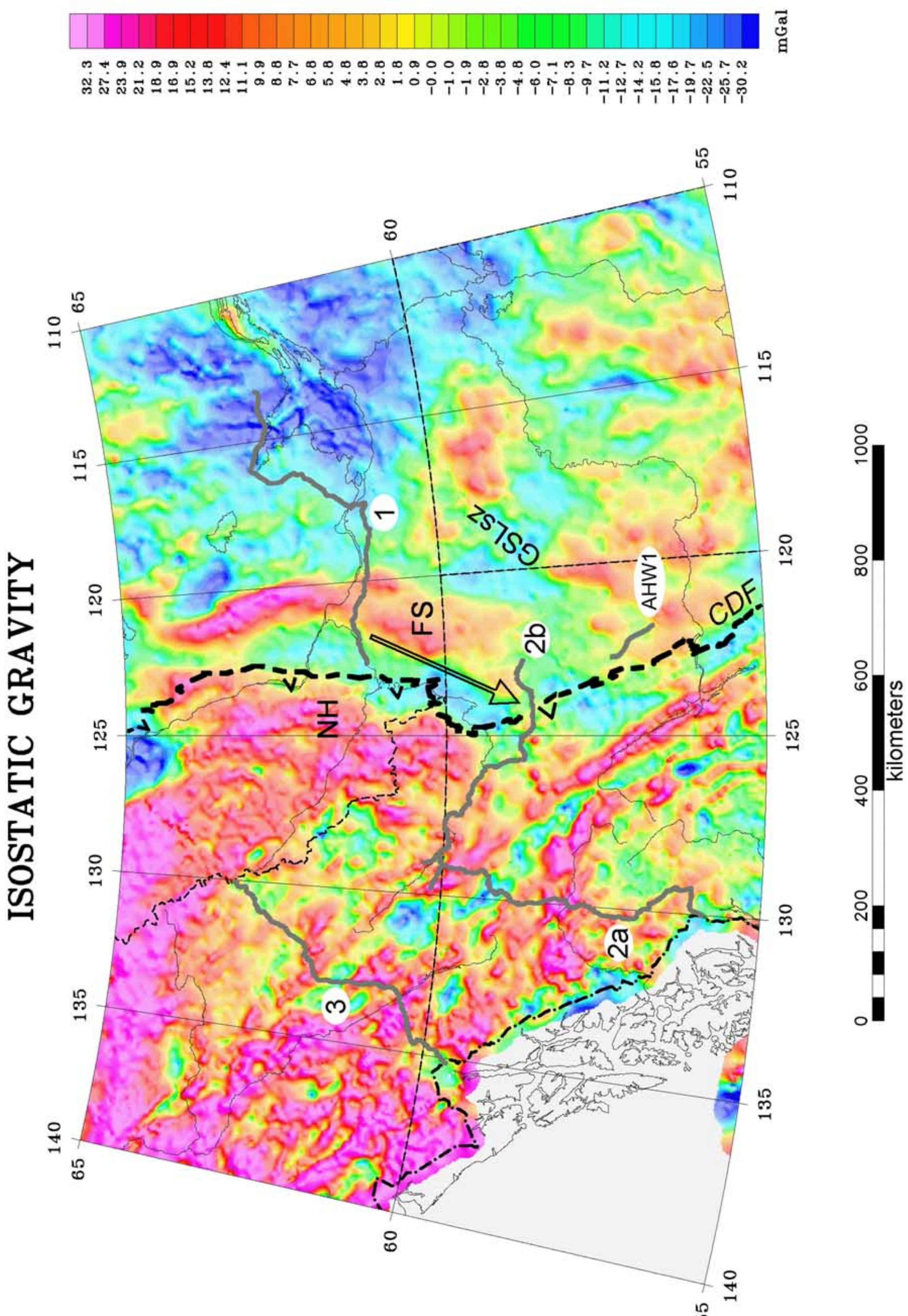


Figure 2. Map of the isostatic gravity [*Geological Survey f Canada*, 1999a] illustrating the correlation from Line 1 to Line 2b. The Fort Simpson anomaly trend (FS) projects southwestward from Line 1 such that the west side of FS, which coincides with the thick Fort Simpson basin on Line 1, projects to the Muskwa anticlinorium along Line 2b (open arrow). Isostatic anomalies calculated assuming Airy compensation to 30 km [*Goodacre et al.*, 1987].

[e.g., *Gabrielse*, 1985], although the preferred estimates are ~400–450 km [*Gordey and Makepeace*, 1999]. Cenozoic magmatic activity included Eocene volcanic and intrusive rocks in northwestern British Columbia and southwestern Yukon that may have been related to the subduction of the Kula plate [*Morris and Creaser*, 1998]. Alkali basalts as young as Holocene cross the Cordillera from near the coast to as far east as Watson Lake, Yukon [*Shi et al.*, 1998; *Edwards and Russell*, 1999; *Francis et al.*, 1999]. The origin of these volcanic rocks is uncertain. They have been interpreted to be from a mantle plume [*Shi et al.*, 1998] or to define an incipient volcanic province from extension similar to the Basin and Range [*Edwards and Russell*, 1999].

3. Previous Geophysical Studies

3.1. Crustal Reflection Profiles

[13] Crustal reflection profiles recorded east of the new Lithoprobe data include SNORCLE profile 1 [*Cook et al.*, 1999], an industry profile south of Fort Nelson (AHW-1 on Figure 1 [*Cook and van der Velden*, 1993]), and several short industry profiles east of the Mackenzie Mountains [*Mitchelmore and Cook*, 1994]. SNORCLE-1 and AHW-1 provided images of layers that thicken westward to more than 20 km beneath the Phanerozoic Western Canada Sedimentary Basin [*Cook and van der Velden*, 1993; *Cook et al.*, 1999]. These layers subcrop beneath the Phanerozoic strata west of Fort Simpson domain in the western Wopmay Orogen (Figure 1a). East of the subcrop location, two wells were drilled from Phanerozoic sediments into 1.85 Ga granitic rocks of the Wopmay Orogen [*Villeneuve et al.*, 1991], whereas uplifted rocks near the eastern front of the Cordillera have more than 6 km of intervening strata (Muskwa assemblage and Lower Paleozoic [*Bell*, 1968; *Taylor and Stott*, 1973; *Ross et al.*, 2001]). Accordingly, although the exact thicknesses of Proterozoic strata in this area not known, much of the layered reflectivity is interpreted as post-Wopmay Proterozoic strata (Fort Simpson Basin [*Cook and van der Velden*, 1993; *Cook et al.*, 1999]).

[14] On the west, the ACCRETE program in the United States addressed the structural and tectonic evolution of the Coast Range and offshore regions [*Morozov et al.*, 2001] by recording marine seismic data from the Pacific Ocean into fjords that cross the western Coast Range. The ACCRETE profiles are located west of Line 2a and end near Stewart, British Columbia [*Morozov et al.*, 1999, 2001]. The closest profile to the SNORCLE data was recorded within Portland Canal southwest of the west end of Line 2a (Figure 1b). The ACCRETE data provide images of structures, interpreted to be a consequence of late to post-orogenic extension, which project beneath the western end of SNORCLE Line 2a.

3.2. Crustal Refraction Profiles

[15] Crustal refraction data were recorded along the SNORCLE corridors to provide a framework of crustal thickness and regional seismic velocity variations. In the east, *Welford et al.* [2001] (Figure 3) presented a profile that nearly follows reflection Line 2b. Key observations from these data are (1) the Moho shallows from ~38 depth km near Fort Nelson to ~34 km depth beneath the "Fort Simpson Basin," deepens to ~38 km beneath the Foreland belt, and then rises to ~34 km near the Tintina fault; (2) there is a region of high velocity in the upper crust in the eastern Foreland belt that occurs in the vicinity of the Muskwa anticlinorium and has been interpreted as an uplift of midcrustal rocks [*Welford et al.*, 2001]; and (3) near the Tintina fault, the refraction velocity structure shows a low velocity "notch" attributed to the fault zone [*Welford et al.*, 2001; *Clowes et al.*, 2001]. The crust may thin by ~1–2 km in the vicinity of the Tintina fault zone, and intracrustal velocities appear to be offset across it [*Clowes et al.*, 2001]. Westward thinning of the Moho near the east edge of the Fort Simpson basin is consistent with results from reflection data to the north [*Cook et al.*, 1999] and to the south [*Cook and van der Velden*, 1993].

[16] Profile 2a is also coincident with a SNORCLE refraction profile. *Hammer and Clowes* [1999] and *Clowes et al.* [2001] interpreted the Moho at ~35–36 km depth west of the Tintina fault and beneath accreted rocks of Stikinia, and proposed that the crust thins by ~4–5 km westward into the Coast Mountains [*Hammer et al.*, 2000].

[17] Preliminary analyses of a refraction profile that was recorded along about two-thirds of reflection Line 3 (MacMillan Pass to ~50 km west of Teslin) delineate the Moho at ~34–37 km depth [*Creaser*, 2000]. The Moho is deepest (~37 km depth) ~25–35 km west of the Tintina fault and rises to ~34 km depth near the Teslin zone. There is no obvious change in the Moho across the Tintina fault, although crustal velocities appear to be higher to the west of the fault. From MacMillan Pass to the Tintina fault, the refraction interpretation includes a wide-angle reflection near 20 km depth. As shown below, this depth corresponds to a prominent transition on the reflection profile.

3.3. Regional Isostatic Gravity Data

[18] Potential field data (gravity and magnetic) are available for most of the region covered by the reflection profiles (e.g., Figure 2). Features that are regionally significant, though not entirely understood, include the following. To the east of the Cordillera, both magnetic and gravity anomalies delineate patterns associated with Precambrian rocks beneath the Western Canada Sedimentary Basin (e.g., FS, GSLsz on Figure 2). The Fort Simpson trend (FS) is subparallel to the eastern front of the Mackenzie Mountains, but appears to merge with the northwest striking anomalies of the Cordillera near 57°–58°N. When these data are bandpass filtered, additional subparallel anomalies are found within the Mackenzie Mountains that correlate with structural highs within the crust, one of which appears to project into Line 2b ~250 km west of Fort Nelson (NH on Figure 2 [*Cook*, 2000]). The Mackenzie Mountains and the northwestern part of the region exhibit one of the largest isostatic gravity highs in North America (Figure 2). This high is also visible in the Bouguer gravity anomalies [*Lowe et al.*, 1994]; its origin is uncertain. The Tintina fault zone appears as a linear anomaly (Figure 2) that may be a residual

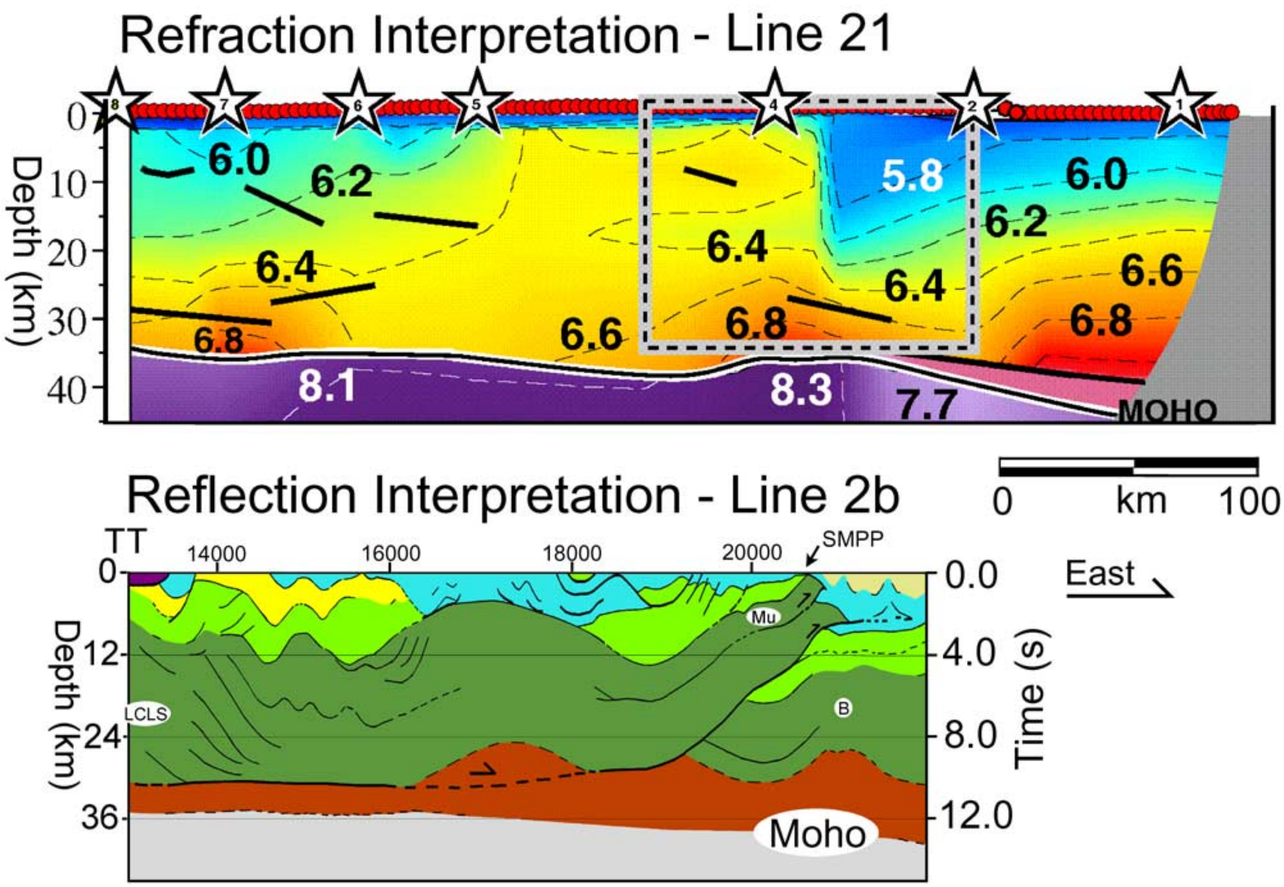


Figure 3. Interpretation of regional crustal refraction profile east of the Tintina fault zone (top; from *Welford et al.* [2001]) compared with interpretation of SNORCLE line 2b near the eastern front of the Cordillera (Foldout 1), both sections plotted with vertical exaggeration of 2.5 to 1. Blue box highlights prominent velocity changes. Colors and numbers indicate *P* wave velocities (km/s). Solid lines indicate wide-angle reflections. Note the uplifted layers east of Stone Mountain Provincial Park (SMPP) on the reflection profile that coincide with high refraction velocities near the surface west of shot point 4. Stars, refraction shot points; Mu, Muskwa complex; B, anticline B described in text; LCLS, Lower crustal layered sequence described in text.; TT, Tintina-Northern Rocky Mountain Trench, NH, Nahanni high; CDF, Cordilleran deformation front; GSLsz, Great Slave Lake shear zone; gray area, no data.

topographic effect rather than a significant subsurface density contrast [*Geiger and Cook*, 2001].

4. Lithoprobe Seismic Reflection Profiles

[19] Reflection data acquisition took place between August 1999 and February 2000. Acquisition parameters (Table 1) were similar to the previous Lithoprobe survey along Line 1, with an increase in the number of channels from 404 to 576, a shorter station spacing (50 m versus 60 m for Line 1), and a vibrator source spacing of 75 m (versus 90 m for Line 1 [*Cook et al.*, 1999]). Processing followed a relatively standard sequence (Table 2). The final step in the processing sequence was coherency filtering [e.g., *Milkereit and Spencer*, 1989] to enhance coherent signal. This allows the most significant reflection events to be visible at scales that are appropriate for regional interpretations.

Table 1. Acquisition Parameters for the 1999–2000 SNORCLE Data

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	12 per station
Vibrator point spacing	75 m
Geophone frequency	10 Hz
Nominal fold	192
Instruments	I/O 2000, 24 bit

Table 2. Processing Sequence and Parameters for the 1999–2000 SNORCLE Data

Parameter	Value
Extended cross correlation	32 s record extended past 12 s
Crooked line geometry	25 × 2000 m bins
Notch filter, where necessary	60 Hz
Trace edits, first break picks	
Refraction statics computation	two layer
Sort to CDP	
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 1% prewhitening
Statics application	
Normal move out correction on common midpoint gathers	
Residual statics computation	4.0–8.0 s window
Cross-dip correction	Cross-dip correction
Trim statics	4.0–12.0 s window ± 12 ms
Stack	Nominally 192 fold
Trace energy balance	8.0–16.0 s window
Combine stacks	Redatum to 750 m (2a and 2b) 1200 m (3)
Migration	Phase shift
Coherency filter	
Plot	1:1 at 6000 m/s

[20] Reflection characteristics common to most of the data include the following: (1) the crust is reflective in most areas; (2) the base of crustal reflectivity (the reflection Moho, when it occurs at a travel time that is appropriate for the refraction Moho) is commonly visible, although it is not always outlined by a distinct reflection; and (3) few reflections from the mantle are observed. However, a significant mantle reflection occurs in the west end of Line 3, where it dips eastward between 14.0 and 21.0 s (~45 to 73 km assuming 6.0 km/s for the crust and 8.0 km/s for the mantle).

5. Structure of the Crust

5.1. Western Canada Sedimentary Basin

5.1.1. Data Description

[21] On line 2b, near-surface layers thicken westward from ~1.5 s at Fort Nelson to ~3.0 s near station 21000 (Foldout 1). In general these are subparallel and appear little deformed until about station 21500 where they delineate an anticline between 0.0 and 2.0 s.

[22] At greater travel times, reflections outline large and complex structures to 8.0–10.0 s. In the east, a prominent west dipping reflection is visible between 8.0 and 10.0 s beneath Fort Nelson. Two broad antiforms crest near stations 21000 and 22800 (A and B on Foldout 1b). The west limb of B is situated beneath about stations 20500–20000 between ~4.0–8.0 s (~12–24 km) and can be followed to about station 20000 where it is truncated by overlying west dipping layers at ~8.0 s (Foldout 1).

[23] Reflections fade downward near 12.0–14.0 s between about stations 22800 and 22200. If this is near the Moho [*Welford et al.*, 2001], then the reflection Moho is more diffuse and has a later arrival time here than it does along strike to either the south [*Cook and van der Velden*, 1993] or to the north [*Cook et al.*, 1999].

5.1.2. Interpretation

[24] The near-surface layers that thicken from 1.5 to 3.0 s are known from drill holes to be Phanerozoic strata of the Western Canada Sedimentary Basin. They are underlain by subparallel layers between 1.5 and 3.0 s (near Fort Nelson) and 3.0–4.0 s (near station 21000) that could include either lower Paleozoic strata, Late Proterozoic strata, or both. Possibilities for the origin and age of the layers and structures at greater travel times are: (1) they are Proterozoic strata of the Fort Simpson basin (with or without intruded sills) that were deformed (e.g., west side of Figure 4 from *Cook et al.* [1999]); (2) they are deformed crystalline "basement" rocks (>1.85 Ga) of the Wopmay Orogen that underlie the Fort Simpson basin strata (e.g., antiforms A1 and A2 in Figure 4 of *Cook et al.* [1999]), or (3) they include both Proterozoic strata and pre-1.85 Ga basement. Three lines of reasoning favor the last of these interpretations.

[25] First, projection along strike from Line 1 can be made using isostatic gravity (Figure 2). The Fort Simpson ramp occurs between about stations 1000 and 2000 on Line 1 [*Cook et al.*, 1999] where there is a prominent transition from high to low gravity values (Figure 2). This gravity transition projects southwestward to near stations 20000 and 21000 on Line 2b where a similar thick (~8.0–10.0 s) layered sequence dips westward (Foldout 1b). Second, drill holes that penetrate into rocks that underlie the known Paleozoic strata in all cases except one intersected sedimentary strata. One drill hole intersected igneous rock that may be a sill [*Ross et al.*, 2000]. Third, shallow layers of the prominent west dipping sequence are correlated with layers projecting to exposures of Proterozoic strata.

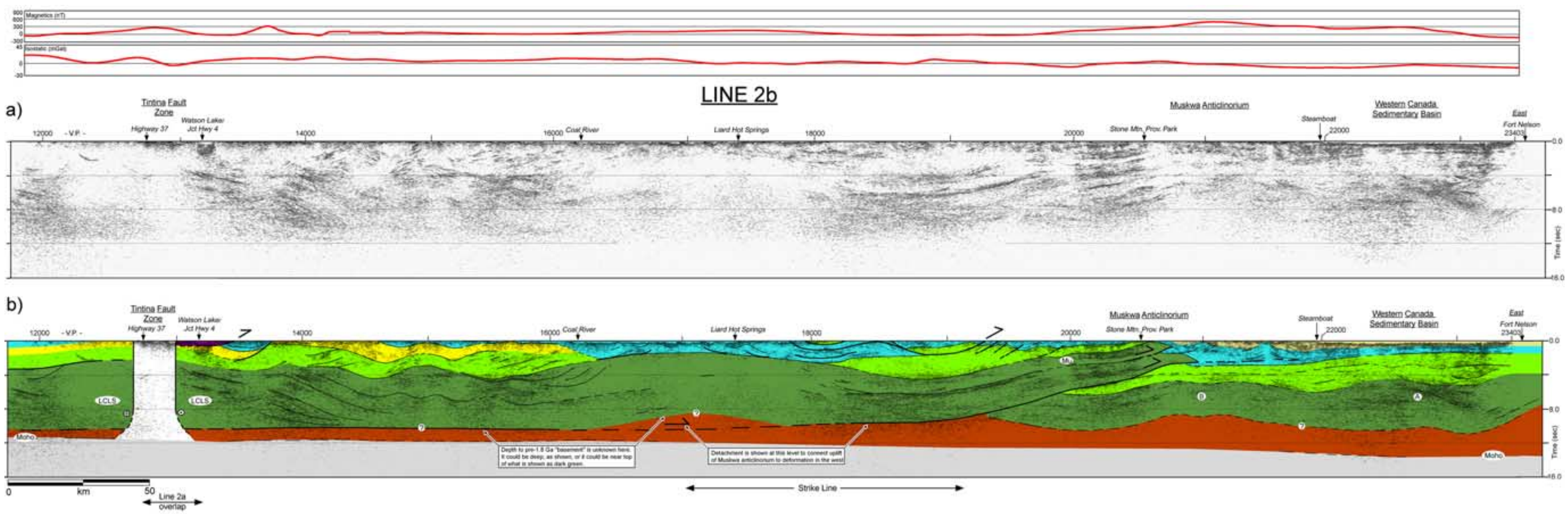
5.2. Foreland Belt to Tintina Fault

5.2.1. Data Description: Line 2b

[26] Prominent reflections dip westward from exposures of Proterozoic strata in the vicinity of Stone Mountain Provincial Park (Mu in Foldout 1b and Figure 4). The apparent dip of the reflections decreases as the profile trends northward toward strike. Nevertheless, the zone of reflections can be followed between ~5.0–9.0 s continuously from there to at least station 18200, where the data quality deteriorates.

[27] From stations 18200 to 17000, the profile generally follows the strike of the surface structures. A broad, near-surface antiformal warp crests near station 17400 and overlies deeper reflections with angular discordance (Foldout 1a). Whether this is a true anticline or whether it is an apparent structure due to the oblique orientation of the line to strike is not known. Structures above the apparent anticline are complex with short wavelengths and appear to flatten into the top of the antiform.

[28] From stations 17000 to ~15500, reflections commonly dip westward from the surface to ~4.0 s, where they appear to flatten (Foldout 1a). At longer travel times, they are generally subhorizontal.



Foldout 1. (a) Migrated seismic reflection data with isostatic and gravity profiles above. Gravity and magnetic data are from *Geological Survey of Canada* [1999a, 1999b]. (b) Interpretation of Line 2b. Thickness of the Proterozoic strata is determined by outcrop (>5 km), prominent reflection layering (~20 km), low seismic refraction velocities (~20 km), and exposures of Proterozoic strata elsewhere in the northern Cordillera. Colors follow the legend of Figure 1. Mu, Muskwa complex; LCLS, lower crustal layered sequence; A and B, see discussion in text. See enlarged version of this figure in the HTML.

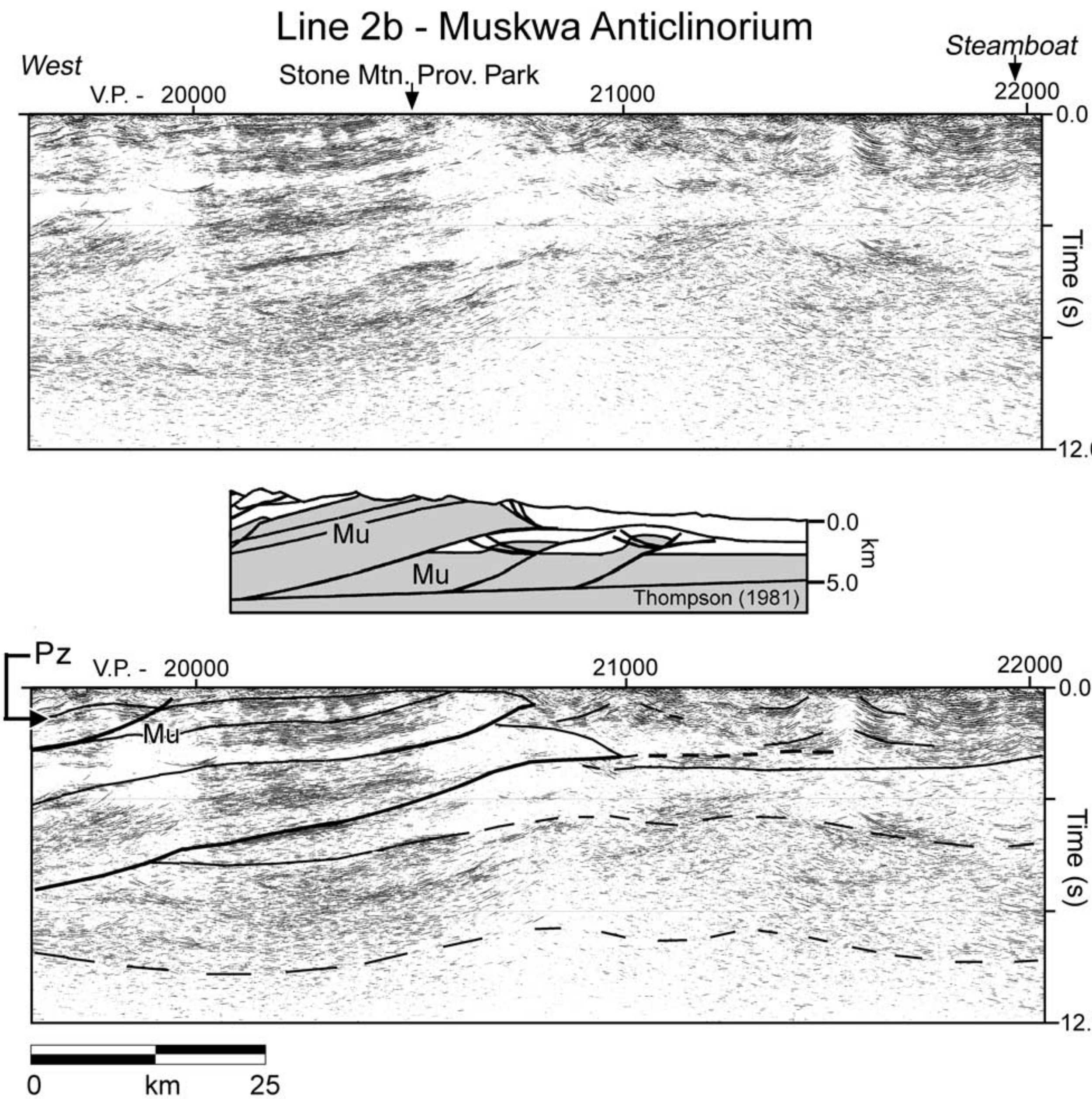
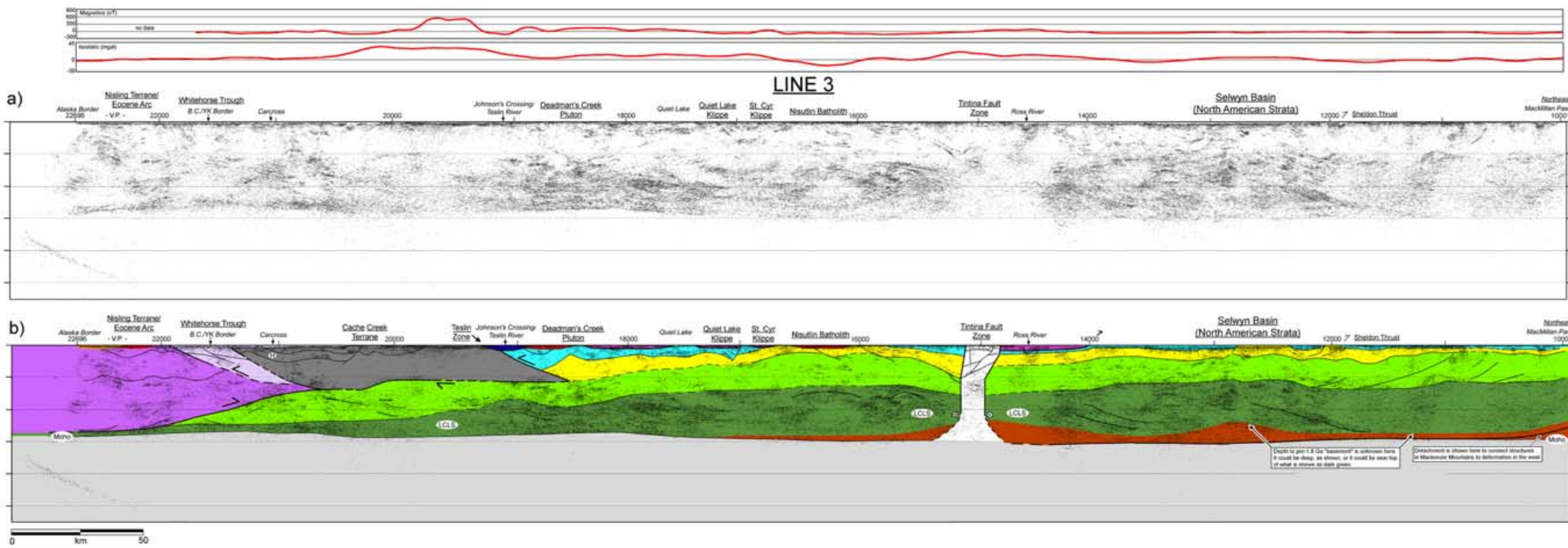


Figure 4. Enlargement of data across the Muskwa anticlinorium (upper), interpretation (lower) and comparison with cross section by *Thompson* [1981] (middle). Mu, Muskwa strata; Pz, Paleozoic-Proterozoic unconformity. The LCLS layering that underlies much of the Cordillera ties to surface at the Muskwa anticlinorium, providing strong evidence for Precambrian layering as the origin of reflectivity within LCLS. A footwall counterpart to the uplift shows that the Precambrian layers may extend to at least 10 s (~30 km). This part of the profile crosses the Tuchodi anticline on the east side of the Muskwa anticlinorium.

[29] Between station 15500 and Watson Lake, both east and west dipping reflections occur near the surface and a dominantly east dipping fabric of strong reflectivity and some complexity is seen below 4.0 s north of about station 14500. Throughout most of this region, reflections in the middle and lower crust appear to flatten above 11.0–12.0 s. The reflection Moho is indistinct, but is characterized by a general decrease in reflectivity below 11.0–12.0 s.

5.2.2. Data Description: Line 3

[30] Line 3 is nearly perpendicular to the strike of surface structures ~350 km to the northwest of Line 2b (Figures 1a and 1c). It crosses most of the Selwyn Basin, a region of westward thickening Neoproterozoic and Lower Paleozoic basinal strata that were penetratively deformed and subsequently intruded by Cretaceous plutons [*Gordey and Makepeace*, 1999].



Foldout 2. (a) Data and (b) interpretation of Line 3. Low seismic refraction velocities to 20–25 km [*Creaser*, 2000] are consistent with thick Neoproterozoic and Mesoproterozoic strata. Projection of surface structures from the east coupled with the reflection geometry permit regional faults such as the Plateau fault to be located at 4.0 s (~12 km) or deeper. H, Horsefeed formation(?). Colors and abbreviations as per legend. See enlarged version of this figure in the HTML.

[31] East of the Tintina fault, Line 3 exhibits similar gross reflection characteristics as Line 2b (Foldout 2a). For example, from the eastern end of the line (station 10000) at Macmillan Pass to about station 12000, reflections dip gently westward from the surface and flatten near 4.0 s. At longer travel times, reflectivity is diffuse and subhorizontal, with the base (reflection Moho) near 12.0 s (~36 km). From about station 12000 to the Tintina fault, reflections in the upper crust are more complex and are underlain by prominent layered reflections, many of which below 5.0 s dip eastward and flatten above 12.0 s. (Foldout 2a).

5.2.3. Interpretation

[32] The ability to relate near-surface reflections to known surface structures in several places provides limits on the range of possible interpretations. The upper layers that project into the subsurface west from Stone Mountain Provincial Park are Proterozoic strata (Mu in Figure 4). Preliminary correlations with a synthetic seismic trace calculated from the stratigraphic data permit correlation of the layering to ~2.0 s near station 20000 with Precambrian strata that are exposed adjacent to the line (S. Siegel and F. Cook, personal communication, 2003). Deeper layers that are parallel to Mu are also likely to be strata because (1) they are obviously layered, (2) they are parallel to known strata above, and (3) they correlate along strike to layers within the Fort Simpson basin on Line 1 that appear to exhibit sedimentary features such as growth faults and unconformities [*Cook et al.*, 1999]. The causes of the reflectivity at depth are uncertain; they may be contacts between sedimentary and intrusive rocks (e.g., sills), contacts between sedimentary rocks of different lithology, or both.

[33] The layering is visible to at least 10.0 s along both lines 2b and 3. If these are largely Proterozoic strata, then either (1) the continental basement beneath these layers was remarkably thin, (2) the basement beneath them could be oceanic crust, or (3) the basement beneath them is continental and the present Moho is not the same as the Proterozoic Moho. In any case, the crust today consists of 25–30 km of stratified rock. This interpretation is similar to that proposed for the west end of Line 1 ~240 km to the north [*Cook et al.*, 1999] as well as for reflection profiles across the Purcell anticlinorium in southern British Columbia [*Cook and van der Velden*, 1995].

[34] The total thickness of post-1.85 Ga Proterozoic strata may exceed 20 km. Collinear seismic refraction data along both Lines 1 and 2b have thick zones with low velocity that coincide with the reflections to at least 20 km depth beneath the Western Canada Sedimentary Basin [*Welford et al.*, 2001; *Fernandez-Viejo and Clowes*, 2003] (Figure 3). Conversely, a region of high refraction velocities spatially coincides with the uplifted reflections Mu near Stone Mountain Provincial Park (Figure 3). Furthermore, the refraction interpretation includes localized regions of high seismic velocities in the lower crust that appear to correspond to deep structural highs observed on the reflection profile (e.g., below refraction shot point 4, between shot points 4 and 5, and east of shot point 8; Figure 3). Taken together, these correlations suggest that (1) the layered reflectors, at least some of which are known to be Proterozoic strata, correspond to velocities of ~6.5 km/s or less, and (2) less reflective zones beneath these stratified rocks correspond to velocities greater than ~6.5 km/s (Figure 3). To a first approximation, therefore, the ~6.5 km/s contour may be near the top of older basement beneath the strata. The change from low refraction velocities in the upper and middle crust east of refraction shot point 4 to intermediate velocities between shot points 4 and 5 can thus be interpreted as a juxtaposition of deep, intermediate velocity layers on the west with lower velocity rocks on the east along a west dipping fault (Figure 3).

[35] Constraints available from outcrop relationships and regional considerations collectively suggest an age of 1.8–0.8 Ga for the rocks near the surface based on the following rationale. First, Proterozoic strata are overlain with angular unconformity by Paleozoic strata [*Taylor and Stott*, 1973]. There is an angular discordance visible on Line 2b dipping westward from near the surface at about station 20500 to ~1.0 s near station 20000, that is reasonably interpreted as this unconformity (Figures 4 and 5; Foldout 1). Second, the Proterozoic strata are older than 779 Ma dikes that cut them [*LeCheminant and Heaman*, 1994; *Ross et al.*, 2001]. Third, detrital zircons within the lower formations of the Muskwa assemblage (Tuchodi, Tetsa and Chischa formations) have recently been interpreted to indicate these strata are part of regional Sequence A (1.84 Ga > t > 1.1 Ga) of *Young et al.* [1979]; more specifically, the lower formations are younger than ~1.766 Ga [*Ross et al.*, 2001]. The age of the overlying Proterozoic rocks is uncertain, but it is greater than 779 Ma.

[36] The paucity of reflections between stations 18000 and 17000 on Line 2b precludes detailed correlation. The loss of signal appears to be a result of poor transmission of energy through thick surface gravel. The first arriving waves, which propagate along the surface and are usually the most prominent on the entire data set, were practically nonexistent. However, as much of this portion of the line is oriented along the strike of surface rocks, regional seismic patterns near station 19025, on the south, can be correlated directly with those near station 17000, to the north (Figure 1a). Thus reflections Mu of the Muskwa anticlinorium correlate in a general sense with similar reflections in the middle and deep crust east of Watson Lake, Yukon.

[37] Between station 16000 and Watson Lake on Line 2b, crustal layering is complexly deformed, with divergent dips and prominent reflectivity throughout. Several lines of evidence lead to the conclusion that the layering includes primarily Proterozoic strata and their basement. First, reflections from the Muskwa layering can be followed into the middle and lower crust as discussed above. Second, although exposures are sparse throughout this region [*Gabrielse*, 1963a], upper crustal layers can be generally correlated with surface outcrop and structures. The mapped units between about stations 15000 and 14000 are predominantly Neoproterozoic sedimentary strata (Hyland Group [*Gabrielse*, 1963a]) that were deposited on older North American rocks. Third, near station 14300, the upper crustal layers are arched in an antiform whose west limb appears to flatten near 3.0 s and overlies deeper, prominent

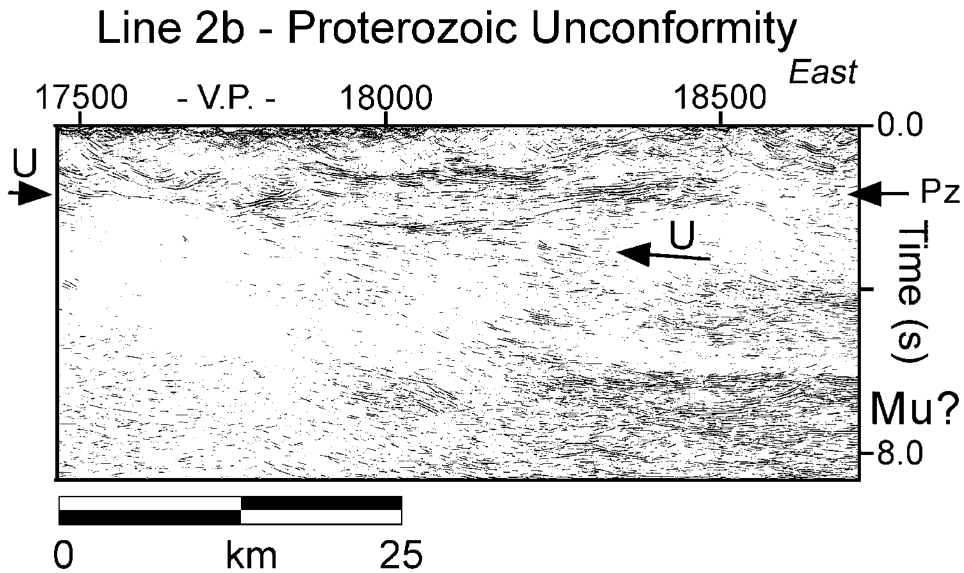


Figure 5. Enlargement of data on the west side of the Muskwa (Mu) anticlinorium. There appear to be two unconformities. One is likely the base of Paleozoic strata, and the deeper one (labeled here as 'U') is probably the unconformity within the Muskwa (Mu) layers [*Taylor and Stott*, 1973].

east dipping layering that is listric into the lower crust (Foldout 1a). The geometry of upper crustal layers (which correlate with ancient North American strata on the surface) overlying deeper layers with angular discordance is consistent with the deep layers representing older North American strata and their basement that project from the east.

[38] The prominent east dipping layers that flatten in the lower crust are situated east of accreted terranes; a ~20–50 km wide part of the Yukon-Tanana terrane is exposed in a thin thrust slice east of Watson Lake and Ross River (Figure 1a; Foldouts 1 and 2). The east dipping reflectors in the middle and lower crust may be North American Mesoproterozoic strata (e.g., Muskwa, Mackenzie Mountains Supergroup) and basement, or, alternatively, they may be rocks associated with the Yukon-Tanana terrane that were tectonically wedged into the lower crust. The latter interpretation is not preferred because reflections that are correlated with Neoproterozoic and Paleozoic rocks near the surface overlie the east dipping layers and because North American strata are present on the surface far to the west. We interpret the deep layers as North American rocks that were deformed in the Proterozoic.

[39] A key observation from these data is that the Paleozoic and upper Proterozoic strata are generally sparsely reflective, with only a few prominent reflections outlining the major structural geometry (Foldout 2a). Reflections from the upper crust typically flatten into layered reflections below, at ~4.0 s (e.g., near stations 11000 and 13000 along Line 3). This result implies that there is a regionally consistent pattern of weaker reflectivity in Paleozoic and Late Proterozoic strata, and more prominent reflectivity in older Proterozoic strata and their (?) basement.

[40] Near the eastern end of Line 3, reflections dip slightly westward from near the surface to ~4.0–4.5 s where they flatten into the deeper subhorizontal layering (stations 10500–11000, Foldout 2a). Even though these layers project to within ~0.5 s (~1.0–1.5 km) of the surface, the Lower Paleozoic rocks here are gently folded with little indication of a correlative structure within them. These west dipping layers may be within older (e.g., Proterozoic) strata that flatten at depth into yet older strata or basement. The thickness of subhorizontal rocks above the dipping layers (~0.5 s) is too small to accommodate normal Hyland Group strata [e.g., *Gabrielse and Campbell*, 1991]. Accordingly, these dipping layers either (1) are older than Hyland Group and represent a topographic high that was beveled prior to Early Paleozoic; (2) include Hyland Group and were beveled before Early Paleozoic time; or (3) were truncated by a shallow (<1.0 s, or <3 km depth) detachment during formation of the Cordillera.

[41] Farther west along Line 3, between stations 12000 and 13000, Neoproterozoic strata are exposed [*Gordey and Makepeace*, 1999] and they are correlated with dipping reflections. Near stations 13000–14000, however, Paleozoic strata are paraconformable or only slightly discordant with the Neoproterozoic strata. The angular discordance observed near station 10600 is thus likely due either to truncation of older Proterozoic layers that were topographically high during the Neoproterozoic (option shown on Foldout 2), or to post-Windermere deformation.

[42] The correlation of shallow (to ~4.0 s) layers with Paleozoic and Late Proterozoic strata along both Lines 2b and 3, coupled with interpretation of deep layers as Mesoproterozoic strata exposed in the Muskwa anticlinorium, leads to the conclusion that the crust east of the Tintina fault is composed primarily of these strata and their basement.

The deep reflection zone is hereafter referred to as the "lower crustal layered sequence," or LCLS.

[43] Approximately 100 km northeast of Line 3, the Plateau thrust fault juxtaposes Mesoproterozoic Mackenzie Mountains Supergroup strata in its hanging wall with middle Devonian strata in the footwall (Figure 1c) and may have ~30 km of horizontal offset [*Cecile and Cook*, 1981; *Cook*, 1991]. According to *Cecile and Cook* [1981], the Plateau thrust projects westward to a depth of ~10 km or more beneath the eastern end of Line 3; the seismic data are consistent with this interpretation if the fault is located at or below 4.0 s (~12 km).

[44] Significant shortening occurs within Lower Paleozoic shale and chert of the Selwyn Basin [*Cecile*, 1984; *Gordey and Thompson*, 1991]. The short wavelength of the deformation and the fact that the stratigraphic level remains approximately constant across this portion of the basin indicate that there must be a shallow detachment within Lower Paleozoic or Upper Proterozoic strata. Thus, there may have been two or more detachment levels during formation of the Cordillera in Mesozoic-Tertiary times: a shallow (<3 km) detachment to accommodate short-wavelength near-surface shortening, and a deeper detachment (>12 km) that is linked with regional structures farther east and outcrops as the Plateau fault (Foldout 2b [see also *Cecile and Cook*, 1981]).

[45] The LCLS is deformed along both Line 2b and Line 3 east of the Tintina fault. The age of deformation is uncertain; it is likely to have been at least partly Precambrian as Precambrian structures have been documented by *Taylor and Stott* [1973] and *Thorkelson et al.* [1998, 2001]. Near stations 11000–13000 along Line 3 (Foldout 2a) and near stations 13000–15000 along Line 2b (Foldout 1a), 4.0–5.0 s (~12–15 km) of structural relief are visible in the LCLS.

[46] Accreted rocks of the Yukon-Tanana terrane that exist east of the Tintina fault appear to be very thin (<2–4 km; Foldouts 1 and 2 [*Tempelman-Kluit*, 1979; *Wheeler and McFeely*, 1991]). Subhorizontal reflections are visible between 0.5–1.5 s near stations 14100–14400 on Line 3. A shallow, west dipping reflection that may be the base of the Yukon-Tanana terrane is visible along the northeastern portion of Line 2a that is parallel to strike (stations 13880–13000; Foldout 3b).

5.3. Tintina Fault Zone

5.3.1. Data Description

[47] Lines 2a, 2b and 3 cross the Tintina-Northern Rocky Mountain Trench fault system. Along Line 3, its surface expression is a ~25 km wide morphologic trench (Figure 6). Along Line 2b, however, it is not exposed and its position in the subsurface was estimated by projection from the north and south. The seismic response is similar along both profiles: layered reflections terminate against a near-vertical 15–20 km wide zone of weak and discontinuous reflectivity (Figure 6). The poorly reflective areas in the vicinity of the Tintina fault zone are not likely due to reduction of data quality because the first arrivals on the field records are visible and prominent. On both profiles the crustal reflectivity exhibits the same gross pattern on both sides of the Tintina fault zone, with the upper 4.0–5.0 s being less reflective than the lower 5.0–11.0 s (Foldouts 1, 2, and 3; Figures 6 and 7).

5.3.2. Interpretation

[48] The observation of prominent reflections that terminate at a poorly reflective zone on Lines 2a, 2b and 3 leads to the conclusion that this is the most likely position of the Tintina fault zone in the subsurface (Figure 6). The lack of coherent reflections from this zone is consistent with a steeply dipping fault or fault zone that penetrates the crust through the LCLS to at least 11.0 s (~33 km depth). The amount of fault displacement cannot be determined from the seismic reflection data. Along both profiles 2 and 3, the LCLS is visible on both sides of the Tintina fault zone. This implies that the LCLS is present in the subsurface from southeast of Watson Lake along strike to Ross River on both the east and west sides of the Tintina fault zone and is evidence of regionally extensive layers.

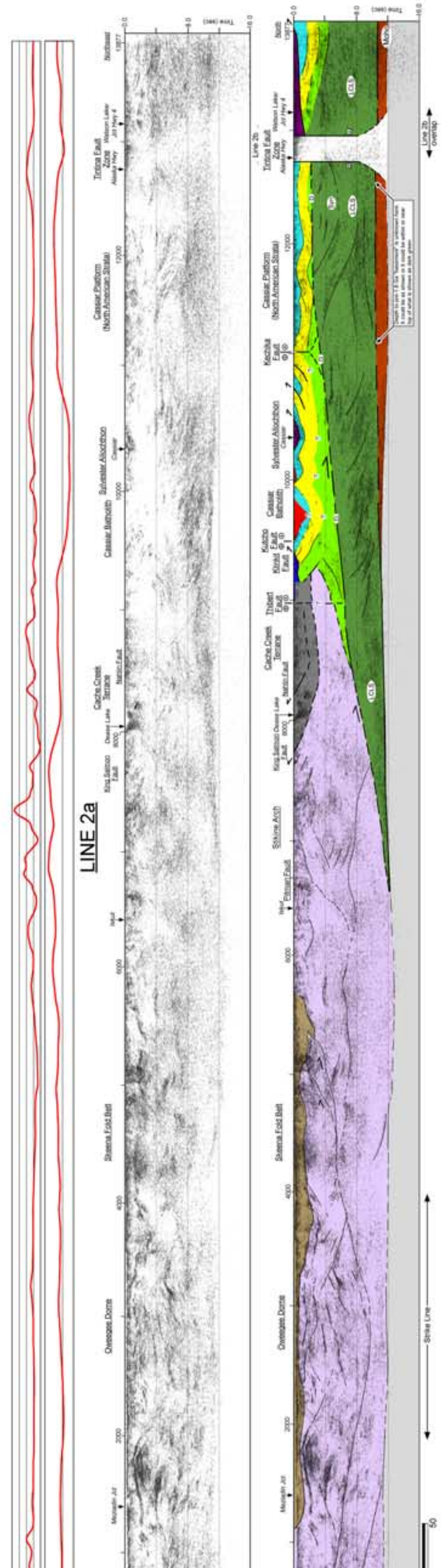
5.4. Tintina Fault Zone to Teslin–Thibert Zone

5.4.1. Data Description: Line 2a

[49] The LCLS on Line 2a is visible on both sides of the Tintina fault zone (Foldout 3 and Figure 6). To the west, the LCLS tapers in the lower crust and disappears at a position ~60 km south of Dease Lake (near station 6800), a distance of ~300 km along the profile from the Tintina fault zone, or ~150 km across strike from the Tintina fault zone.

[50] A major structural feature of the crust within the LCLS between the Tintina fault zone and the Kechika fault is the synform between 3.0 and 8.0 s at about stations 11100–13000 (Syn in Figure 8 and Foldout 3). The top of the synform is bounded by a narrow zone of horizontal to slightly southwest dipping reflections and/or truncations at ~2.0 s. This zone (D in Figure 8 and Foldout 3) appears to correlate westward with discontinuous reflections and reflection truncations that for the most part delineate the top of LCLS as it tapers westward in the lower crust to about station 6800. Between stations 10500 and 8900, the LCLS is deformed into a series of east dipping structures that generally do not project above D (Foldout 3). Reflections above D in this region are dominantly west dipping and in some areas are even listric into D (e.g., near station 9100 at 5.0–6.0 s; Figure 9 and Foldout 3). In contrast, reflections below D appear to dip primarily east and are commonly listric into the lower crust or Moho. Accordingly, zone D may be an important structural and/or stratigraphic boundary.

[51] Visible reflections are generally consistent with the surface geology and thus provide links to important stratigraphy and structures. For example, between the Tintina fault zone and approximately station 11100, the profile crosses deformed Paleozoic and Neoproterozoic strata that were deposited on or adjacent to North American crust and subsequently deformed into west dipping structures and isoclinal folds [see *Gabrielse*, 1963b; *Gabrielse and Yorath*, 1991, Figure 17.1f]. The reflections also dip generally westward from the surface to ~2.0 s. However, Neoproterozoic strata are also deformed into recumbent isoclines in areas to the south that may project to depth in this region.



Foldout 3. (a) Data and (b) interpretation of Line 2a. The depth to pre-1.85 Ga "basement" is not known, so is speculatively shown to thin westward beneath the westward tapering LCLS. See enlarged version of this figure in the HTML.

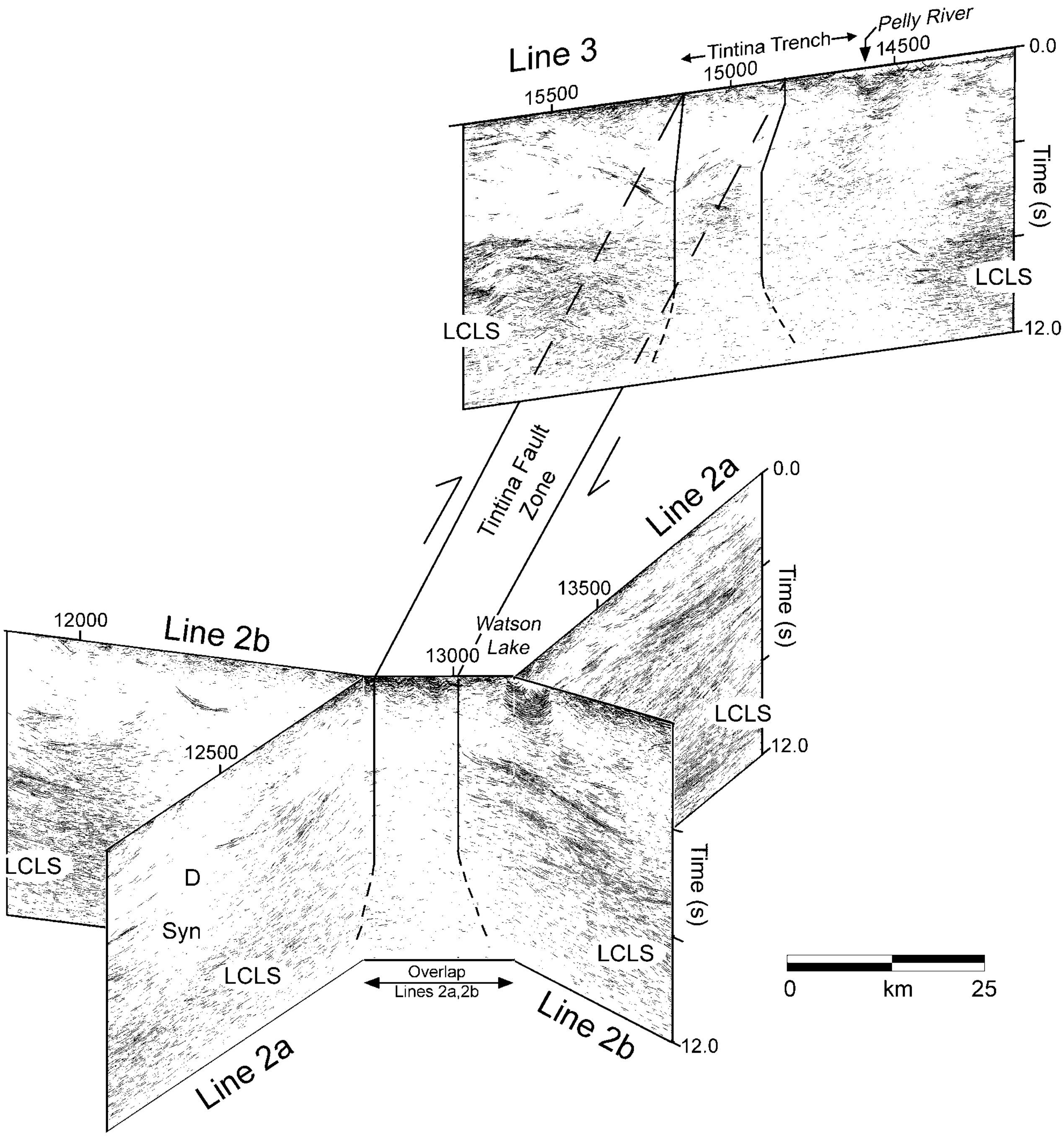


Figure 6. Enlargement of data in the vicinity of the Tintina fault zone, plotted in a perspective view. Note that the Tintina is a zone of poor reflectivity throughout the crust on all lines. D and Syn same as Foldout 3b. Note that LCLS is present on both sides of the Tintina fault zone, implying that this zone is continuous both along and across strike. The distance between lines 2 and 3 is not to scale.

[52] The Kechika fault, a north-northwest striking dextral strike-slip fault that was active as late as Oligocene [*Gabrielse*, 1985] and may be kinematically linked to the northern Rocky Mountain Trench fault [*Evenchick*, 1988], projects northward to the vicinity of stations 11100–11300. Although the deep reflections of the LCLS fade slightly between stations 11200 and 10800, most reflections are nearly continuous across this region and do not exhibit any obvious characteristics of disruption by a steep fault zone (Foldout 3 and Figure 8). Some shallow reflections have a relatively shallow dip to the northeast between the surface trace of the fault and the area of structure "Syn."

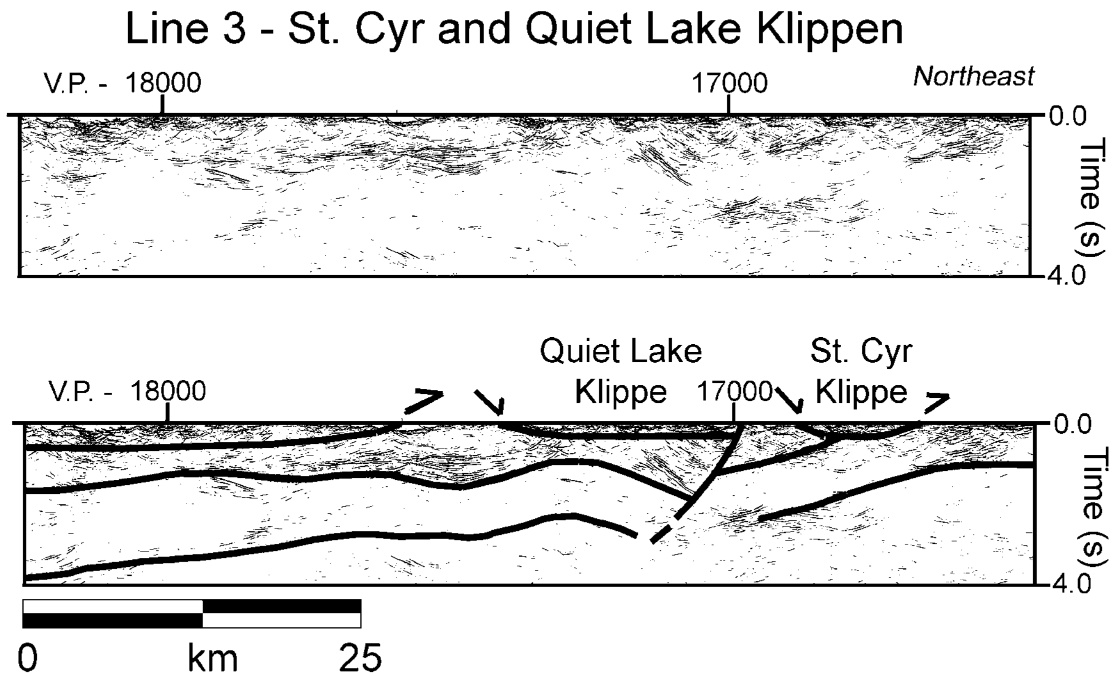


Figure 7. Enlargement of data along Line 3 in the vicinity of Quiet Lake, illustrating the thin klippen and the relationship with deeper layers of probable Paleozoic and Proterozoic strata. The prominent layering below the klippen are likely Paleozoic strata that were deformed prior to emplacement of the klippen. The normal fault is a late Mesozoic or early Tertiary structure that offsets the two fragments of the originally continuous allochthons.

[53] Near station 10600, rocks of the Slide Mountain terrane (Yukon-Tanana terrane) were thrust northeastward over the North American strata [*Harms et al.*, 1988]. Reflections near the surface between stations 10200 and 10600 define a synform that correlates with the syncline cored by the Sylvester allochthon. Rocks defining the synform correlate with rocks of the Sylvester allochthon between about stations 10300 and 10500 and with pericratonic rocks of North America northeast of about station 10600 and southwest of about station 10300 (Figure 9 and Foldout 3).

[54] Between stations 9800 and 9450, the profile crosses the mid-Cretaceous Cassiar batholith, a late orogenic granitic pluton with mixed isotopic characteristics [*Driver et al.*, 1998]. On its east side, near station 9800, the batholith has an intrusive contact with Cassiar Platform strata. On its west side, near station 9400, it is separated from Cassiar platform strata by the Kutcho-Cassiar fault [*Wheeler and McFeely*, 1991; *Gabrielse*, 1998]. From about station 9800 to 9600, the data exhibit little reflectivity from the surface to >5.0 s (Foldout 3). However, near the southwestern margin of the Cassiar batholith, in the vicinity of the Kutcho-Cassiar fault, northeast dipping reflections rise from ~3.0 s to near the surface (Figure 9). These northeast dipping reflections are underlain by southwest dipping reflections that are listric into surface D. At travel times greater than 4–5 s, the west dipping zone D and the westward tapering LCLS with its prominent northeast dipping reflectivity dominate the reflection patterns (Foldout 3). Reflection zone D can be followed westward to at least station 8200.

[55] From the Tintina fault zone to the disappearance of the LCLS along line 2a, the base of the crustal reflectivity (reflection Moho) occurs between ~11.0 s and 12.0 s (Foldout 3a). In some regions it is diffuse (e.g., between stations 12500 and 10800); in others it is more distinct (stations 10800 to 6800). It is commonly overlain by reflections that are listric into it or flatten slightly above it. The reflection Moho exhibits regional variations in travel time and reaches its shallowest level of ~11.0 s near station 9500, but deepens both to the north and to the south.

5.4.2. Data Description: Line 3

[56] As observed along Line 2a, the LCLS layering is visible to at least stations 19500–19800, where reflectivity deteriorates between 6.0 and 12.0 s. Reflections below 6.0 s west of this zone are likely a continuation of the LCLS, which appears to taper to a thin (~1.0 s) zone near the base of the crust at the end of the line (Foldout 2a). Crustal reflectivity from the Tintina fault zone to the Teslin zone is grossly similar to that east of the Tintina. For example, with the exception of several key reflections that correlate with Paleozoic and upper Proterozoic strata near the surface, the upper crust is relatively devoid of reflections and the middle and lower crust (~4.5–11.0 s) are dominated by layered reflections (LCLS).

[57] The travel time to the reflection Moho varies from ~12.0 s near station 16000 and 11.8 s near station 20000, to ~10.8 s near stations 18000 and 21000 (Foldout 2a). In

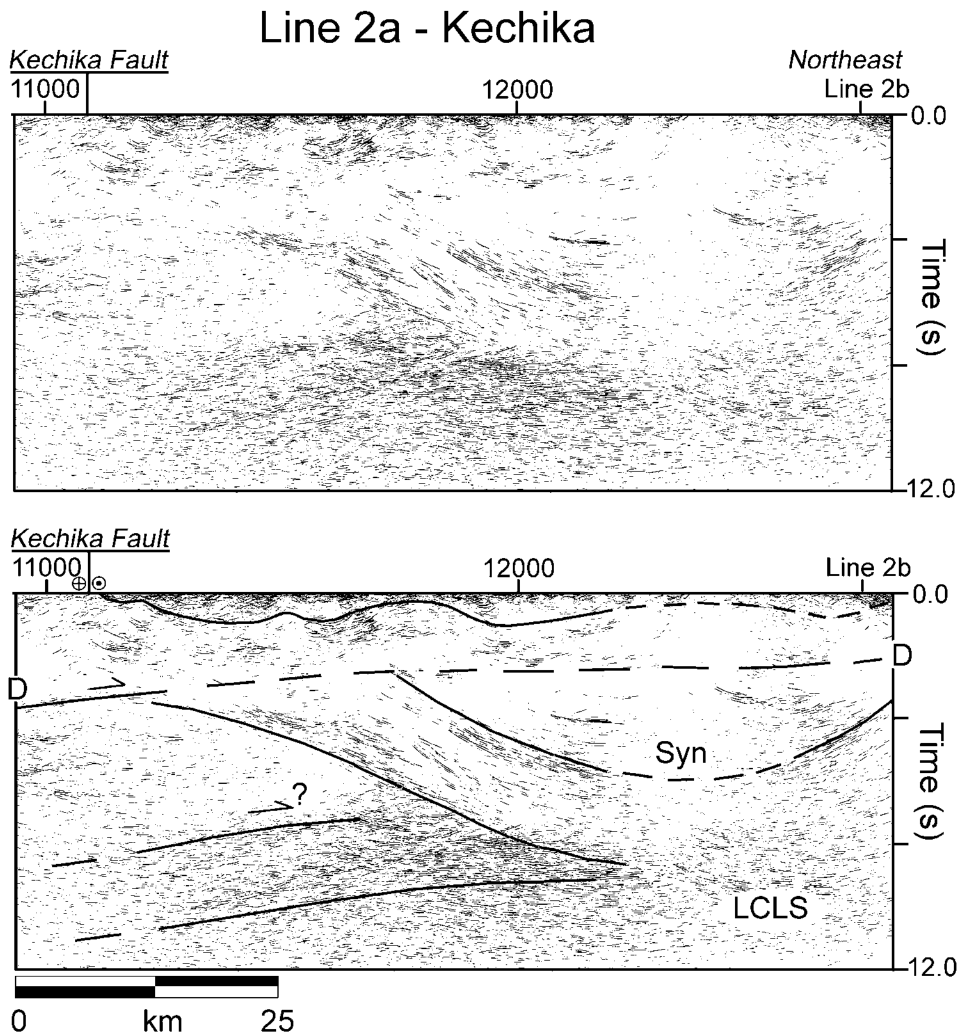


Figure 8. Enlargement of data in the vicinity of the Kechika fault near Line 2a. Note the large syncline (Syn) that has more than 10 km of relief and is truncated by zone D. There is no expression of this structure at the surface, leading to its interpretation as pre-Cordilleran, and probably Precambrian. D and LCLS same as Foldout 3b. There is no expression of this structure at the surface, nor is there a known hanging wall counterpart exposed to the east. Therefore structure Syn, and by inference the prominent lower crustal reflections that underlie the Cordillera (LCLS), are interpreted to be pre-Cordilleran and possibly Precambrian.

some areas the Moho is diffuse, with no clear reflection (e.g., stations 15000–17200), whereas in others it is distinct (e.g., near stations 17300–19200 and 20200 westward). Reflections above it are commonly subparallel to the boundary, although several structures are listric into the reflection Moho or slightly above it (e.g., near stations 21000–22000; Foldout 2a).

[58] Near-surface reflections between the Tintina fault and Teslin fault zones provide important links to outcrop. For example, between stations 15200 and 16000, subhorizontal to shallow east dipping reflections between 0.0–4.0 s correlate on the west with Early Paleozoic and Neoproterozoic rocks of the Cassiar Platform (Foldout 2). Similarly, between stations 16600 and 18500, the profile crosses the St. Cyr and Quiet Lake klippen of the Yukon-Tanana terrane, both of which overlie Cassiar Platform strata [*Fallas et al.*, 1998]. The Cassiar strata appear to be characterized by dipping reflections (Figure 7).

5.4.3. Interpretation

[59] The reflection geometry and the ability to correlate some features with surface structures leads to the interpretation that supracrustal rocks which were deposited on or adjacent to the ancient North America continental margin persist at depth far to the west of the Tintina fault zone. As stated previously, the LCLS probably represents autochthonous or parautochthonous Proterozoic strata and their basement. The continuity of this zone of reflections to a location ~60 km south of Dease Lake requires that North American crust and underlying lithosphere project to at least that point. A consequence of this interpretation is that all of

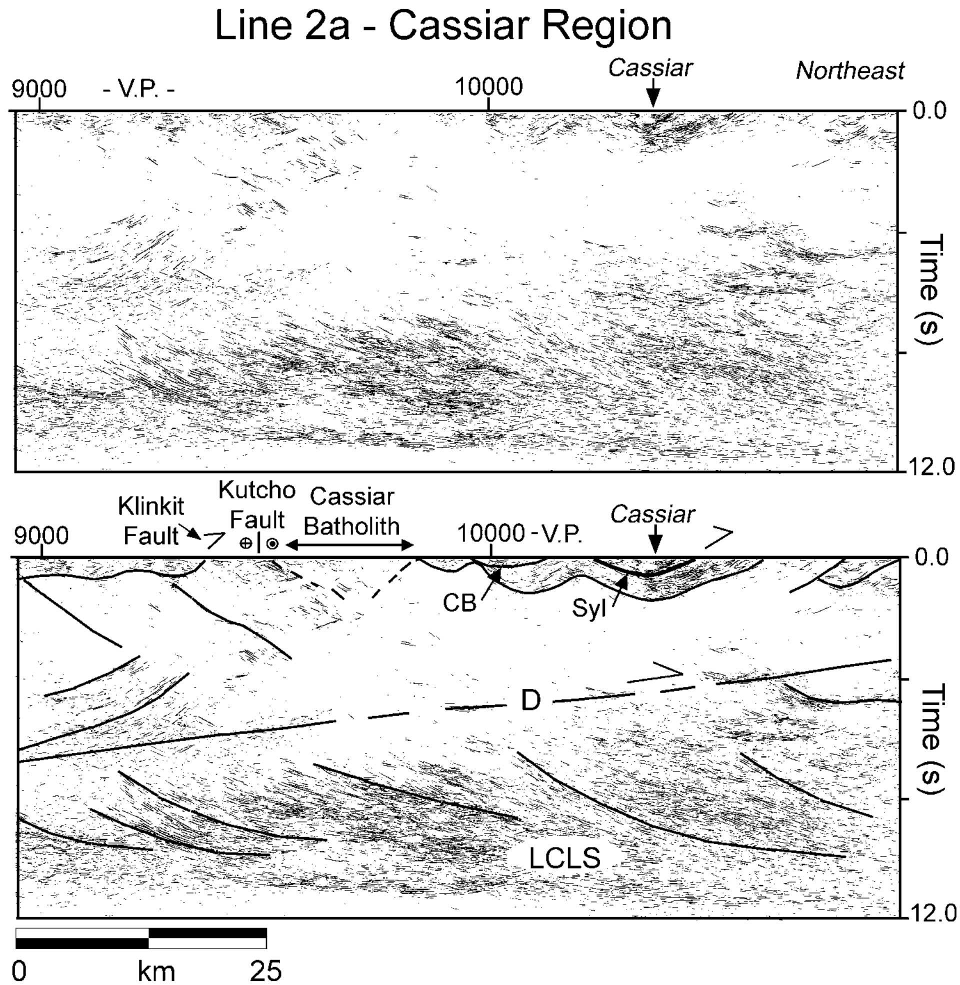


Figure 9. Enlargement of data in the vicinity of the Sylvester allochthon near Cassiar, British Columbia. Note the dish shape of the allochthon and underlying strata. Note also that moderately northeast dipping layers underlie the Kutcho fault, a near-vertical strike-slip fault on the surface. There is no obvious expression of a vertical fault on the data, although it may be detached at D, or it may flatten into D. CB, a thin "tongue" of the Cassiar batholith; Syl, Sylvester allochthon.

the rocks above the LCLS were transported onto the ancient margin. Although the exact position of the base of the accreted klippen is uncertain, most of the layering that is visible to ~2.0 s in this region is likely due to underlying strata (e.g., Figure 7).

[60] Vertical faults are difficult to delineate on crustal-scale reflection data. They are most easily identified when signals such as diffractions (before migration) and disruptions of layered reflections (after migration) are observed. The Tintina fault zone correlates with an obvious disruption of the LCLS, but other faults (Teslin, Kechika, Kutcho-Cassiar, and possibly the Thibert) do not. The youngest faults, whose latest movement post-dated regional contraction and would thus be expected to disrupt all older layers, are the Tintina and Kechika [*Gabrielse*, 1985]. There are several possible explanations for why the Kechika fault does not disrupt LCLS, including: (1) it loses displacement northward [*Gabrielse*, 1985] and may not have significant offset at the position of the profile; (2) it offsets layers parallel to strike; (3) it becomes gently dipping to the east and is parallel to subhorizontal reflections; or (4) layering observed beneath the Kechika is younger than the latest (Oligocene) motion on the fault (e.g., late intrusions). This last possibility is considered to be the least likely because layers observed beneath the projected position of the Kechika fault can be correlated with older structures and post-Oligocene intrusives are rare.

[61] The LCLS was apparently deformed prior to formation of zone D because it is truncated at D (e.g., Syn; Foldout 3b). The change in structural orientation at D is interpreted as a truncation surface, either structural or stratigraphic. Whatever the cause, the results are local listric reflections above it and truncations of dipping reflections below. Two possible structural interpretations are (1) the deformation of the LCLS was Precambrian and predated the development of the Paleozoic miogeocline and the formation of the Cordillera, or (2) the deformation within the LCLS was Cordilleran. In the first case there are abundant

surface data (outcrop, map pattern [*Gabrielse*, 1967; *Eisbacher*, 1981; *Thorkelson et al.*, 1998, 2001]) and subsurface data (seismic, potential field [*Cook*, 1988a, 1988b; *Cook and Maclean*, 1995; *Cook*, 2000]) that provide evidence of widespread Proterozoic (~1.8–0.8 Ga) orogenic activity throughout the northern Cordillera. In the second case, the LCLS structures would have formed during an early stage of Cordilleran development (e.g., in the late Paleozoic or early Mesozoic [*Gordey et al.*, 1987]) and then were truncated along D at a later time.

5.5. Teslin–Thibert Zone to Coast Mountains

5.5.1. Data Description: Line 2a

[62] The dextral Thibert fault marks the contact between Quesnellia on the northeast and the Late Devonian to Triassic Cache Creek terrane that accreted during the middle Jurassic [*Gabrielse et al.*, 1991; *Gabrielse*, 1998]. Near the Nahlin fault (Figure 1b), reflections dip predominantly northeastward to ~2.0 s and then flatten northward. Between the Nahlin fault and the King Salmon fault at about station 7700, the near surface reflectivity is disrupted by a data gap around the community of Dease Lake. However, prominent north dipping reflections project from the King Salmon fault to ~2.0 s (Foldout 3a). There is no indication that they project to greater depth, as there are discordant west dipping reflections at ~3.0–4.0 s. It is possible that these reflections correlate with subhorizontal reflections between 2.5–3.0 s beneath the Cache Creek terrane, but this is uncertain. At longer travel times, the reflection geometry is more complex, at least down to surface D at ~7.0 s. Below zone D, the LCLS tapers and disappears near station 6800 (Foldout 3a).

[63] South of Dease Lake, Line 2a crosses Late Triassic-Jurassic imbricated metasedimentary and metavolcanic rocks of northern Stikinia in the footwall of the King Salmon fault [*Monger et al.*, 1991; *Gabrielse*, 1998]. From station 5700–1200, Stikinia is overlain stratigraphically by Jurassic to Cretaceous sedimentary rocks of the Bowser Lake group, and both display contractional structures associated with the Cretaceous Skeena fold belt [*Evenchick*, 1991a, 1991b]. From station 1200 to its southern termination, the line crosses Stikinia [*Monger et al.*, 1991; *Gabrielse et al.*, 1991].

[64] South of the King Salmon fault, sedimentary and volcanic strata of northern Stikinia dip northward and are visible on this profile as a homocline of north dipping reflections to at least 1.5 s, or ~4.5 km depth (Foldout 3). South of this homocline, the upper 2.0 s (~6 km) are strongly reflective near stations 7600–7400. South of there, Devonian-Permian arc rocks are exposed in a region where reflections appear gently folded. The arc strata are bounded by steep faults [*Evenchick*, 1991c; *Ash et al.*, 1997], however, making extrapolation of reflections to the deep crust problematic. The problem is compounded by the fact that Stikinia was not a "layer cake" prior to Mesozoic deformation and much of it is covered by younger successions.

[65] At about station 5700 the line crosses the angular unconformity between faulted Triassic and Jurassic strata of Stikinia and the Bowser Lake group [*Evenchick*, 1991a, 1991b]. The line remains in Bowser Lake group and the Skeena fold belt to about station 1200, in two locations (5100–4100 and 3300–2800) approaching and then diverging from the contact with underlying Stikinia. In the southern location, gently north dipping reflections in the upper 0.5–1.0 s appear to correspond with the gentle north plunge of the culmination of Stikinia. In the north the line is at a high angle to local structural trends of the Skeena fold belt, but from VP 5100–4100 and 3300–1500, the line is subparallel with Skeena fold belt trends. Reflections in the upper 2.5 s are in the Bowser Lake group, but do not clearly exhibit the magnitude of shortening documented in the fold belt [*Evenchick*, 1991a, 1991b]. This is probably due to the scale of structures as well as the strike-parallel orientation of parts of the line. Deep reflections define large zones of apparent north and south dip, and many truncations can be recognized.

[66] The westernmost 40–50 km of the profile are dominated by a northeast dipping fabric outlined by a prominent zone of discordance (C) that dips northeast from ~4.0 s (~12 km) near Stewart (station 159) to ~8.0 s near station 2200. It flattens as the line turns to the northwest and may correlate with a northeast dipping reflection between stations 4100 and 4800 near the reflection Moho. Reflections above C exhibit a variety of dips, but flatten into it or are truncated at it (Foldout 3 and Figure 10). Nearly all reflections beneath C dip northeastward more steeply than C, and appear to be truncated at or merge with C. The crustal fabric exhibits a fundamental change in orientation from variably north and south dipping, northeast of station 5000, to consistently northeast dipping southwest of station 5000. The Coast shear zone, a near vertical structure which records multiple phases of thrusting followed by extension, is located along the updip projection of C about 70–80 km to the west of the profile [*Diebold et al.*, 1998; see also *Morozov et al.*, 1999; *Hammer et al.*, 2000].

[67] The reflection Moho travel time along the southern portion of Line 2a varies from ~11.0 s near station 9200 to ~13.0 s near station 5700. Its deepest point appears to be located near the convergence of the southwest and northeast dipping crustal fabrics (Foldout 3), although the Moho is diffuse and difficult to identify in this region.

5.5.2. Data Description: Line 3

[68] The Teslin zone, an enigmatic linear zone of high strain, is located near station 19100, where the line crosses the Teslin River. The crust is generally non-reflective to ~5.0 s, but is highly reflective below 5.0 s as the LCLS continues westward. Between about stations 19500 and 19800, the LCLS is less obvious due to a poor data area, but it is again visible west of station 19800 (Foldout 2a). West of that point, the relatively simple appearance of the LCLS as a thick layered sequence of subhorizontal reflections changes to a westward thinning zone of greater complexity between ~4.0 and 9.0 s. Above this zone, northeast dipping reflections appear to project from ~1.0 s at station 18900 to ~4.5 s near station 18500.

[69] Between the Teslin zone and about station 20000, reflectivity is weak from 0.0 to ~4.0 s. From station 20000

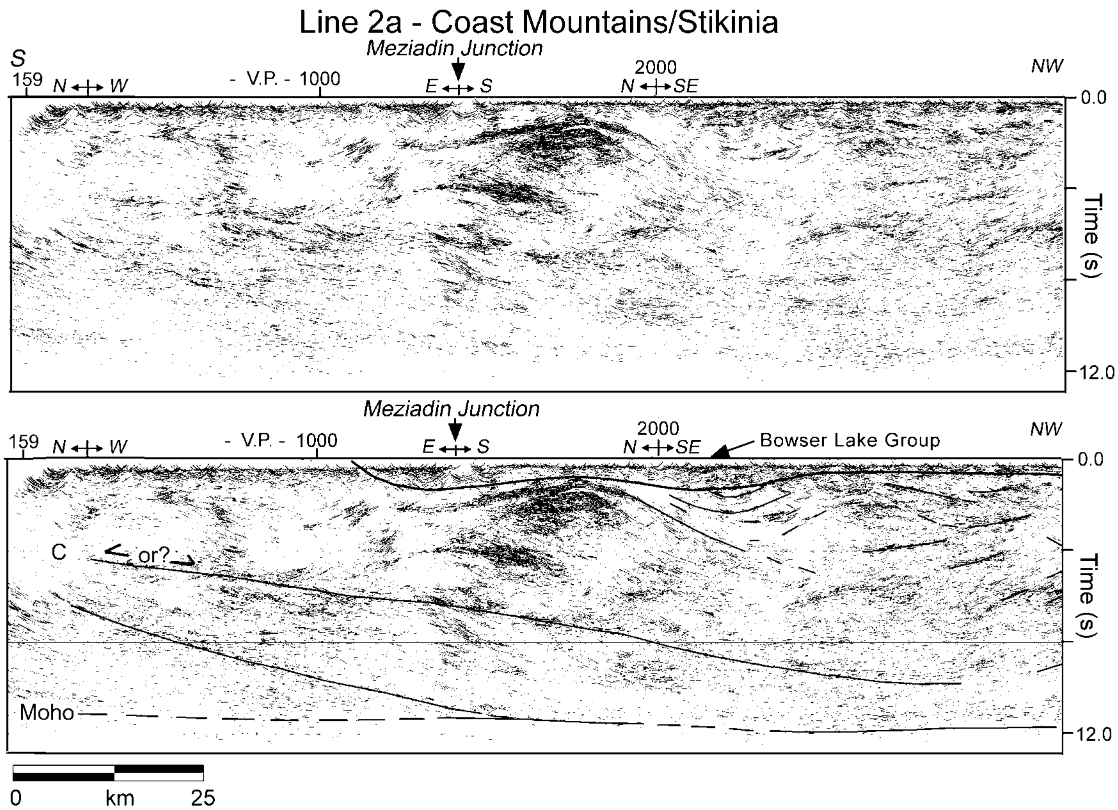


Figure 10. Enlargement of data along the western portion of Line 2a illustrating the thin surface layering of the Bowser Lake Group and complex structures within Stikinia rocks at depth. Zone C dips northeastward from ~4.0 s to near the Moho and appears to truncate underlying reflections. Prominent reflections within Stikinia outline large structures, but the origin of these is uncertain because they are unconformably overlain by Bowser strata. Zone C may be related to an accretionary complex, or it may be related to Miocene extension. Stewart, BC is located at station 159.

to about station 21000, subhorizontal layering that is visible from 0.0–2.0 s is deformed into broad, open folds with wavelengths of 20–25 km (Foldout 2a). Deep crustal reflections west of the Teslin zone appear to be the continuation of the LCLS that tapers to less than 1.0 s, at 10.5–11.5 s near the Alaska border (station 22696). Reflections between 6.0 and 9.0 s dip primarily westward and flatten into layering near the reflection Moho.

[70] Line 3 shows the only obvious mantle reflection from the entire data set. A distinctive reflection dips eastward between 14.0 and 21.0 s (~45–73 km depth assuming 6.0 km/s average velocity in the crust and 8.0 km/s average velocity in the upper mantle; Figure 11). It underlies a region of Eocene magmatism that was likely related to subduction (Figure 11 [*Morris and Creaser*, 1999]).

5.5.3. Interpretation

[71] The most striking feature of both Lines 2a and 3 is the westward tapering wedge shape of the LCLS. The layers are interpreted as the toe of a westward thinning prism of ancient North American rocks extending to the Alaska border on Line 3, and to ~60 km southwest of Dease Lake on Line 2a (Foldouts 2b and 3b). The exact nature of the rocks within these layers is uncertain. They may be Proterozoic (Fort Simpson) metasedimentary rocks, arc rocks of the Wopmay orogen, or a combination of both for the following reasons. First, the LCLS layers are zones of reflections that can be followed from outcrops of Mesoproterozoic strata in the Foreland belt to their tapered edge. Second, surface strata above the LCLS are known from geological mapping to be Neoproterozoic and Paleozoic along both Lines 2a and 3 to the east and northeast of the Teslin-Thibert zone.

[72] The deep structure of Stikinia and Skeena fold belt is characterized by severe deformation throughout the crust. Further analysis is required to delineate the relative ages of different structures, their magnitudes and their styles of deformation. Three large-scale structures appear to reach the surface. These are near the Stikine arch, the Oweegee dome, and the culmination at the south end of the profile (Foldout 3). In these regions the reflection layering is consistent with mapped Paleozoic and Mesozoic supracrustal strata of Stikinia, although the total thickness of these rocks is unknown. In short, interpretations of reflections beneath Stikinia are difficult because Stikinia was complexly deformed prior to the Mesozoic. Reflections could be from stratigraphic, structural or intrusive features and the orientation of the seismic profile at various angles to strike means that the observed dips are mostly apparent.

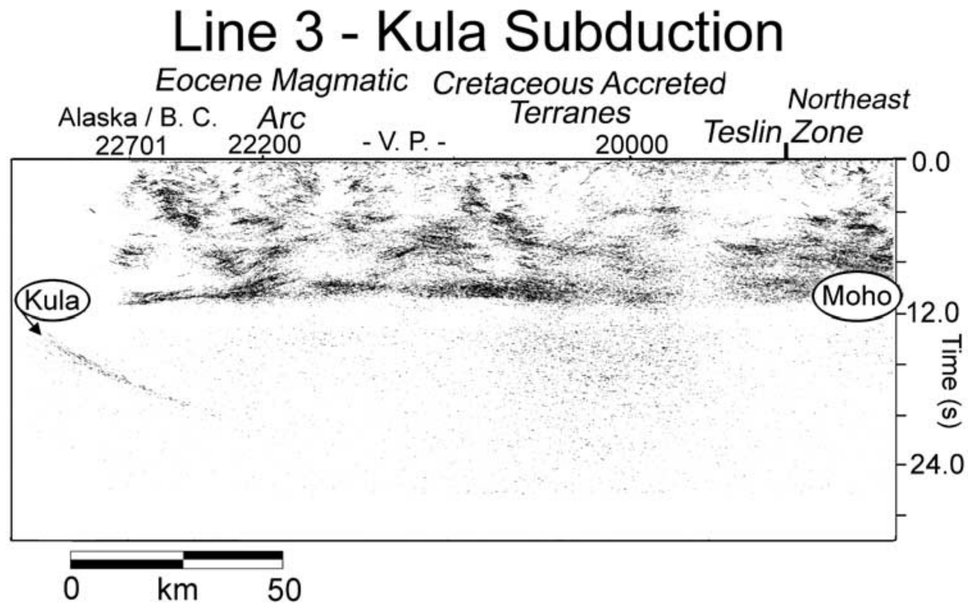


Figure 11. Enlargement of data along the western end of Line 3. A reflection zone dips from ~14.0 s (~45 km) eastward to ~20.0 s (~73 km) and is interpreted as a relict portion of the Kula plate. There does not appear to be any obvious enhancement or degradation of either Moho or crustal reflectivity due to late volcanism.

[73] Zone C may be associated with either northeast dipping crustal fabric seen in the eastern Coast Mountains and western Skeena fold belt, which may be related to Cretaceous lithospheric convergence [*Evenchick*, 1991a, 1991b; see Figure 12a from *Evenchick*, 1991a], or with Miocene lithospheric extension [*Heah*, 1990; *Evenchick et al.*, 1999; *Klepeis and Crawford*, 1999; *Chardon et al.*, 1999].

[74] Along Line 3, the crust above the LCLS exhibits arcuate structures and discontinuous reflectivity. There are no obvious effects of Eocene magmatism on the LCLS or overlying structures; hence the crust along this portion of the line is considered to be mostly accreted material of Cache Creek and Nisling terranes (Foldout 2b). Between 0.0 and 2.0 s west of station 20000, the folded layered reflections (H on Foldout 2b) project to the surface west of Carcross. These may be rocks of the Upper Paleozoic Horsefeed formation, the underlying Nakina Group volcanics, or both [*Monger*, 1977a]. Rocks that underlie these supracrustal strata are not observed at the surface. The crust west of the Whitehorse Trough is part of the Nisling terrane that includes rocks of largely continental affinity [e.g., *Gehrels et al.*, 1990; *Mihalynuk*, 1999]. Structures within the Nisling terrane include broad folds above the LCLS that appear to be truncated or to terminate near the Whitehorse Trough and the adjacent Cache Creek terrane.

6. Discussion

6.1. Chemical Signatures of Young Igneous Rocks

[75] Variations in the chemical composition of late or postorogenic igneous rocks across the northern Cordillera, both isotopic and major element, indicate distinct differences in compositions. In general, late granitic intrusions are more evolved in the east and more primitive in the west [e.g., *Morris and Creaser*, 1999]. The transition occurs near the Teslin zone, where plutons exhibit a mixed signature. An interpretation in the context of the results from the reflection patterns is that the evolved granites in the east (e.g., in the Selwyn Basin) were derived from, or contaminated by, metasedimentary and/or basement rocks of the LCLS. Conversely, intrusions that formed west of the thickest part of the LCLS exhibit more primitive signatures. The Cassiar batholith, one of the largest in the northern Cordillera, exhibits isotopic signatures that indicate mixing of juvenile and evolved crustal sources [*Driver et al.*, 1998]. It is located in an intermediate position above the LCLS.

[76] Volcanic rocks derived from the mantle provide evidence of variations in mantle composition across the Cordillera. *Francis et al.* [1999] summarized results of isotopic variations of Cenozoic basaltic rocks and interpreted changes across the Tintina fault and the Cassiar/Teslin zones to indicate changes in the mantle over distances of tens of kilometers. The change across the Tintina fault is consistent with the seismic characteristics if the Tintina fault offsets the upper mantle, as the disruption in the LCLS suggests. In contrast, the Cassiar-Kutcho fault and Teslin zone do not appear to disrupt the LCLS. In order to explain the isotopic variations across the Cassiar-Kutcho fault, either the mantle undergoes a change, possibly as a result of the westward thinning of the LCLS, or the eastward thickening of the LCLS influenced the isotopic characteristics.

[77] Compositional characteristics of Eocene magmatic rocks near the western end of Line 3 indicate that these rocks were derived from a mantle source, most likely due to

subduction [*Morris and Creaser*, 1998]. The observation of east dipping reflections in the mantle between ~45 and 73 km depth (Figure 11) is consistent with this interpretation, and it is likely an image of the relict Kula plate. Reflections near the Moho have no obvious indications of disruption by passage of magma from the overlying mantle wedge into the crust.

6.2. Structure and Timing of Deformation

[78] One of the most powerful tools for interpreting the origin of deep crustal reflections is the tracing of reflections from exposed surface features to depth. In some cases, this approach is successful for following boundaries such as faults (e.g., Wind River thrust [*Smithson et al.*, 1979]); fault zones (e.g., Grenville front [*Green et al.*, 1988]); subduction zones [*Clowes et al.*, 1987]; or layered strata (e.g., Purcell anticlinorium [*Cook and van der Velden*, 1995]). However, while individual faults or structures can sometimes be projected into the lower crust, it is rarely possible to trace a coherent zone of reflections through a region as large as the northern Cordillera. Yet the LCLS persists across and along strike for hundreds of kilometers. If its interpretation as Mesoproterozoic strata and basement is correct, then a large part of the northern Canadian Cordillera is underlain by crust and lithosphere that extend west from shield and platform areas of the North American craton.

[79] In the absence of direct ties to outcrop or drill holes in the western part of the Cordillera, alternative interpretations for LCLS are possible. For example, it could be a series of layered igneous intrusives that were deformed during Cordilleran orogenesis, or it could be a series of ductile flow structures that were formed as the Mackenzie Mountains were being uplifted. In some respects, both of these have merit as partial explanations for LCLS.

[80] In the first case, the large amount of magmatism that is visible on the surface throughout the region is likely indicative of similar rocks at depth. There is no evidence from outcrop data throughout the region, even where more than 15–20 km of Proterozoic strata are exposed in the Wernecke Mountains northwest of Line 3 [*Delaney*, 1981; *Thorkelson et al.*, 1998], that massive regional igneous "underplating" was an important process. Furthermore, there is no evidence from regional refraction profiles for high (>7 km/s) lower crustal velocities in the vicinity of LCLS [*Creaser*, 2000; *Creaser and Spence*, 2000; *Hammer and Clowes*, 1999; *Welford et al.*, 2001].

[81] Regional ductile flow in the hinterlands of some orogens has been proposed as a mechanism for deforming lower crustal rocks in wide "channels" [e.g., *Beaumont et al.*, 2001]. While deformation at depth was ductile, it is not known whether ductile deformation zones were relatively narrow and acted as detachments within and around LCLS, or whether the observed layers are completely transposed from their original configuration. In the first case, while the deformation (and flow) was ductile, the observed reflectivity originated as the supracrustal strata and basement. In the second case, the observed layering may not have retained its original characteristics. The ability to correlate some of the layers on the east with known stratigraphy, coupled with the observation that 15–20 km of Mesoproterozoic strata are exposed only 150 km along strike to the northwest of Line 3 provide evidence that much of the layering east of the Tintina fault is due to supracrustal rocks. West of the Tintina, large structures within the layering rise to within 4–6 km of the surface where the surface rocks are essentially unmetamorphosed (e.g., Syn in Foldout 3b). Accordingly, it appears that ductile flow in these layers, if present, would likely have been confined to discrete zones.

[82] All of the accreted rocks east of the western limit of the LCLS are restricted to the upper crust above the taper and are thus detached from their lithospheres. Although such detachment has been proposed previously for the Yukon-Tanana terrane [*Tempelman-Kluit*, 1979; *Erdmer*, 1987; *Erdmer et al.*, 1998; *Fallas et al.*, 1998], the Dorsey assemblage [*Harms and Stevens*, 1996; *Nelson et al.*, 1998], and the Sylvester allochthon [*Harms et al.*, 1988; *Gabrielse et al.*, 1991], it has not been known for the Quesnel, Cache Creek, Nisling, and eastern Stikinia terranes. If the LCLS layering indeed underlies all of Line 3, for example, and if the interpretation of the LCLS as part of the ancient North American plate is valid, then all of the terranes along this profile are tectonic flakes.

[83] The regional configuration of the LCLS constrains the structural geometry, relative ages, and tectonic processes by which the northern Cordillera has evolved. Foremost among these is that Cordilleran structures observed at the surface are, with few exceptions, confined to the upper crust above the LCLS. Exceptions may include structures such as the Muskwa anticlinorium, which may have been uplifted partly during the Proterozoic and partly during the Late Cretaceous–Paleocene. The detachment level for this uplift must have been deep and thus may underlie the LCLS. Similar processes and structures may be responsible for other uplifted Proterozoic rocks near the front of the northern Cordillera [e.g., *Mitchelmore and Cook*, 1994], but this remains to be tested.

[84] Other than the Tintina–Northern Rocky Mountain Trench fault system, major faults (e.g., Kechika, Kutcho, and Thibert) apparently do not penetrate deep into the crust along the profiles. For any proposed displacements on the Tintina fault exceeding 350–400 km, an implication of our interpretation is that the LCLS layers underlie most of the northern Cordillera from the central Yukon to north central British Columbia, a strike distance of up to 1000 km. Proterozoic (~1.85 Ga) orthogneiss is found in the Sifton Ranges west of the Northern Rocky Mountain Trench, ~100 km southwest of the Muskwa anticlinorium [*Evenchick et al.*, 1984; *Evenchick*, 1988], and Fort Simpson strata are not present at this location. It is uncertain how much displacement occurred on the Northern Rocky Mountain Trench fault system at this location as the displacement probably decreases southward, but restoration of 200 to 400 km of dextral offset are required place the Sifton Range gneiss adjacent to the southward projection of the ~1.85 Ga Fort Simpson trend (Figure 2).

[85] The presence of the LCLS beneath accreted rocks of the Yukon-Tanana, Slide Mountain, Quesnellia, and Cache

Creek terranes suggests that these are detached, thin flakes that were emplaced over North American strata from hundreds of kilometers to the west. For example, the eastern limit of the Yukon-Tanana terrane along Line 3 is ~20 km east of the Tintina fault. If the LCLS indeed projects to the lower crust near White Pass, then the leading edge of the Yukon-Tanana terrane must have been thrust a minimum of 300 km from the west, not counting shortening of underlying North American margin strata, which is considerable [e.g., *Gordey*, 2002]. In contrast, the eastern edge of Stikinia may have been a tectonic wedge that was driven into the western margin of ancient North America.

6.3. Comparison to Southern Canadian Cordillera

[86] The new cross sections are a second major transect of the Canadian Cordillera completed by Lithoprobe. Between 1984 and 1995, the southern Canadian Cordillera transect was undertaken between the Foreland belt on the east and the Cascadia convergent margin on the west. Although detailed comparisons are being addressed by ongoing research, some key elements relevant to the seismic reflection structure are the following. The pervasive LCLS observed in the north does not appear to have a counterpart in the southern Cordillera. There are several possible reasons for this, including (1) postcontractional extension in the south that is not observed in the north may have disrupted the layering if it had been present following accretion; and (2) the layering in the north is interpreted to have been initiated as a result of extension following the development of the Wopmay Orogen at ~1.8 Ga, whereas major crustal extension and basin formation in the south may not have begun until ~1.5 Ga when the Belt-Purcell basin formed.

[87] The thickness of the crust is more or less uniform (33–36 km) in the Omineca and Intermontane Belts in both the southern [*Cook*, 1995] and northern Cordillera. The reasons for this are not yet clear as the postcontraction lithospheric extension that is offered as an explanation for thin crust in the south is not visible in the north. Nevertheless, the depth of the Moho increases eastward by about 6–10 km in both regions. In the south the change occurs near the southern Rocky Mountain Trench on the west side of the Foreland belt [*Clowes et al.*, 1995; *Cook*, 1995], whereas in the north a similar change is visible nearly 150 km east of the Mackenzie Mountains [*Cook et al.*, 1999; *Fernandez-Viejo and Clowes*, 2003].

[88] Despite the fundamental differences in lithospheric evolution, accreted rocks in both regions are interpreted to be thin, detached from their lithospheres, and thrust over North American crust and upper mantle. As has been observed in many other parts of the world, detachment and large lateral translation of accreted rocks appears to be a fundamental process in orogens of many different ages.

6.4. Development of the Northern Canadian Cordillera

[89] Most of the northern Canadian Cordillera is underlain by ancient (Mesoproterozoic or older) crust and lithosphere that are continuous with lower crust and lithosphere of the Canadian Shield beneath the Western Canada Sedimentary Basin. Likely candidates for the LCLS reflection layering are Mesoproterozoic sedimentary layers (with or without sills) and/or Mesoproterozoic "basement." Accordingly, three stages in the tectonic development of the Northern Canadian Cordillera are illustrated in Foldout 4.

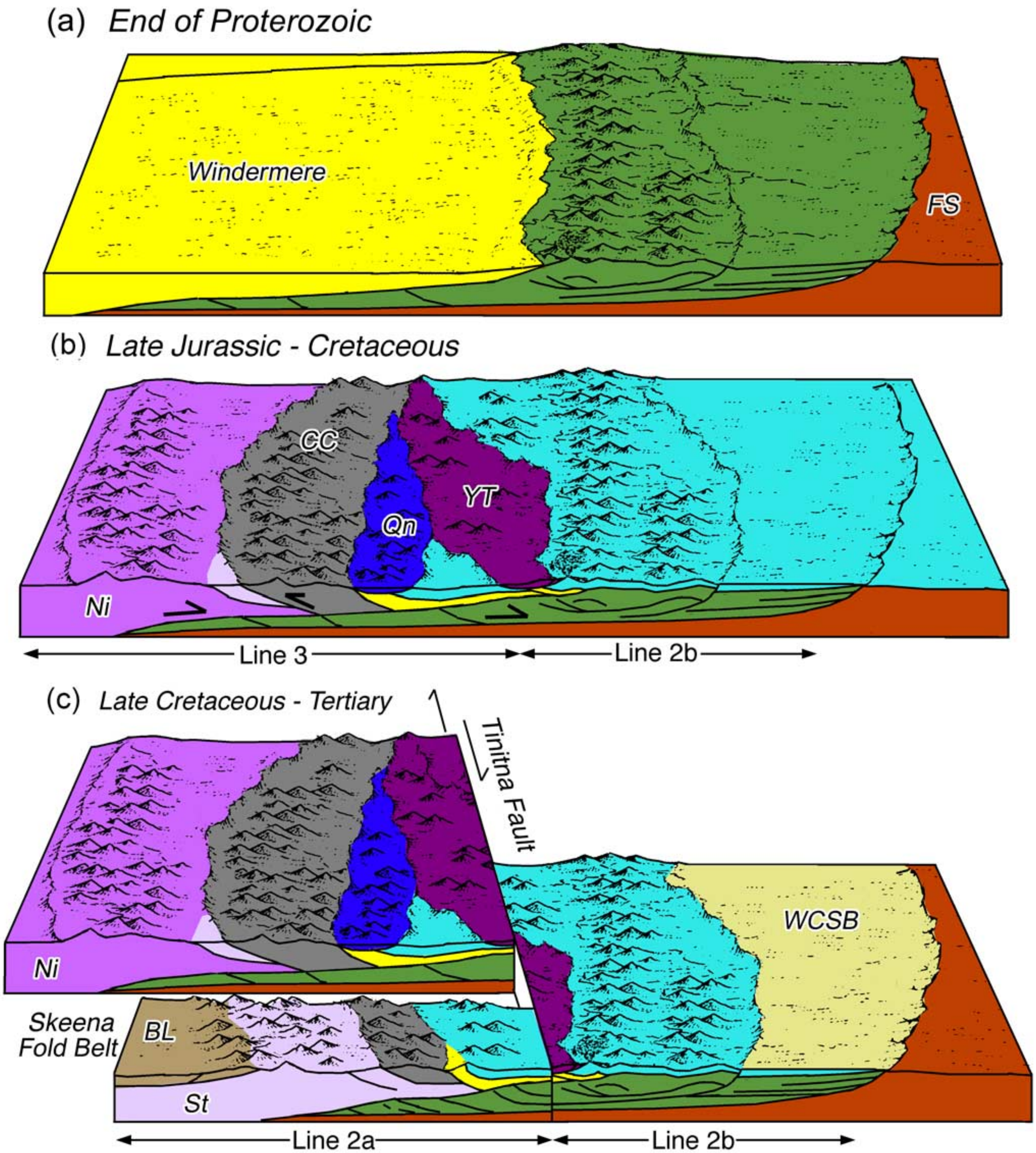
[90] At the end of the Proterozoic, the westward tapering wedge included Mesoproterozoic strata and their basement that were overlain by Neoproterozoic Windermere strata. The Windermere strata probably thicken westward as deeper Mesoproterozoic rocks thin, although the stratigraphic contact between them is not observed in the seismic profiles. Conversely, the Windermere strata thin eastward above the Mesoproterozoic layers. The deep Mesoproterozoic layers then taper eastward against the west facing Fort Simpson ramp (Foldout 4a).

[91] Accretion of terranes during the construction of the Cordillera resulted in a series of thin flakes emplaced upon the Proterozoic middle and lower crust east of Stikinia. Stikinia may comprise most of the crust of the western Cordillera. Its eastern tip was apparently wedged into rocks of LCLS and its western margin was overlain by late orogenic strata of the Skeena fold belt (Foldout 4b). The thin terranes and underlying Proterozoic and Paleozoic layers of LCLS were subsequently offset by the Tintina-Northern Rocky Mountain trench faults system (Foldout 4c).

7. Conclusions

[92] About 1895 km of deep reflection data along two profiles across the Cordillera in northwestern Canada provide images of layered rocks that can be followed westward from the orogenic foreland where they are exposed, for as much as 500 km to where they thin in the lower crust. They are at least 15–20 km thick and constitute a coherent sequence of layering that persists westward; the upper portions of them correlate with Proterozoic sedimentary rocks in the eastern part of the Cordillera. Whether the layering consists dominantly of Mesoproterozoic metasedimentary rocks, some combination of Mesoproterozoic metasedimentary rocks and their pre-1.8 Ga basement, or dominantly of pre-1.8 Ga basement is not certain at this time because no direct samples are yet available. Nevertheless, these deep layers are most easily interpreted as ancient North American rocks that project beneath exposed Neoproterozoic and Paleozoic rocks in the eastern part of the Cordillera and beneath crustal flakes of accreted terranes in the west. Only the Stikinia terrane may occupy the entire crust and possibly the upper mantle.

[93] The crust-mantle transition is visible as a downward loss of reflectivity along most of the profiles. It generally occurs between 11.0 and 12.0 s (~33–36 km), although from the Western Canada Sedimentary Basin to the Tintina fault, it is diffuse and is only a distinct reflection between Coal River and Watson Lake. West of the Tintina fault along Line 2a, the reflection Moho is more prominent, and outlines broad ~150 km wide arches. The reflection Moho appears to deepen westward on the south end of Line 2a. The only clear mantle reflection is observed at the western end of Line 3 where a suite of reflections dips east from



Foldout 4. Schematic diagram illustrating three stages in the development of the northern Canadian Cordillera and the relationship to the seismic structure. (a) End of Proterozoic time(Windermere) showing deformed undifferentiated Mesoproterozoic strata, the ∼1.85 Ga Fort Simpson ramp (FS) and westward thickening continental margin Windermere (Proterozoic-C) strata. (b) Jurassic-Cretaceous time after amalgamation of accreted terranes with North America and before major transcurrent faulting. The western portion of the section is part of Line 3 whereas the eastern portion is part of Line 2b. Accreted terranes are interpreted as thin flakes. (c) Configuration after dextral displacement of ∼300–400 km on Tintina fault. Here, the section along Line 2a is juxtaposed with Line 2b. WCSB, Western Canada Sedimentary Basin. See enlarged version of this figure in the HTML.

~14.0 s (~45 km depth) to ~21.0 s (~73 km depth) and is interpreted as a relict fragment of the Kula plate which was active in the region during the Eocene.

[94] **Acknowledgments.** The Natural Sciences and Engineering Research Council of Canada and the Geological Survey of Canada fund Lithoprobe. The data were recorded by Kinetex, Inc. of Calgary in 1999-2000 in challenging field conditions and were processed at the Lithoprobe Seismic Processing Facility at the University of Calgary. We wish to thank John Amor, Kate Bone, Phil Hammer, Jeff Schragge, Kim Welford, Gervais Perron, and Kris Vasudevan for assistance in the field work, Rolf Maier for assistance with the processing, and Elissa Lynn for assistance with some of the graphics. Early versions of the manuscript were reviewed by Bob Thompson, John McBride, Andy Calvert, and Craig Jones. Figure 2 was provided by Warner Miles of the Geological Survey of Canada. Lithoprobe contribution 1324. Geological Survey of Canada contribution 2002080.

References

Ash, C. H., R. W. J. Macdonald, and R. M. Friedman (1997), Stratigraphy of the Tatogga Lake area, northwestern British Columbia (104H/12 & 13, 104G9 & 16), in *Geological Fieldwork 1996*, edited by D. V. Lefebure, W. J. McMillan, and J. G. MacArthur, *Paper 1997-1*, pp. 283–290, Br. Columbia Minist. of Employment and Investment, Victoria, B. C., Canada.

Baragar, W. R. A., and J. A. Donaldson (1973), Coppermine and Dismal Lakes map area, *Pap. Geol. Surv. Can.*, *71-39*.

Beaumont, C., R. A. Jamieson, M. Nguyen, and B. Lee (2001), Himalayan tectonics explained by extrusion of a low-viscosity crustal channel coupled to focused surface denudation, *Nature*, *414*, 738–742.

Bell, J. S. (1973), Late Paleozoic orogeny in northern Yukon, in *Proceedings of the Symposium on the Geology of the Canadian Arctic*, pp. 23–28, Can. Soc. of Pet. Geol., Calgary, Alberta.

Bell, R. T. (1968), Proterozoic stratigraphy of northeastern British Columbia, *Pap. Geol. Surv. Can.*, *67–68*.

Cecile, M. P. (1984), Geology of the northwest Niddery map area, Yukon, *Geol. Surv. Can. Open File 1066*.

Cecile, M. P., and D. G. Cook (1981), Structural cross section, northern Selwyn and Mackenzie Mountains, *Geol. Surv. Can. Open File, 807*.

Chardon, D., C. L. Andronicos, and L. S. Hollister (1999), Large-scale transpressive shear zone patterns and displacements within magmatic arcs: The Coast Plutonic Complex, British Columbia, *Tectonics*, *18*, 278–292.

Clowes, R. M., M. T. Brandon, A. G. Green, C. J. Yorath, A. Sutherland-Brown, E. R. Kanasewich, and C. S. Spencer (1987), Lithoprobe-southern Vancouver Island: Cenozoic subduction complex imaged by deep seismic reflections, *Can. J. Earth Sci.*, *24*, 31–51.

Clowes, R. M., C. A. Zelt, J. R. Amor, and R. M. Ellis (1995), Lithospheric structure in the southern Canadian Cordillera from a network of seismic refraction lines, *Can. J. Earth Sci.*, *32*, 1485–1513.

Clowes, R. M., P. T. C. Hammer, and J. K. Welford (2001), Trans-Cordilleran comparison between SNORCLE seismic datasets: Near vertical to wide angle profiles of the lithosphere along corridor 2, in *Lithoprobe Slave-Northern Cordillera Lithospheric Evolution Transect and Cordilleran Tectonics Workshop*, edited by F. A. Cook and P. Erdmer, *Lithoprobe Rep. 79*, pp. 59–66, Lithoprobe, Vancouver, B. C., Canada.

Coflin, K. C., F. A. Cook, and W. T. Geis (1990), Evidence for Ellesmerian convergence in the subsurface east of the Mackenzie Delta, *Mar. Geol.*, *93*, 289–301.

Cook, D. G. (1991), Mackenzie Mountains, Franklin Mountains and Colville Hills, in *Geology of the Cordilleran Orogen in Canada*, edited by H. Gabrielse and C. J. Yorath, pp. 642–646, Geol. Surv. of Can., Ottawa, Ont., Canada.

Cook, D. G., and B. C. Maclean (1995), The intracratonic Forward orogeny, and implications for regional correlations, Northwest Territories, Canada, *Can. J. Earth Sci.*, *32*, 1991–2008.

Cook, F. A. (1988a), Proterozoic thin-skinned thrust and fold belt beneath the Interior Platform in northwest Canada, *Geol. Soc. Am. Bull.*, *100*, 877–890.

Cook, F. A. (1988b), Middle Proterozoic compressional orogen in northwestern Canada, *J. Geophys. Res.*, *93*, 8985–9005.

Cook, F. A. (1995), The reflection Moho beneath the southern Canadian Cordillera, *Can. J. Earth Sci.*, *32*, 1520–1530.

Cook, F. A. (2000), Proterozoic structural highs beneath the Mackenzie Mountains, northwest Canada, discovered with filtered potential field and seismic data, *Geosci. Can.*, *27*, 49–55.

Cook, F. A., and A. J. van der Velden (1993), Proterozoic crustal transition beneath the Western Canada Sedimentary basin, *Geology*, *21*, 785–788.

Cook, F. A., and A. J. van der Velden (1995), Three-dimensional crustal structure of the Purcell anticlinorium in the Cordillera of southwestern Canada, *Geol. Soc. Am. Bull.*, *107*, 642–644.

Cook, F. A., M. Dredge, and E. A. Clark (1992), The Proterozoic Fort Simpson structural trend in northwestern Canada, *Geol. Soc. Am. Bull.*, *104*, 1121–1137.

Cook, F. A., A. J. van der Velden, K. W. Hall, and B. J. Roberts (1999), Frozen subduction in Canada's Northwest Territories: Lithoprobe deep lithospheric reflection profiling of the western Canadian Shield, *Tectonics*, *18*, 1–24.

Creaser, B. S. (2000), Crustal seismic velocity structure of the northern Cordillera: Interpretation of LITHOPROBE SNORE Line 31, southern Yukon Territory, M. Sc. thesis, 134 pp., Univ. of Victoria, B. C., Canada.

Creaser, B., and G. Spence (2000), Crustal seismic velocity structure of the Northern Cordillera, southern Yukon Territory–LITHOPROBE SnoRE Line 31, in *Lithoprobe Slave-Northern Cordillera Lithospheric Evolution Transect and Cordilleran Tectonics Workshop*, edited by F. A. Cook and P. Erdmer, *Lithoprobe Rep. 72*, pp. 110–113, Lithoprobe, Vancouver, B. C., Canada.

Delaney, G. D. (1981), The mid-Proterozoic Wernecke Supergroup, Wernecke Mountains, Yukon Territory, in *Proterozoic Basins of Canada*, edited by F. H. A. Campbell, *Pap. Geol. Surv. Can.*, *81–10*, 1–23.

Diebold, J., T. Das, L. S. Hollister, and K. Rohr (1998), ACCRETE MCS data synthesis reflects tectonic history of the northern continental margin of British Columbia (abstract), *Eos Trans. AGU*, *79*(45), Fall Meet. Suppl., F803.

Driver, L. A., T. Chacko, R. Creaser, and P. Erdmer (1998), Petrogenesis of the Cretaceous Cassiar batholith, Yukon-B. C., Canada, in *Lithoprobe Slave-Northern Cordillera Lithospheric Evolution Transect and Cordilleran Tectonics Workshop*, edited by F. A. Cook and P. Erdmer, *Lithoprobe Rep. 64*, pp. 158–164, Lithoprobe, Vancouver, B. C., Canada.

Edwards, B. R., and J. K. Russell (1999), Northern Cordilleran volcanic province: A northern Basin and Range?, *Geology*, *27*, 243–246.

Eisbacher, G. H. (1981), Sedimentary tectonics and glacial record in the Windermere Supergroup, Mackenzie Mountains, northwestern Canada, *Pap. Geol. Surv. Can.*, *80-27*, 40 p.

Engebretson, D. C., A. V. Cox, and R. G. Gordon (1985), Relative motions between oceanic and continental plate in the Pacific basin, *Spec. Pap. Geol. Soc. Am.*, *206*, 59 pp.

Erdmer, P. (1987), Blueschist and eclogite in mylonitic allochthons, Ross Rover and Watson Lake, southeastern Yukon, *Can. J. Earth Sci.*, *24*, 1439–1449.

Erdmer, P., E. D. Ghent, D. A. Archibald, and M. V. Stout (1998), Paleozoic and Mesozoic high pressure metamorphism at the margin of ancestral North America in central Yukon, *Geol. Soc. Am. Bull.*, *110*, 615–629.

Evenchick, C. A. (1988), Stratigraphy, metamorphism, structure and their tectonic implications in the Sifton and Deserters ranges, Cassiar and northern Rocky Mountains, northern British Columbia, *Geol Surv. Can. Bull.*, *376*, 90 pp.

Evenchick, C. A. (1991a), Geometry, evolution and tectonic framework of the Skeena fold belt, north central British Columbia, *Tectonics*, *10*, 527–546.

Evenchick, C. A. (1991b), Structural relationships of the Skeena fold belt west of the Bowser Basin, northwest British Columbia, *Can. J. Earth Sci.*, *28*, 973–983.

Evenchick, C. A. (1991c), Jurassic stratigraphy of east Telegraph Creek and west Spatsizi map areas, British Columbia, *Pap. Geol. Surv. Can.*, *91-1A*, 155–162.

Evenchick, C. A., R. R. Parrish, and H. Gabrielse (1984), Precambrian gneiss and late Proterozoic sedimentation in north-central British Columbia, *Geology*, *12*, 233–237.

Evenchick, C. A., M. L. Crawford, V. J. McNicholl, L. D. Currie, and P. B. O'Sullivan (1999), Early Miocene or younger normal faults and other Tertiary structures in west Nass River map area, northwest British Columbia and parts of adjacent Alaska, *Pap. Geol. Surv. Can.*, *1999-A*, 1–11.

Fallas, K. M., P. Erdmer, D. A. Archibald, L. M. Heaman, and R. A. Creaser (1998), The St. Cyr Klippe, south-central Yukon: An outlier of the Teslin Tectonic Zone, in *Lithoprobe Slave-Northern Cordillera Lithospheric Evolution Transect and Cordilleran Tectonics Workshop*, edited by F. A. Cook and P. Erdmer, *Lithoprobe Rep. 64*, pp. 131–138, Lithoprobe, Vancouver, B. C., Canada.

Fernandez-Viejo, G., and R. M. Clowes (2003), Lithospheric structure beneath the Archean Slave Province and Proterozoic Wopmay Orogen, northwest Canada, from a Lithoprobe refraction/wide-angle reflection survey, *Geophys. J. Int.*, *153*, 1–19.

Francis, D., A. C. Abraham, and M. Polve (1999), Volcanic probes of the craton's edge beneath the northern Canadian Cordillera, in *Lithoprobe Slave-Northern Cordillera Lithospheric Evolution Transect and Cordilleran Tectonics Workshop*, edited by F. A. Cook and P. Erdmer, *Lithoprobe Rep. 69*, pp. 142–145, Lithoprobe, Vancouver, B. C., Canada.

Gabrielse, H. (1963a), Geology, Rabbit River, scale: 1:253,440, *Geol. Surv. Can. Map*, *46-1962*.

Gabrielse, H. (1963b), McDame map area, Cassiar district, British Columbia, *Mem. Geol. Surv. Can.*, *319*, 138 pp.

Gabrielse, H. (1967), Tectonic evolution of the northern Canadian Cordillera, *Can. J. Earth Sci.*, *4*, 271–298.

Gabrielse, H. (1985), Major dextral transcurrent displacements along the northern Rocky Mountain Trench and related lineaments in north central British Columbia, *Geol. Soc. Am. Bull.*, *96*, 1–14.

Gabrielse, H. (1998), Geology of Dease Lake (104J) and Cry Lake (104I) Map Areas, North-central British Columbia, *Bull. Geol. Surv. Can.*, *504*, 147 pp.

Gabrielse, H., and R. B. Campbell (1991), Upper Proterozoic assemblages, in *Geology of the Cordilleran Orogen in Canada*, edited by H. Gabrielse and C. J. Yorath, pp. 127–149, Geol. Surv. of Can., Ottawa, Ont., Canada.

Gabrielse, H., and C. J. Yorath (Eds.) (1991), *Geology of the Cordilleran Orogen in Canada*, 844 pp., Geol. Surv. of Can., Ottawa, Ont., Canada.

Gabrielse, H., J. W. H. Monger, J. O. Wheeler, and C. J. Yorath (1991), Morphogeologic belts, tectonic assemblages, and terranes, in *Geology of the Cordilleran Orogen in Canada*, edited by H. Gabrielse and C. J. Yorath, chap. 2, pp. 15–28, Geol. Surv. of Can., Ottawa, Ont., Canada.

Gehrels, G. E., W. C. McClelland, S. D. Samson, P. J. Patchett, and J. L. Jackson (1990), Ancient continental margin assemblage in the northern Coast Mountains, southeast Alaska and northwest Canada, *Geology*, *18*, 208–211.

Geiger, H., and F. A. Cook (2001), Analyses of crustal structure from bandpass and directionally filtered potential field data: An example from western Canada, *Can. J. Earth Sci.*, *38*, 953–961.

Geological Survey of Canada (1999a), Gridded isostatic anomaly data of Canada, 5 km grid spacing, Geophys. Data Cent., Ottawa, Ont., Canada.

Geological Survey of Canada (1999b), Gridded aeromagnetic data of Canada, 2 km grid spacing, Geophys. Data Cent., Ottawa, Ont., Canada.

Geological Survey of Canada (1999c), Geological map of Canada [CD-ROM], Ottawa, Ont., Canada.

Goodacre, A. K., R. F. Grieve, and J. F. Halpenny (1987), Isostatic gravity anomaly map of Canada, in *Canadian Geophysical Atlas*, Map 4, scale 1:1,000,000, Geol. Surv. of Can., Ottawa, Ont., Canada.

Gordey, S. P. (2002), Is Yukon-Tanana terrane North America?, in *Lithoprobe Slave-Northern Cordillera Lithospheric Evolution Transect and Cordilleran Tectonics Workshop*, edited by F. A. Cook and P. Erdmer, *Lithoprobe Rep. 82*, pp. 39–42, Lithoprobe, Vancouver, B. C., Canada.

Gordey, S. P., and A. J. Makepeace (1999), Yukon digital geology, *Geol, Surv. Can. Open File*, *D3826*.

Gordey, S. P., and R. I. Thompson (1991), Ancestral North America, in *Geology of the Cordilleran Orogen in Canada*, edited by H. Gabrielse and C. J. Yorath, chap. 17, pp. 625–628, Geol. Surv. of Can., Ottawa, Ont., Canada.

Gordey, S. P., J. G. Abbott, D. J. Tempelman-Kluit, and H. Gabrielse (1987), Antler clastics in the Canadian Cordillera, *Geology*, *15*, 103–107.

Green, A. G., B. Milkereit, A. Davidson, C. Spencer, D. R. Hutchison, W. F. Cannon, M. W. Lee, W. F. Agena, J. C. Behrendt, and W. J. Hinze (1988), Crustal structure of the Grenville front and adjacent terranes, *Geology*, *16*, 788–792.

Hammer, P. T. C., and R. M. Clowes (1999), Crustal structure across the Cordilleran orogen of northwest British Columbia, preliminary results of Lithoprobe SNORCLE refraction experiment line 22, in *Lithoprobe Slave-Northern Cordillera Lithospheric Evolution Transect and Cordilleran Tectonics Workshop*, edited by F. A. Cook and P. Erdmer, *Lithoprobe Rep. 69*, pp. 275–282, Lithoprobe, Vancouver, B. C., Canada.

Hammer, P. T. C., R. M. Clowes, and R. M. Ellis (2000), Crustal structure of NW British Columbia and SE Alaska from seismic wide-angle studies: Coast Plutonic Complex to Stikinia, *J. Geophys. Res.*, *105*, 7961–7981.

Harms, T. A., and R. A. Stevens (1996), Assemblage analysis of the Dorsey terrane, in *Lithoprobe Slave-Northern Cordillera Lithospheric Evolution Transect and Cordilleran Tectonics Workshop*, edited by F. A. Cook and P. Erdmer, *Lithoprobe Rep. 50*, pp. 199–201, Lithoprobe, Vancouver, B. C., Canada.

Harms, T. A., J. Nelson, and J. Bradford (1988), Geological transect across the Sylvester Allochthon north of Blue River, northern British Columbia (104 P/12), in *Geological Fieldwork, 1987*, *Paper 1988-1*, pp. 245–248, Br. Columbia Minist. of Energy, Mines and Pet. Resour., Victoria, B. C., Canada.

Heah, T. S. T. (1990), Eastern margin of the Central Gneiss Complex in the Shames River area, Terrace, British Columbia, *Pap. Geol. Surv. Can.*, *90-1E*, 159–169.

Hildebrand, R. S., and R. S. Baragar (1991), On folds and thrusts affecting the Coppermine River Group, *Can. J. Earth Sci.*, *28*, 523–531.

Hildebrand, R. S., P. F. Hoffman, and S. A. Bowring (1987), Tectonomagmatic development of the Great Bear magmatic arc, northwestern Canadian Shield, *J. Volcanol. Geotherm. Res.*, *32*, 99–118.

Hoffman, P. F. (1980), Wopmay orogen: A Wilson cycle of Early Proterozoic age in the northwest of the Canadian Shield, in *The Continental Crust and Its Mineral Deposits*, edited by D. W. Strangway, *Geol. Assoc. Can. Spec. Pap.*, *20*, 523–549.

Hoffman, P. F. (1987), Continental transform tectonics: Great Slave Lake shear zone (ca. 1.9 Ga), northwest Canada, *Geology*, *15*, 785–788.

Irving, E., P. J. Wynne, D. J. Thorkelson, and P. Schiarizza (1996), Large (1000 to 4000 km) northward movements of tectonic domains in the northern Cordillera, 83 to 45 Ma, *J. Geophys. Res.*, *101*, 17,901–17,916.

Kelley, K. (1993), Relative motions between North America and oceanic plate of the Pacific basin during the past 130 million years., M. Sc. thesis, 89 pp., Western Washington Univ., Bellingham, Washington.

Klepeis, K. A., and M. L. Crawford (1999), High-temperature arc-parallel normal faulting and transtension at the roots of an obliquely convergent margin, *Geology*, *27*, 7–10.

LeCheminant, A. N., and L. M. Heaman (1994), 779 Ma mafic magmatism in northwest Canadian Shield and northern Cordillera, *U. S. Geol. Surv. Circ.*, *1089*, 187 pp.

Lowe, C., R. B. Horner, J. K. Mortensen, S. T. Johnston, and C. F. Roots (1994), New geophysical data for the southern Yukon Territory: Implications for the structure and tectonics of the northern Canadian Cordillera, *Can. J. Earth Sci.*, *31*, 891–904.

Mihalynuk, M. G. (1999), Geology and mineral resources of the Tagish Lake area (NTS 104M8/9/10E/15 and 104N12W), *Bull. 105*, 201 pp., Br. Columbia Minist. of Energy and Mines, Victoria, B. C., Canada.

Mihalynuk, M. G., J. L. Nelson, and L. J. Diakow (1994), Cache Creek terrane entrapment: Oroclinal paradox within the Canadian Cordillera, *Tectonics*, *13*, 575–595.

Milkereit, B., and C. Spencer (1989), Noise suppression and coherency enhancement of seismic data: In *Statistical Application in the Earth Sciences*, edited by F. P. Agterberg and G. Bonham-Carter, pp. 243–248, Geol. Surv. of Can., Ottawa, Ont., Canada.

Mitchelmore, M. D., and F. A. Cook (1994), Inversion of the Proterozoic Wernecke basin during tectonic development of the Racklan Orogen, northwestern Canada, *Can. J. Earth Sci.*, *31*, 447–457.

Monger, J. W. H. (1977a), Upper Paleozoic rocks of northwestern British Columbia, *Pap. Geol Surv. Can.*, *77-1A*, 255–261.

Monger, J. W. H. (1977b), Upper Paleozoic rocks of the western Canadian Cordillera and their bearing on Cordilleran evolution, *Can. J. Earth Sci.*, *14*, 1832–1859.

Monger, J. W. H., J. O. Wheeler, H. W. Tipper, H. Gabrielse, T. Harms, L. C. Struik, R. B. Campbell, C. J. Dodds, G. E. Gehrels, and J. O'Brien (1991), Upper Devonian to Middle Jurassic, in *Geology of the Cordilleran Orogen in Canada*, edited by H. Gabrielse and C. J. Yorath, pp. 281–328, Geol. Surv. of Can., Ottawa, Ont., Canada.

Morozov, I., S. B. Smithson, L. Hollister, and J. Diebold (1999), Wide-angle seismic imaging across accreted terranes, southeastern Alaska and western British Columbia, *Tectonophysics*, *299*, 281–296.

Morozov, I. B., S. B. Smithson, J. Chen, and L. S. Hollister (2001), Generation of new continental crust and terrane accretion in southeastern Alaska and western British Columbia: Constraints from P- and S-wave wide-angle seismic data (ACCRETE), *Tectonophysics*, *341*, 49–67.

Morris, G. A., and R. A. Creaser (1998), The petrogenesis of the Mount Skukum/Bennett Lake igneous complex, Yukon Territory, and reflections on the Eocene tectonics of the Canadian Cordillera, in *Lithoprobe Slave-Northern Cordillera Lithospheric Evolution Transect and Cordilleran Tectonics Workshop*, edited by F. A. Cook and P. Erdmer, *Lithoprobe Rep. 64*, pp. 165–170, Lithoprobe, Vancouver, B. C., Canada.

Morris, G. A., and R. A. Creaser (1999), Mid-Cretaceous granites of the southern Yukon: Initial report, in *Lithoprobe Slave-Northern Cordillera Lithospheric Evolution Transect and Cordilleran Tectonics Workshop*, edited by F. A. Cook and P. Erdmer, *Lithoprobe Rep. 69*, pp. 120–121, Lithoprobe, Vancouver, B. C., Canada.

Nelson, J. L., and M. G. Mihalynuk (1993), Cache Creek ocean: Closure or enclosure?, *Geology*, *21*, 173–176.

Nelson, J., T. Harms, and J. Mortensen (1998), The southeastern Dorsey terrane: Characteristics and correlations, in *Lithoprobe Slave-Northern Cordillera Lithospheric Evolution Transect and Cordilleran Tectonics Workshop*, edited by F. A. Cook and P. Erdmer, *Lithoprobe Rep. 64*, pp. 279–288, Lithoprobe, Vancouver, B. C., Canada.

Ross, G. M., and C. Kerans (1989), Geology, Hornby Bay and Dismal Lakes Groups, Coppermine Homocline, District of Mackenzie, Northwest Territories, scale 1:250,000, *Geol. Surv. Can. Map, 1663A*.

Ross, G. M., H. C. Annell, and M. A. Hamilton (2000), Lithology and U-Pb geochronology of shallow basement along the SNORCLE line, southwest Northwest Territories: A preliminary report, in *Lithoprobe Slave-Northern Cordillera Lithospheric Evolution Transect and Cordilleran Tectonics Workshop*, edited by F. A. Cook and P. Erdmer, *Lithoprobe Rep. 72*, pp. 76–80, Lithoprobe, Vancouver, B. C., Canada.

Ross, G. M., M. E. Villeneuve, and R. J. Theriault (2001), Isotopic provenance of the lower Muskwa assemblage (Mesoproterozoic), Rocky Mountains, British Columbia): New clues to correlation and source areas, *Precambrian Res.*, *111*, 57–77.

Shi, S., D. Francis, J. Ludden, A. Frederiksen, and M. Bostock (1998), Xenolith evidence for lithospheric melting above anomalously hot mantle beneath the northern Canadian Cordillera, *Contrib. Mineral. Petrol.*, *131*, 39–53.

Smithson, S. B., J. A. Brewer, S. Kaufman, J. E. Oliver, and C. A. Hurich (1979), Structure of the Laramide Wind River uplift, Wyoming, from COCORP deep reflection data and from gravity data, *J. Geophys. Res.*, *84*, 5955–5972.

Snyder, D. B., R. M. Clowes, F. A. Cook, P. Erdmer, C. A. Evenchick, A. J. van der Velden, and K. W. Hall (2002), Proterozoic prism arrests suspect terranes: Insights into the ancient Cordilleran margin from seismic reflection data, *GSA Today*, *12*, 4–10.

Taylor, G. C., and D. F. Stott (1973), Tuchodi Lakes map area, British Columbia, *Mem. Geol. Surv. Can.*, *373*, 37 pp.

Tempelman-Kluit, D. J. (1979), Transported cataclasite, ophiolite, and granodiorite in Yukon: Evidence of arc-continent collision, *Pap. Geol. Surv. Can. 79-14*, 27 pp.
Thompson, R. I. (1981), The nature and significance of large 'blind' thrusts within the northern Rocky Mountains of Canada, in *Thrust and Nappe Tectonics*, *Geol. Soc. Spec. Publ.*, *9*, 449–462.
Thorkelson, D. J., J. G. Abbott, J. Mortensen, R. Creaser, and M. Villeneuve (1998), Proterozoic sedimentation, magmatism, metasomatism, and deformation in the Wernecke and Ogilvie Mountains, Yukon, in *Lithoprobe Slave-Northern Cordillera Lithospheric Evolution Transect and Cordilleran Tectonics Workshop*, edited by F. A. Cook and P. Erdmer, *Lithoprobe Rep. 64*, pp. 110–119, Lithoprobe, Vancouver, B. C., Canada.
Thorkelson, D. J., J. K. Mortensen, R. A. Creaser, G. J. Davidson, and J. G. Abbott (2001), Early Proterozoic magmatism in Yukon, Canada: Constraints on the evolution of northwestern Laurentia, *Can. J. Earth Sci.*, *38*, 1479–1494.
Villeneuve, M. E., R. J. Theriault, and G. M. Ross (1991), U-Pb ages and Sm-Nd signature of two subsurface granites from the Fort Simpson magnetic high, *Can. J. Earth Sci.*, *28*, 1003–1008.
Welford, J. K., R. M. Clowes, R. M. Ellis, G. D. Spence, I. Asudeh, and Z. Hajnal (2001), Lithospheric structure across the craton-Cordilleran transition of northeastern British Columbia, *Can. J. Earth Sci.*, *38*, 1169–1189.
Wheeler, J. O., and P. McFeely (1991), Tectonic assemblage map of the Canadian Cordillera and adjacent parts of the United States of America, *Geol. Surv. Can. Map, 1712A*, scale 1:2,000,000.
Young, G. M. (1984), Proterozoic plate tectonics in Canada with emphasis on evidence for a late Proterozoic rifting event, *Precambrian Res.*, *25*, 233–256.
Young, G. M., C. M. Jefferson, D. G. F. Long, G. D. Delaney, and G. M. Yeo (1979), Middle and Late Proterozoic evolution of the northern Canadian Cordillera and shield, *Geology*, *7*, 125–128.

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