

# Electromagnetic images of crustal structures in southern and central Canadian Cordillera<sup>1</sup>

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**Abstract:** Data from more than 400 magnetotelluric soundings, made since the early 1980's in the Canadian Cordillera over a 300 000 km<sup>2</sup> area between 49 and 53.5°N, are used to image qualitatively regional three-dimensional crustal variation in electrical conductivity by means of phase maps, phase–frequency sections, and maps of resistivity at depth. Two hundred of the soundings were acquired as part of Lithoprobe Southern Canadian Cordillera Transect activities, and their locations were coordinated with the seismic reflection and refraction experiments. The lower crust has a generally pervasive, low resistivity (1–100 Ω·m) throughout the Cordillera west of the Foreland Belt. Within this “Canadian Cordilleran Regional” conductor, the magnetotelluric data reveal both two-dimensional structures, with highest conductivities along the Coast Belt and Omineca Belt, and three-dimensional variation along geological strike. This conductor, mapped over a volume in excess of 10<sup>6</sup> km<sup>3</sup>, is most probably caused by fluids — saline waters and silicate melts — in fractures and along interconnected grain boundaries. The observed lateral variations in conductivity may result from variations in fracture density, temperature, and the sources of hot fluid, such as the subducting Juan de Fuca plate under the Coast Belt, and mantle upflow under the Omineca Belt. In addition, we report a major east–west-trending geophysical discontinuity in the upper and middle crust of the Omineca Belt at a latitude of 50°N, with highly resistive rocks (>1000 Ω·m) to the south and more conductive rocks to the north (30–300 Ω·m). Seismic refraction models, residual gravity, and filtered magnetic maps correlate changes in compressional-wave velocity, density, and magnetization along this cross-strike discontinuity.

**Résumé :** Les données recueillies dans plus de 400 sondages magnétotelluriques effectués depuis le début des années 1980 dans la Cordillère au Canada, représentant une aire plus grande que 300 000 km<sup>2</sup>, entre 49 et 53,5°N, sont utilisées pour imager qualitativement, à l'échelle régionale, les variations crustales en trois dimensions de la conductivité électrique à l'aide de cartes de phase, de coupes de fréquence–phase et de cartes de résistivité de niveau profond. Parmi ces sondages, 200 furent enregistrés lors des travaux du transect de la Cordillère méridionale du projet Lithoprobe, et leurs localisations ont été planifiées en coordination avec les levés de sismique réfraction et réflexion. La croûte inférieure présente généralement une faible résistivité (1–100 Ω·m), répandue au travers la Cordillère à l'ouest du Domaine de l'avant-pays. À l'intérieur des limites de ce conducteur de la « Cordillère canadienne régionale », les données magnétotelluriques révèlent, d'une part, une structure à deux dimensions avec les conductivités les plus fortes mesurées le long du Domaine côtier et du Domaine d'Omineca, et d'autre part, une variation à trois dimensions longeant la direction géologique. Ce conducteur, cartographié sur un volume excédant 10<sup>6</sup> km<sup>3</sup>, doit son origine la plus probable à la présence de fluides — eaux salées et liquides silicatés — dans les fractures et le long des surfaces limites des grains interconnectés. Les variations latérales de la conductivité observées peuvent résulter de variations du nombre de fractures par unité de surface, de la température et des sources de fluides chauds comme la plaque Juan de Fuca subductée qui plonge sous le Domaine côtier, et les liquides magmatiques ascendants qui sont issus du manteau sous la Domaine d'Omineca. En plus, nous décrivons une discontinuité géophysique majeure orientée est–ouest, apparaissant dans la croûte supérieure et médiane du Domaine d'Omineca à la latitude 50°N, avec des roches de résistivité élevée (>1000 Ω·m) au sud et des roches caractérisées par une plus grande conductibilité au nord (30–300 Ω·m). Les modèles de sismique réfraction, la gravité résiduelle et les cartes magnétiques filtrées permettent d'établir une mise en corrélation des changements de vitesses des ondes de compression, de la densité et de l'aimantation le long cette discontinuité qui transecte la direction.

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## Introduction

The Cordillera of North America is an excellent natural laboratory for studying accretionary tectonics, as a considerable amount of juvenile material has been added to the continental crust since the beginning of the Mesozoic (245 Ma). This contrasts with other Phanerozoic orogens, such as the Himalayan, Caledonian and Hercynian, which comprise mainly preexisting crustal material that has been reworked or remobilized (Samson et al. 1989; Monger 1993). It is in the Canadian Cordillera that terrane concepts of crustal accretion were first developed (Monger et al. 1972; Monger and Price 1979; Coney et al. 1980; Monger et al. 1982), and the recent Decade of North American Geology (DNAG) volume on the Canadian Cordillera (Gabrielse and Yorath 1991a) gives excellent background information on the state of knowledge of the tectonic framework of British Columbia up until the late-1980's. During the last 200 Ma, subduction processes added elongated strips of island-arc, pericratonic, and oceanic crust to North America, with associated intrusions of magmatic rocks, mainly from partial melting of the subducted plate (plates?). Notwithstanding the vast amount of geological knowledge of the region, fundamental questions remain to be answered (Gabrielse and Yorath 1991c). Some of these can be addressed by electromagnetic (EM) studies, such as the structure of the basement under the accreted terranes, and the distribution of fluids in the crust from the active subduction of the Juan de Fuca plate to the thrust and fold belt of the Rocky Mountains. Extensional faults in the Omineca Belt and Kootenay Arc, major strike-slip fault systems such as those of the Fraser River and northern Rocky Mountain Trench, and major normal faults like the Slocan Lake Fault have signatures in electrical conductivity that complement the vertical-incidence seismic sections. The high conductivities of molten silicates give EM signatures of magma chambers in and beneath the crust. Sulphide and other ore minerals also produce local enhancement of crustal conductivity. Accordingly, large programs of EM studies, using principally the natural-source magnetotelluric (MT) technique, were initiated and carried out within southern and central British Columbia during the last decade.

All available MT sounding sites are shown in relation to the belts in Fig. 1a, and provide the largest MT sampling of the Earth's crust in a region of active accretionary tectonics. The various MT data discussed herein comprise two main subsets, plus other, smaller subsets (Table 1). The larger of the two main subsets was recorded as part of the Canadian multidisciplinary Lithoprobe program of study of the Canadian Cordillera in southern British Columbia (Clowes et al. 1992), and MT soundings were made at over 200 sites, mainly along two east-west transects between latitudes 49 and 51°N, along the more northerly of which seismic reflection data were also acquired. These data cover a range of frequencies broad enough to respond to conductive structures from close to the surface to the upper mantle. The smaller subset of the two comprises 91 sites recorded by the University of Alberta (UofA), which extend coverage along strike to the north to 53.5°N, with a frequency range which limits penetration to principally the upper half of the crust. Together, the Lithoprobe and UofA data extend across the five recognized morphogeological belts of the Canadian Cordillera. Additional datasets were provided by industry

and as part of the federal Geothermal Energy Program of MT data acquisition around geothermal prospects.

In this paper, we present all these data together for the first time in a uniform way in both pseudosection and map displays. From these plots of MT parameters with spatial location and with decreasing frequency, which is equivalent to increasing depth penetration, we draw qualitative conclusions about the variation of electrical conductivity in this region of over 300 000 km<sup>2</sup>. We then present a new synthesis of the conductive structure of the region, and relate it to the accretionary evolution of the Cordilleran crust.

## Tectonic setting

We briefly outline the gross tectonic framework of the Canadian Cordillera, with reference to the morphogeological belts depicted in Figs. 1a and 1b, based in the main on the DNAG publication (Gabrielse and Yorath 1991a), in particular Gabrielse and Yorath (1991b). The west margin of the stable craton, termed Ancestral North America, was long-lived, from its rift inception in the Mid-Proterozoic to the Mesozoic (~1500 to ~400 Ma), and comprised a passive margin setting with a deep miogeocline. The Intermontane Superterrane (Composite Terrane I of Monger et al. 1982), consisting of a collage of terranes amalgamated outboard of Ancestral North America, docked with the craton in the Jurassic, with attendant eastward-directed subduction. A second superterrane, the Insular Superterrane (Composite Terrane II of Monger et al. 1982), lay in unknown relationship to the Jurassic continental margin. Subsequently, the Insular Superterrane collided with the amalgamated Intermontane Superterrane - North American craton in the mid-Cretaceous (ca. 90-100 Ma), possibly as a result of sinistral movements along the margin (Monger et al. 1994b), causing a compressive thickening of the crust and mantle lithosphere in the Intermontane Superterrane, and significant northward displacements of elements of the Insular Superterrane along dextral strike-slip faults. In southeastern British Columbia, a short-lived (~10 Ma) extensional event occurred during the Early Eocene, between 59 and 46 Ma (Parrish et al. 1988; Bardoux and Mareschal 1994), which extended and thinned the crust by about 50% (Bardoux and Mareschal 1994), resulting in the present-day crustal thickness of ~32-34 km from the original preextensional value of 50-60 km (Parrish et al. 1988; Bardoux and Mareschal 1994). This extension was accommodated mainly on east-dipping fault systems between 58 and 52 Ma, and subsequently on west-dipping ones from 52 to 45 Ma (Parrish et al. 1988). Such a thin crust for an orogen is anomalous, with values averaging 50 km for young orogens and 42 km for older orogens being typical (Christensen and Mooney 1995).

These successive tectonic and subsequent plutonic events are reflected in the five present-day morphogeological belts of the Canadian Cordillera (Figs. 1a, 1b):

The *Foreland Belt* (FB), previously termed the Rocky Mountains Belt or Rocky Mountains Fold-and-Thrust Belt, contains rocks mainly deposited from latest Proterozoic to Late Jurassic time along the ancient continental margin (formed by Late Proterozoic rifting and separation of a supercontinent), as well as late Mesozoic synorogenic clastics, deformed in folds and thrusts. These sequences were

foreshortened and translated northeastwards onto cratonic North American basement.

The *Omineca Belt* (OB) straddles the complex boundary between deformed miogeoclinal Foreland rocks and the easternmost accreted terrane (Composite Terrane I of Monger et al. 1982) of Mid-Jurassic age. The Omineca Belt comprises metamorphic rocks, derived mainly from clastic strata from the craton, with minor craton fragments and pericratonic terranes. Its evolution is complex; a record of Paleozoic intrusion, metamorphism, and deformation is overprinted by the record of Mesozoic (<185 Ma) compression, metamorphism, and granitic intrusion, and early Tertiary extension and intrusion. The east–west extension in Tertiary time was accommodated by listric normal faults bounding horsts such as the Valhalla and Shuswap metamorphic gneiss core complexes (Parrish et al. 1988; Ranalli et al. 1989).

The *Intermontane Belt* (IB) is more physiographically subdued than the two flanking belts, and comprises upper Palaeozoic to mid-Mesozoic marine volcanosedimentary sequences overlain by Cretaceous and Tertiary nonmarine sequences. The volcano-sedimentary sequences represent remains of former intraoceanic arcs and accretionary complexes that were accreted to rocks deposited on North America in the latest Early Jurassic, along a boundary now in the Omineca Belt. The belt is well defined by the aeromagnetic map of the Cordillera (see Fig. 5b).

The *Coast Belt* (CB), previously called the Coast Plutonic Complex, consists dominantly (>80%) of Late Jurassic to early Tertiary (165–46 Ma) granitic rocks, with minor septa of variously metamorphosed volcanic and sedimentary rocks whose protolith ages range from Paleozoic (possibly locally Proterozoic) to Tertiary.

The *Insular Belt*, which lies oceanward of the Coast Belt, comprises locally latest Proterozoic, to Paleozoic, to Neogene volcanic and sedimentary rocks and subordinate granitic rocks added to the continent since 165 Ma. The belt is bounded on its west by the small oceanic Juan de Fuca plate, which is currently subducting beneath Vancouver Island and the west coast of British Columbia and the states of Washington and Oregon. The Juan de Fuca plate is a remnant of the formerly extensive Farallon plate.

Subduction could have been episodic, with this offshore subduction at the present time being the third major subduction event of the last 200 Ma, each displaced to the southwest of its predecessor, in a fixed North America reference frame. However, it is more likely to have been almost continuous along this margin since at least the mid-Paleozoic, with westward “step-outs” when large terranes rafted in and blocked the subduction system (J.W.H. Monger, personal communication, 1995).

### Kootenay Arc

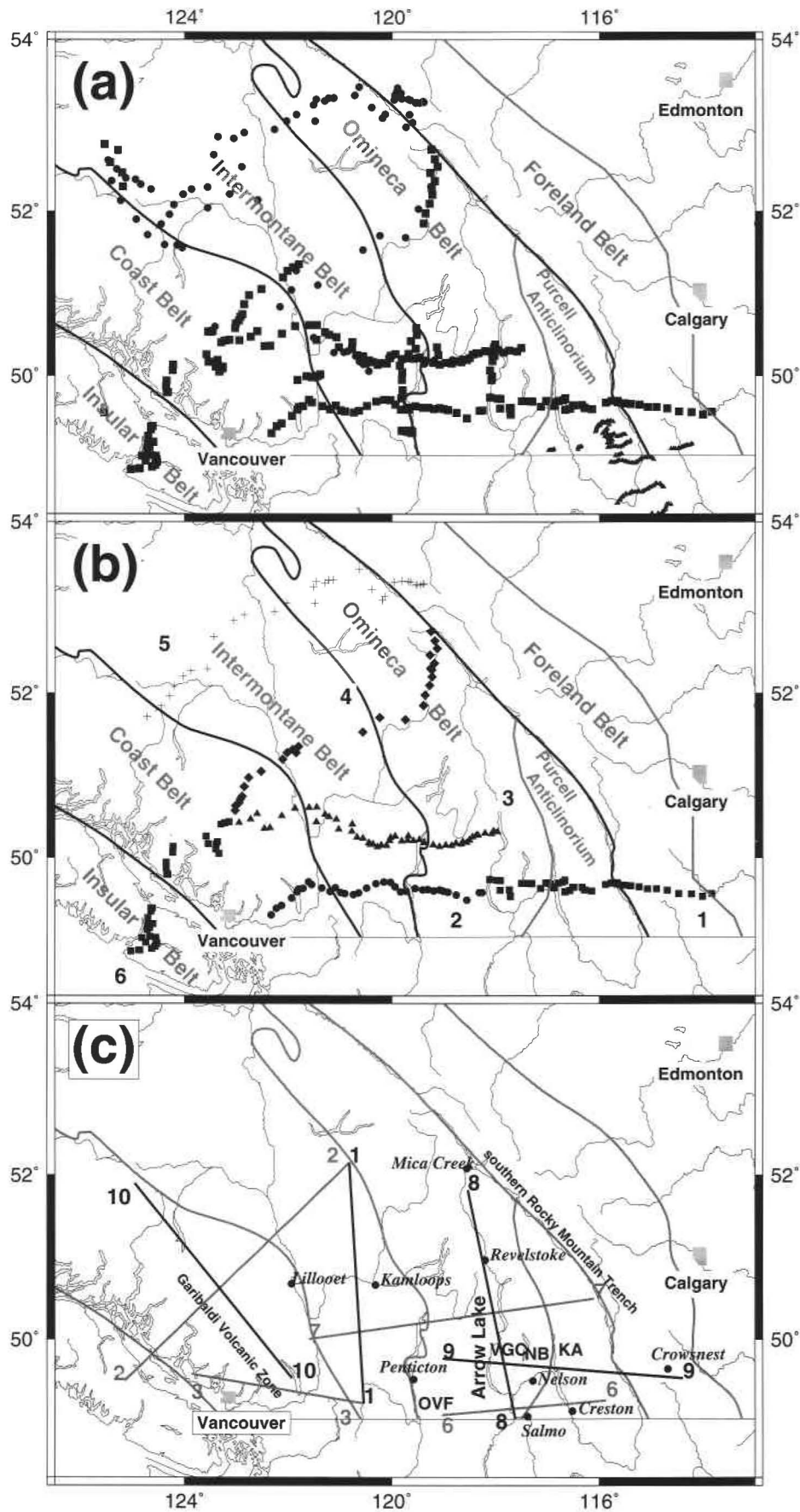
The boundary between the Foreland Belt and Omineca Belt is most recently defined, in the DNAG volume (Gabrielse and Yorath 1991a), as along the southern Rocky Mountain Trench (Gabrielse et al. 1991). Previously, this boundary was defined as along the Kootenay Arc, placing the Purcell Anticlinorium (PA) within the Foreland Belt rather than the Omineca Belt (e.g., Monger and Price 1979; Monger et al. 1982). The Kootenay Arc is shown by several geophysical and geological parameters to be a structurally significant fea-

ture. It marks the eastern limit of both Jurassic and Eocene magmatism (Armstrong and Ward 1991, 1993), the eastern limit of exposed metamorphic core complexes, is close to the eastern limit of the accreted terranes, and is the deep crustal location of a major extensional fault of crustal extent, the Slocan Lake Fault (SLF), which is exposed some 50 km to the west. This fault may have provided a conduit for mantle and lower crustal Pb and CO<sub>2</sub> to the upper crust (Beaudoin et al. 1991), which is important for EM studies in terms of the cause of enhanced conductivity. The contour showing initial <sup>87</sup>Sr/<sup>86</sup>Sr = 0.704 in Triassic – Early Jurassic igneous rocks lies along the Kootenay Arc (Armstrong and Ghosh 1990; Clowes et al. 1992). This low value is characteristic of mantle-derived rocks. A change in lower crustal compressional-wave velocity occurs near the Kootenay Arc (C.A. Zelt and White 1995). The first signature showing the Kootenay Arc as a major boundary was, however, in electrical conductivity, as will be discussed below.

### Eocene extension and timing of intrusives in the southern Omineca Belt

The record of Early to Middle Eocene extension in southeastern British Columbia is clearly displayed within a pie-shaped segment extending across the Intermontane Belt, Omineca Belt, and western Foreland Belt near latitude 49°N in the south to a point near the Malton Gneiss (north of Mica Dam, Fig. 1c) in the southern Rocky Mountain Trench at latitude 52.5°N (Monger et al. 1994a, and references therein). This extension exposed metamorphic core complexes, of which the most striking is the 350 km long north–south Okanagan Complex, of the Shuswap Complex, from the Columbia Plateau to the Monashee Dome, and possibly as far north as 52.5°N (Carr 1992). The extension took place between 59 and 46 Ma within a period of ~10 Ma (Parrish et al. 1988). The amount of extension at the latitudes of Nelson and Penticton (49.5°N, Fig. 1c) was assessed at 80% by Parrish et al. (1988), but a value of 52%, based on more conservative estimates of displacements on individual listric faults, was recently derived by Bardoux and Mareschal (1994). Farther north, at the latitude of Revelstoke (51°N, Fig. 1c), the magnitude of extension decreases to no greater than 30% (Brown and Journeay 1987; R.L. Brown, personal communication, 1995). The extension on the mappable faults ceases farther to the north at the location of the Mica Dam on Mica Creek (52°N, Fig. 1c, R.L. Brown, personal communication, 1995). Most of this extension was on normal fault systems that bound the metamorphic core complexes, with east-dipping faults mainly active between 58 and 52 Ma, and west-dipping faults dated at 52–45 Ma (Parrish et al. 1988).

Coeval with this extension was the “Kamloops–Challis–Absaroka” prominent magmatic episode, dated in the interval 55–40 Ma (Armstrong 1974; Armstrong et al. 1977; Armstrong and Ward 1991), which at its peak around 50 ± 1 Ma extended from southern Idaho to Alaska. It began in Canada and northern Washington at ~60 Ma, and was in decline by 45 Ma, ceasing by 40 Ma in the north but was transgressing southward to the Basin and Range tectonic province. Subsequent magmatism continued in the later Cenozoic sporadically throughout western North America (Armstrong and Ward 1991).



**Fig. 1.** (a) Location map of all MT soundings, with base map and outlines of the morphogeological belts. The squares and triangles denote MT sites at which data were acquired with a Phoenix system (frequency band 0.00055–384 Hz), whereas the circles denote sites at which measurements were made with UofA's SPAM system (0.016–64 Hz, no vertical magnetic field). (b) Location map with MT soundings along the six transects for which sections are presented. Line 1: ■; line 2: ●; line 3: ▲; line 4: ◆; line 5: +; line 6: ▣. Outlines of morphotectonic belts. (c) Location map of refraction lines, towns, and geological elements. KA, Kootenay Arc; NB, Nelson batholith; OVF, Okanagan Valley Fault; VGC, Valhalla Gneiss Complex.

**Table 1.** Dates, locations, and details of MT surveys in the 1980's in British Columbia.

Year	Location	Equipment	Site	References
1982	Mount Meager	MT-16	7	Flores et al. (1985) Jones and Dumas (1993)
1982	Thompson Valley	MT-16	10	
1983	Cayley Mountain	MT-16	5	Jones and Dumas (1993)
1984	Rocky Mountain Trench	SPAM Mk II	25	Hutton et al. (1987)
1984	Anahim Lake	MT-16	5	
1984	Vancouver Island	MT-16	26	Kurtz et al. (1986, 1990)
1984	White Lake Basin	MT-16	15	
1985	Flathead Basin	MT1	12	Gupta and Jones (1990)
1985–1986	Purcell Anticlinorium	MT-16	57	Gupta and Jones (1995) Cook and Jones (1995)
1986–1990	South-central B.C.	SPAM Mk II	101	Gough and Majorowicz (1992) Majorowicz and Gough (1994)
1987	Southeastern B.C.	MT-16	27	Jones et al. (1988 1993) Eisel and Bahr (1993)
1988	Nelson batholith	AET-EMAP	14	Jones et al. (1989)
1989	Southern B.C.	V-5	80	Jones et al. (1992a, 1992b)
1990	Southern B.C.	V-5	80	Jones et al. (1992a, 1992b)

Locally for the region of interest, just prior to the extension, in the Late Paleocene to Middle Eocene, there was emplacement in the southern Omineca Belt of anatectic leucogranites of the Ladybird suite (Carr 1992). The crust and lithosphere were heated as a result of overthickening from compressive deformation caused by the bulldozing of the continent into the welter of accreted terranes. The granites postdate the thermal peak of metamorphism, and were generated during the final stages of thrusting (Carr 1992). The granites extend as far north as the Adams River area (almost 52°N), and have the same geochemistry as those of the early Tertiary Keller Butte Suite of the Colville batholith in the Okanagan Dome, northeast Washington (Carr 1992).

During the subsequent extension, mantle-source syenites of the Coryell Suite were erupted. They are presently exposed between the Okanagan Valley Fault to the west, and the Valhalla Gneiss Complex to the east, in the southernmost part of the Omineca Belt. Their most northerly exposure is south of 50°N. The Coryell plutons have been dated at late Early to Middle Eocene ( $51.1 \pm 0.5$  Ma; Carr and Parkinson 1989), which prompted Templeman-Kluit and Parkinson (1986) to suggest that the Coryell intrusives were contemporaneous with extension across the low-angle west-dipping Okanagan shear zone. The earlier east-dipping extension phase predates these intrusions. Note that these intrusives do not extend as far north as the whole of the Eocene extension does, as expressed by the exposures of metamorphic core complexes, but are limited to the southernmost Omineca Belt. This will be discussed again later.

The cause for the onset of extension is contentious (see, e.g., Bardoux and Mareschal 1994), but what is known is that the onset is approximately coeval with the well-documented loss of the mantle lithosphere at about 60 Ma. Either the overthickening of the crust and upper mantle led to gravitational collapse and subsequent "passive" detachment of the mantle lithosphere (Coney and Harms 1984; Ranalli et al. 1989; Monger et al. 1994a), possibly aided, or even initiated, by a change in relative plate motions, which caused dextral transpression to become dextral transtension, and subsequently thinned the lithosphere. Or the mantle lithosphere was lost due to an "active" upwelling of asthenosphere caused by mantle convection, such as the continent overriding a spreading centre (e.g., Gough 1986a; Wilson 1991). Whichever mechanism is responsible, modelling suggests that rapid removal of the mantle lithosphere will quickly heat up the remaining lithosphere and always trigger extension (J.-C. Mareschal 1994).

## Geophysical results

The state of geophysical knowledge of the southern Canadian Cordillera prior to the advent of Lithoprobe studies is summarized in Gough (1986a) and Sweeney et al. (1991), and briefly in the introductory paper to this issue (Cook 1995). Those results we consider most pertinent and significant are taken up later, together with our new electromagnetic results.

The Lithoprobe southern Cordillera seismic reflection survey began with data acquisition along the western half of MT transect 1 (Fig. 1b), from the Rocky Mountain Trench

to the Valhalla Gneiss Complex (VGC, Fig. 1c) (Cook et al. 1987, 1988). The interpretations of these data are generally consistent with prior tectonic models (Monger et al. 1985). To the east, latest Proterozoic to Mesozoic miogeoclinal and platform stratified rocks of the Foreland Belt, developed along the long-lived western margin of North America, form an eastward-tapering prism detached from the lower crust along a regional décollement at mid-crustal depths (12–25 km). These rocks have been shortened (by 50%) and translated up to 200 km to the northeast onto the craton, during Late Jurassic to early Tertiary convergence between the craton and allochthonous terranes to the west (Monger et al. 1982). The Foreland Belt is dominantly “thin-skinned” in structural style, with deformation considered to be concentrated mainly in the sedimentary wedge above autochthonous North American basement.

These structures are interpreted to be truncated beneath the Kootenay Arc (KA, Fig. 1c) by a major normal fault of Eocene age (Parrish et al. 1988), the Slocan Lake Fault, which possibly extends through the whole crust. Varsek and Cook (1991) present two tectonic models of this boundary, based on the seismic data, differing only in the style of early development of the accretionary complex. In both models, the Slocan Lake Fault offsets autochthonous North American basement by 7–8 km, east side down, and may offset the Moho some 75 km east of the surface expression of the Slocan Lake Fault. Interpretation of isotopic ratios places the western edge of North American basement about 100 km west of the Slocan Lake Fault outcrop, within the Omineca Belt (Armstrong et al. 1977; Armstrong and Ghosh 1990). Discussing the complete seismic reflection dataset from the Lithoprobe Southern Canadian Cordillera transect, Cook et al. (1991) suggest that the North American basement extends some 500 km west of the Slocan Lake Fault as far as the Fraser Fault, at the boundary between the Intermontane and Coast belts. Near the western end of transect 1 (Fig. 1b), beneath the Valhalla Gneiss Complex (VGC, Fig. 1c), the upper crust is seismically transparent to about 2.0–4.0 s, to a prominent reflector, named the Valhalla reflector, beneath which is a complex reflective zone down to about 8 s (22 km). Few reflections are seen in the lower crust, depth range 22–35 km, beneath the Valhalla Gneiss Complex. Weak reflections at about 35 km near its western edge may represent the Moho (Cook et al. 1987, 1988; Eaton and Cook 1990).

In the Intermontane and Omineca belts, the data are interpreted to have imaged three large crustal antiforms or structural culminations. These are, from east to west, the Monashee Complex, the Vernon antiform, and the central Nicola horst, and were initially formed during Jurassic to Eocene compression and subsequently modified and exposed as a consequence of Eocene extension. The eastern two of these are interpreted to be cored by para-autochthonous North American rocks, whereas the Nicola horst is interpreted to be cored by allochthonous rocks. Beneath these culminations, the lower crust is interpreted to be a relatively thin tongue of modified North American cratonic basement, to the Fraser Fault. Finally, the crust–mantle boundary, as imaged as the reflection Moho, beneath these belts is remarkably flat at ~32–33 km with less than 1 s of relief over a distance of more than 250 km. This contrasts sharply with

the high structural relief in the crust, and with the obvious surface topographic relief.

The Fraser Fault is preferentially interpreted to be a listric fault dipping west into a mid-crustal reflective zone at 6.3 s (Varsek et al. 1993). To the west, rocks of the Intermontane Belt are interpreted to have been wedged into the middle crust of the eastern and central Coast Belt during contraction in the Late Cretaceous. The west side of the Coast Belt exhibits deep east-dipping reflections (16 s), which can be traced updip to reflections on the sections from Vancouver Island (Clowes et al. 1987). Moho is at 11–11.5 s beneath the Coast Belt. The “C reflector” on Vancouver Island is interpreted as the base of Wrangellia, whereas the deeper “E reflector” is a shear zone that can be traced sub-Moho beneath the western Coast Belt. Below the E reflector to the top of the descending Juan de Fuca oceanic plate is subducted lower continental crust and mantle material (Varsek et al. 1993).

The SCoRE89 and SCoRE90 seismic refraction experiments within the Southern Canadian Cordillera Transect of Lithoprobe recorded shots along eight lines (Fig. 1c) with 1 km receiver spacing (D.J. White et al. 1992; B.C. Zelt et al. 1992; O’Leary et al. 1993; Kanasevich et al. 1994; C.A. Zelt and White 1995; Clowes et al. 1995). The published models for the three strike lines within the eastern Coast Belt, Intermontane Belt, and Omineca Belt (line 10: O’Leary et al. 1993; line 1: B.C. Zelt et al. 1992; line 8: Kanasevich et al. 1994; respectively) show relatively uniform along-strike crustal structure within each belt. The model of the Coast Belt line (line 10, Fig. 1c) includes a velocity increase, from 6.20–6.35 km/s to 6.35–6.40 km/s, at 9–10 km, which is interpreted to be the base of the pluton. The middle and lower crust is relatively slow, 6.35–6.50 km/s, with a thin higher velocity layer, 6.65–6.80 km/s, just above the Moho. Moho has an average depth of 34 km, and exhibits some lateral variation with long wavelength undulations of 2–3 km in height.

For the Intermontane Belt line (line 1, Fig. 1c), the upper crust shallows from 15 km in the south to 11 km to the north. Below this, the middle crust is slow, 6.2–6.4 km/s, but the lower crust is relatively fast, with compressional-wave velocities from 6.5 km/s at the top increasing to 7.0 km/s towards the base. Moho is transitional, with its top at about 32–33 km depth, but with 2–3 km undulations shallowing to 30 km beneath the southern part of the line. The thin transitional layer is of high-velocity material, here 7.5–7.7 km/s, and has a flat base. The base of the crust is thus flat at 33–34 km. The top of the Moho transition layer correlates well with “reflection Moho” imaged by the vertical incidence data.

The model for the Omineca Belt line (line 8, Fig. 1c) exhibits a thick mid-crustal low-velocity zone, of 6.0–6.1 km/s and thickness 7–10 km. The depth to its top varies significantly, from 12 to 18 km, with the deepest part to the north and shallowest to the south. Its base is also deeper to the north (26 km) and shallower to the south (21.6 km). Beneath this, the velocities in the lower crust are typical of young crust, from 6.4–6.5 km/s at the top to 6.5–6.8 km/s at the base. The top of the Moho transition zone is at 35–37 km, with a slight north–south dip. The transition layer is generally 1–2 km thick. For this belt, there appears to be

a minor disagreement between the reflection Moho at 35 km, and the 1–2 km thick refraction transitional Moho starting at 35–37 km and ending at 37–38 km. However, this difference is within the modelling accuracies of refraction data ( $\pm 2$  km, D. White, personal communication, 1995) and the errors introduced by adoption of erroneous velocities when converting the reflection times to depth.

For the dip line crossing the Foreland and Omineca belts (line 9, Fig. 1c), the model shows a crust increasing in thickness eastward, from 35 km beneath the Omineca Belt to 43 km beneath the Foreland Belt (C.A. Zelt and White 1995), and is consistent with the location suggested by the models based on seismic reflection data (Cook et al. 1987, 1988). The refraction results indicate higher compressional-wave velocities in the crust below 20 km depth, in the Omineca Belt (6.6–6.7 km/s), than beneath the Foreland Belt (6.2–6.5 km/s, mainly 6.4 km/s). The velocity changes over an east–west distance of 85 km or less, at the location of the Slokan Lake Fault (D.J. White et al. 1992; C.A. Zelt and White 1995). Such a velocity change argues against a simple continuation of unmodified North American basement to the Fraser Fault, as implied in the geometries of Cook et al. (1987, 1988). In addition, the lower crust beneath the Omineca and Intermontane belts, with velocities observed of 6.5 km/s at the top to 6.8 km/s at the base, shows no evidence for a thick high-velocity basal layer of  $\sim 7.1$  km/s typical of the North American craton (Kanasewich et al. 1994, and references therein). However, the Foreland Belt is also anomalous compared with the rest of the North American craton in not exhibiting such a high-velocity layer.

On the scale of the continental lithosphere, several geophysical signatures show strong differences between the Foreland Belt and Omineca Belt. The crust is thicker in the Foreland Belt (41 km) than in the Omineca Belt (34 km), as already noted. Lowe and Ranalli (1993) model temperatures 250°C higher under the Omineca Belt (1100–1200 K) than under the Foreland Belt (850–950 K), where they indicate a brittle rheology in the lower crust. Majorowicz et al. (1993b) and Marquis et al. (1995) show a high correlation between depths to the top of the crustal conductor and the 450°C isotherm, both rising to a minimum depth of 10–15 km in the Omineca Belt. Finally, there is strong evidence that the lithosphere thickness is much smaller under the Canadian Cordillera (40–50 km) than under the craton (130 km) (Wickens 1971; Fulton and Walcott 1975; Gough 1986a; Lowe and Ranalli 1993).

### Previous results from electromagnetic geophysics in southwestern Canada

An indicator of conductivity variation used during the 1960's was to compute the ratio  $I = (\text{vertical magnetic field amplitude})/(\text{total horizontal magnetic field amplitude})$ , at various periods from three-component magnetic observations at many locations, and to compare them graphically. From early in the decade 1960–1970, magnetovariation studies showed that the crust and upper mantle of southern British Columbia have higher electrical conductivity (low  $I$ ) than the craton (high  $I$ ). Hyndman (1963) discovered a sharp change in  $I$  across the Kootenay Arc from low values to the west to high values to the east, at a latitude of 49.5°N. Caner and

Cannon (1965) showed that low values of  $I$  are typical for the North American Cordillera, and coupled this fact with high heat flows and low  $P_n$  seismic velocities to argue for a rise in mantle isotherms beneath western North America, as proposed by W.R.H. White and Savage (1965).

Caner et al. (1967) used an east–west profile of magnetovariational stations at 51°N to show a marked change at longitudes of 117–118°W from low to high values of  $I$ , consistent with Hyndman's result at 49.5°N. They showed that a model with a highly conducting layer (resistivity of  $1 \Omega \cdot \text{m}$ ) at depths of 25–35 km accounted for observed  $I$  values in southwestern British Columbia west of Kootenay Lake. Caner et al. (1969) later made MT soundings across southern British Columbia and Alberta. In British Columbia east of Kootenay Lake and in Alberta, their results were fitted by resistive crust ( $>250 \Omega \cdot \text{m}$ ) 30–35 km thick over a more conductive mantle ( $30\text{--}50 \Omega \cdot \text{m}$ ) (Caner 1971). In contrast, soundings west of Kootenay Lake showed a lower crustal conductive layer ( $10 \pm 5 \Omega \cdot \text{m}$ ) 20–40 km thick over a mantle of the same resistivity as that under their eastern stations. Twenty magnetovariational sites were used by Lajoie and Caner (1970) in a detailed study of the transition zone near Kootenay Lake. They found that the lower crustal resistive structure was more complex than a simple two-dimensional (2D) change at 117–118°W, and modelled their results in three dimensions with a major structural change from north–south to east–west just north of the United States–Canada border, passing beneath the town of Salmo (Fig. 1c). Further magnetovariational studies in south-central British Columbia mapped the region of anomalously low  $I$  up to the Rocky Mountain Trench (Caner et al. 1971). Caner et al. (1971) proposed speculative models based on these results, with a lower crustal conductor west of Kootenay Lake and a lower crustal–upper mantle conductor farther south. Dragert and Clarke (1977) refined these models by suggesting that the southern conductor might be laterally bounded and trend near 45°N. They tentatively associated it with a buried Precambrian rift in southern Alberta (Kanasewich 1968). Farther north, Dragert (1973) used magnetovariational fields to locate a highly conductive crustal feature along the Rocky Mountain Trench near 53.5°N.

The magnetovariational studies so far noted were made with small numbers of magnetometers, usually in a line across a structure of interest. The development of large 2D arrays of Gough–Reitzel magnetometers (Gough and Reitzel 1967) enabled conductive structures to be discovered and mapped over an extensive region; Lilley (1975), Gough and Ingham (1983), and Gough (1989) give reviews. The first 2D-array studies revealed highly conductive uppermost mantle beneath the western United States (Reitzel et al. 1970; Porath et al. 1970), which correlated with high heat flow and seismological indications of anomalously high temperatures (Gough 1974). In the Canadian Cordillera, 2D arrays (Camfield et al. 1970; Gough et al. 1982) confirmed the presence of a highly conductive lower crust over much of the region, which Gough (1986a) named the Canadian Cordilleran Regional conductor. The eastern limit of the Canadian Cordilleran Regional conductor lies along Kootenay Lake and the Rocky Mountain Trench north of 51.5°N, in agreement with the results of Lajoie and Caner (1970), Dragert (1973), and Dragert and Clarke (1977). Bingham et al.

**Table 2.** Characteristics of the MT systems used.

Equipment	Manufacturer	No. of channels	Frequency range
MT-16	Phoenix (Toronto)	7 or 10	384 Hz – 1820 s
SPAM Mk II	Edinburgh University	5	162.5 Hz – 80.5 s
MT1	Z-Axis	5	250 Hz – 1052 s
AET-EMAP	AET (Houston)	10	256 Hz – 362 s
V-5	Phoenix (Toronto)	7	384 Hz – 1820 s

(1985), using a more closely spaced 2D array from the Omineca Belt to the Alberta foothills between 51.5 and 54.5°N, showed that in this latitude range a highly conductive crust lies beneath the western half of the Main Ranges of the Rocky Mountains as well as beneath the Rocky Mountain Trench, where it earlier had been discovered by Dragert (1973). MT soundings by Hutton et al. (1987) provided resistivity–depth sections that confirmed the extension of the highly conductive crust beneath the Main Ranges. Ingham et al. (1987) showed that magnetovariational responses, along a profile across strike of the Cordillera from Edmonton to Lillooet, fit a model with generally very conductive crust in the Cordillera and complicated structure near the Rocky Mountains and Rocky Mountain Trench, and much more resistive crust in the pre-Mesozoic continent. Gough et al. (1989) used the magnetovariational array deployed in the EMSLAB (Electromagnetic Studies of the Lithosphere and Beyond—Juan de Fuca) international experiment to show that the Canadian Cordilleran Regional conductor extends southward in the U.S.A. to link with thicker regional conductors beneath the Cascade volcanic belt and the Basin and Range tectonic province. The same array showed that a linear crustal conductor that crosses southern Alberta and the southeast corner of British Columbia (Gough 1986a; Wang 1988; Wang et al. 1989) ends at Creston, British Columbia, where the Kootenay River crosses the international boundary. Gough et al. (1989) suggested that the currents in this “southern Alberta – British Columbia” (SABC) conductor disperse in the widespread Canadian Cordilleran Regional conductive crust there. This SABC conductor is the one originally discovered by Dragert and Clarke (1977). Recent MT profiles in Alberta, as part of the Lithoprobe Alberta basement transect activities, have mapped highly conductive features in the basement spatially coincident with the “Red Deer High” (Boerner et al. 1995) at the boundary of the sub-Phanerozoic Lacombe and Laverna domains (Ross and Parrish 1991). The conjectured westward extrapolation of this domain boundary (Ross and Parrish 1991; Ross et al. 1995) follows almost exactly the trace of the SABC mapped by Wang (1988) in his inversion of magnetovariational array data for causative currents, which associates the SABC with the basement boundary between the Archean Hearne Province and the Proterozoic Rimbey magmatic arc. Boerner et al. (1995) suggest that the anomalies in enhanced conductivity are within the top 20 km of the crust, not in the range 30–120 km as reported by Wang (1988), and that they are caused by graphitic metasedimentary rocks in the euxinic-flysch facies of a Paleoproterozoic foredeep sequence, similar to that of the Wopmay Orogen, rather than fluids as suggested by Wang (1988).

### Magnetotelluric soundings since 1980 in the Canadian Cordillera

For more quantitative study of structure, with higher definition, a good technique is to locate conductive structures with one or more magnetovariational arrays, and then make MT soundings across them. This two-stage process was used by Bingham et al. (1985) and Hutton et al. (1987) on the interface between craton and Cordillera between 51°N and 54°N. On a larger scale, numerous MT soundings in the Cordillera reveal many significant structures within this region of the Earth's crust, which magnetovariational responses show simply as generally conductive. Magnetotelluric data have been acquired in a number of surveys since 1980, and the locations of all of the soundings are shown in Fig. 1a and those used in discussion of six phase–frequency sections, in Fig. 1b. The dates and locations of the surveys are listed in Table 1, and the characteristics of the equipment used, in Table 2. The 1984 Vancouver Island survey and the surveys in 1987, 1988, 1989, and 1990 were undertaken under the auspices of Lithoprobe.

For viewing the data in pseudosection format, we have assigned the MT sites to six lines based on their spatial coverage (Fig. 1b):

*Line 1* is an east–west line starting on the eastern limit of the Cordilleran deformation front, crossing the Foreland Belt and Purcell Anticlinorium, and ending at Arrow Lake on the western edge of the Valhalla Gneiss Complex in the middle of the Omineca Belt. Seismic reflection lines 2–5 were shot along the western half of this line (Cook et al. 1987, 1988).

*Line 2* is an east–west line starting to the west of Arrow Lake, in the middle of the Omineca Belt, crossing the whole of the Intermontane Belt, and ending in the middle of the Coast Belt.

*Line 3* is an east–west transect parallel to line 2 some 100–150 km farther north. Seismic reflection lines 7–13 were shot along this line (Cook et al. 1991, 1992).

*Line 4* is a northeast–southwest line starting at the northern end of the North Thompson Valley at the edge of the Foreland and Omineca belts, crossing the Omineca and Intermontane belts in their entirety, and ending at the centre of the Coast Belt. There is sparse coverage of the Omineca–Intermontane boundary, and no coverage in the centre of the Intermontane Belt.

*Line 5* is a line parallel to line 4, starting in the northeast on the east side of the Rocky Mountain Trench, crossing both the Intermontane and Omineca belts in their entirety, and ending just on the Coast Belt. All of the sites on this line were recorded with the UofA's SPAM Mark II MT system, with



its narrower recording band.

*Line 6* is a northeast–southwest line that starts at the middle of the Coast Belt, on the east side of the Garibaldi Volcanic Belt (Fig. 1c), crosses the volcanic belt (Mounts Cayley and Meager), runs along the Powell River road, and includes the MT sites recorded by Kurtz et al. (1986, 1990) on Vancouver Island. Seismic line 16 is coincident with the Powell River MT stations, and line 15, with some of the stations close to the volcanic front (Cook et al. 1991; Varsek et al. 1993).

### Conductive structures of the southern Canadian Cordilleran crust

The conductive structures of the Cordilleran region are discussed with reference to phase–frequency sections and two sets of maps. The phases we choose to display for the purposes of showing the qualitative lateral and vertical variation of conductivity are the MT phases for current travelling in the direction N30W. This is the dominant direction of the morphogeological belts and associated structures, and so defines the ‘E-polarization’ MT responses for a 2D Earth. The MT phase is the phase lead of the observed horizontal electric field component over the perpendicular horizontal magnetic field component. Above a uniform half-space, the MT phase will be constant at a value of 45° and frequency independent. If, in a multidimensional Earth, there exists a zone of resistivity lower (conductivity higher) than that of the region above it, then the MT phase will increase above 45° to a maximum then relax back to 45° as frequency decreases, that is, as penetration into the Earth increases (see, e.g., Jones 1992). The theoretical limit for 1D (one-dimensional) and 2D Earths is a phase of 90° when there exists a perfect conductor (infinite conductivity) at depth. For a resistive structure, then, the converse is true: MT phase will drop below 45° then recover to 45° with decreasing frequency. The limit for a perfect resistor is 0°, that is, the electric and magnetic horizontal field components are totally in phase.

The phase–frequency sections (Fig. 2) indicate structure along the six MT lines located in Fig. 1b. In these figures, the horizontal axis shows distance along the MT line, and the vertical axis, frequency on a logarithmic scale. Such phase–frequency sections are related nonlinearly to true depth sections, with lower frequencies corresponding to greater depths. Warm colours (yellows and reds) indicate more conductive regions, cold colours (greens and blues) resistive regions. Figure 3 shows phase maps at three frequencies, corresponding approximately to penetration to the upper crust, middle crust, and lower crust respectively. Figure 4 shows maps of estimated resistivity from Niblett–Bostick transformations (Niblett and Sayn-Wittgenstein 1960; Bostick 1977; Jones 1983) of the E-polarization responses at each MT site, at depths of 5, 10, 20, and 30 km. The Niblett–Bostick depth transform of the MT data is a representation of the data in terms of an approximate value of true resistivity at a given depth, instead of period, and gives first-order qualitative information about the gross subsurface conductivity variation. For more precise definition, especially of fine features, multidimensional modelling and inversion must be applied. We will describe features on these sections and maps, from east to west.

In the Foreland Belt, the lower crust is quite resistive, as shown by the cold colours in Fig. 2 at frequencies of 0.01–1 Hz. Modelling in progress of the data from the eastern half of line 1 suggests a lower crust in excess of 300  $\Omega \cdot \text{m}$  (Gupta and Jones<sup>3</sup>). This value is commensurate with the earlier estimate of Caner et al. (1969). For the part of line 1 crossing the eastern part of the Purcell Anticlinorium, Gupta and Jones (1995) analyzed data from almost 200 MT sites on both sides of the United States–Canada border (small triangles in Fig. 1a), and found very low resistivities (1  $\Omega \cdot \text{m}$ ) at a few kilometres depth in the upper crust, which they ascribe to mineralization (sulphides, etc.). The magnetovariational responses for the central part of line 1 (not shown) exhibit a strong conductivity anomaly, with a reversal in the real induction arrows, at long periods (~100 s) for sites just to the west of the Rocky Mountain Trench. Comparison with the published current-concentration maps of Wang (1988) identifies this anomaly with the SABC conductor described above. This conductor is apparent in the Niblett–Bostick resistivity maps as the localized two- or three-station anomaly at 49.5°N, 116°W (Fig. 4). As discussed above, this position marks the approximate location of the basement boundary between the Archean Hearne Province and the Proterozoic Rimbey magmatic arc (Ross and Parrish 1991), and recent MT profiles in Alberta have associated enhanced crustal conductors with graphitic foredeep sequences (Boerner et al. 1995).

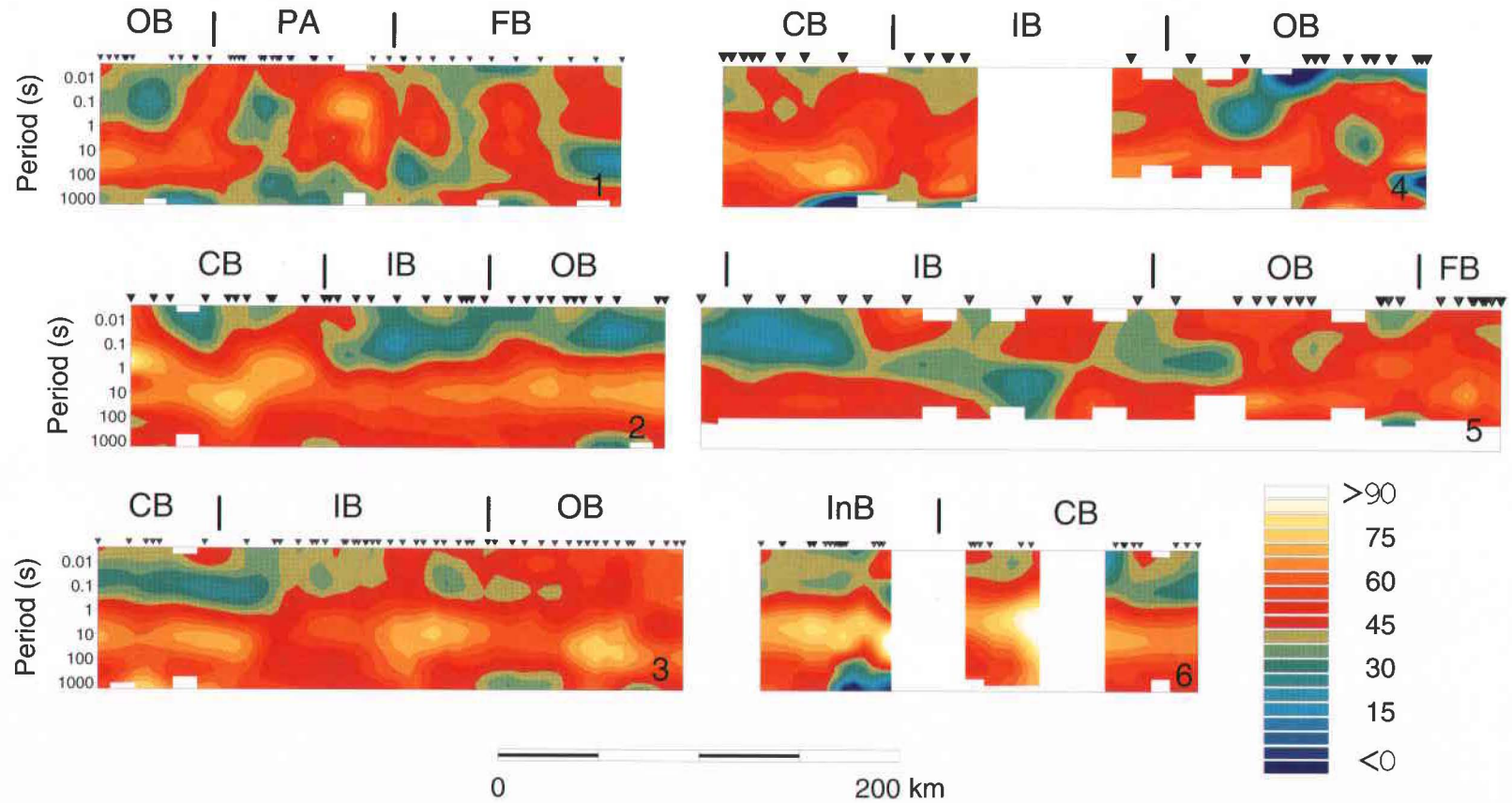
Along the three southern profiles (lines 1–3 in Fig. 1b), the lower crust is generally conductive west of the Kootenay Arc, as is seen in Figs. 2 and 3 from the high phases at low frequencies (0.01–1 Hz) corresponding to the lower crust. These MT soundings thus confirm the existence of the Canadian Cordilleran Regional conductor mapped with magnetometer arrays and that its eastern boundary is near the Kootenay Arc.

High conductivities occur under the Omineca Belt (Jones et al. 1992a; Marquis et al. 1995). High phase values in the 33 Hz map (Fig. 3a) over much of the Omineca Belt indicate high conductivities rising into the upper crust. Majorowicz et al. (1993a, 1993b) and Marquis et al. (1995) have estimated depths to the 450°C isotherm from heat flow measurements in the Cordilleran region, and found that this depth correlated with the depth to a mid-crustal conductor. Both depths have values of ~10 km in the southern Omineca Belt (8–10 km, Marquis et al. 1995; 10–12 km, Majorowicz et al. 1993a), compared with ~17 km in the southern Intermontane Belt (15–17 km, Marquis et al. 1995; 15–20 km, Majorowicz et al. 1993a). This variation at depth on either side of the Intermontane–Omineca boundary is evident in the Niblett–Bostick map for depths of 20 and 30 km (Figs. 4c and 4d, respectively).

The 33 Hz map (Fig. 3a) shows generally resistive upper crust in both the Intermontane and Coast belts, with local high-conductivity anomalies probably associated with conducting clays in the Coast Belt (see Jones and Dumas 1993). A small resistive block in the Intermontane Belt near 52.7°N, 123°W is seen in line 5 data (Figs. 3b and 3c). In

<sup>3</sup> J.C. Gupta and A.G. Jones. Electrical conductivity structure of a deformed continental margin: southwestern Alberta to southeastern British Columbia, Canada. In preparation.

**Fig. 2.** Phase–frequency pseudosections for all six lines (line number in bottom right corner of each pseudosection). The site locations are indicated by the inverted triangles. FB, Foreland Belt; PA, Purcell Anticlinorium; OB, Omineca Belt; IB, Intermontane Belt; CB, Coast Belt; InB, Insular Belt. The warm colours (reds to yellows) represent phases greater than  $45^\circ$ , and depict an increase in conductivity with increasing period (depth), whereas the colder colours (greens and blues) depict the converse.



**Fig. 3.** Phase maps at frequencies of (a) 33, (b) 1, and (c) 0.033 Hz. These frequencies generally correspond to penetration depths equivalent to upper, middle, and lower crust respectively.

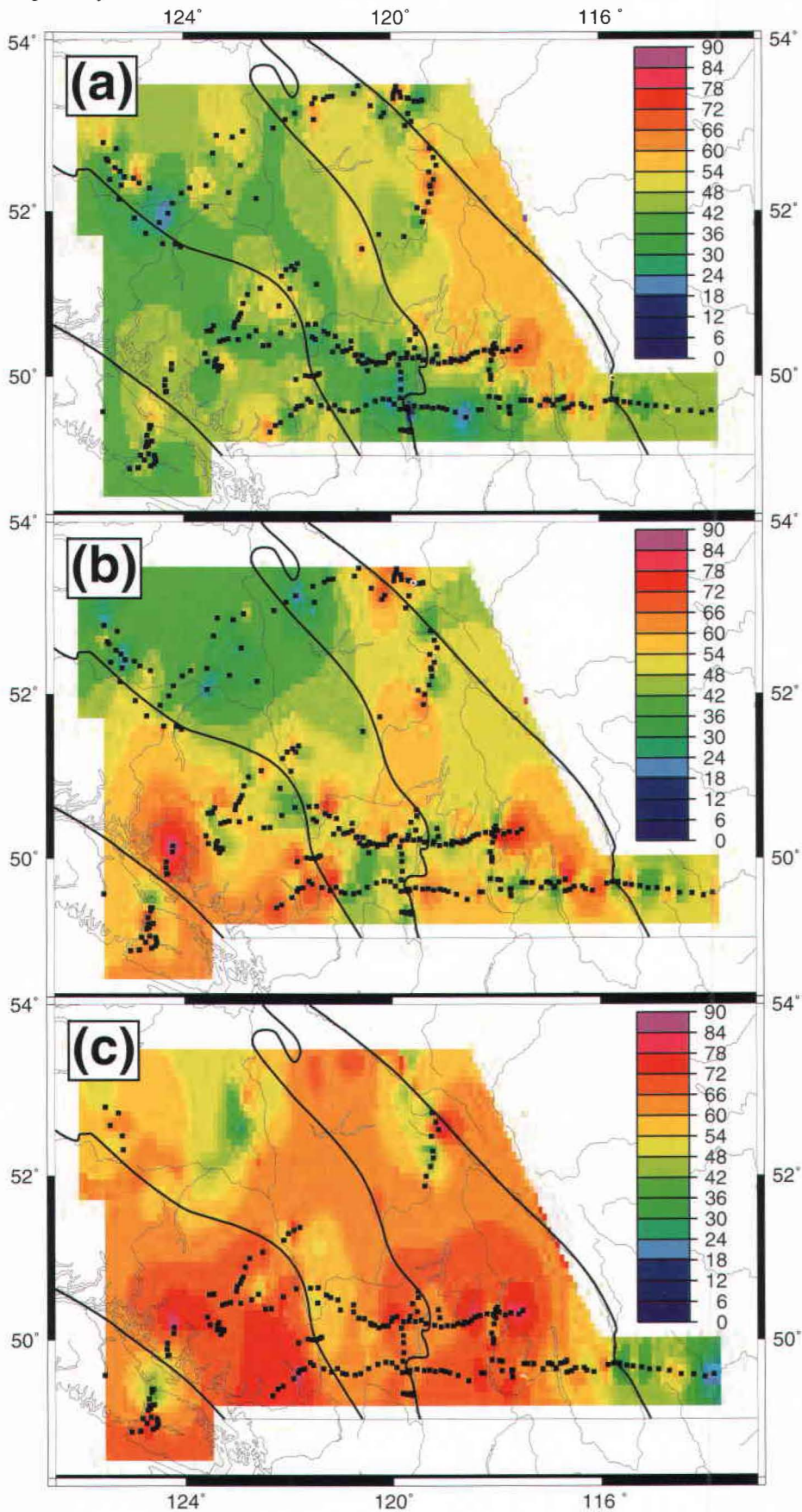
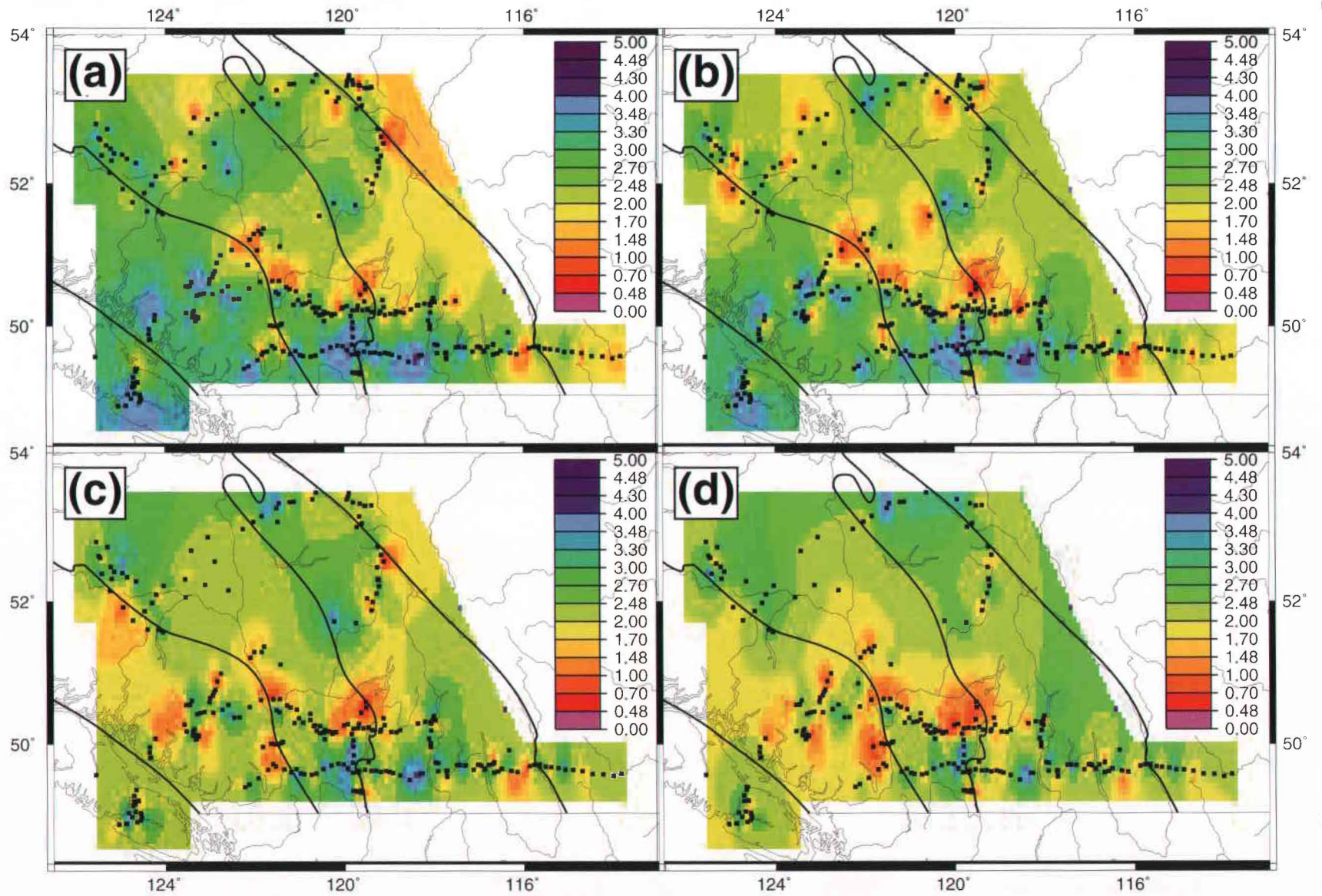


Fig. 4. Niblett–Bostick  $\log_{10}(\rho_{xy})$  maps for depths (a) 5 km, (b) 10 km, (c) 20 km, and (d) 30 km.



a model section of upper crustal conductivity on that line, Majorowicz and Gough (1994) noted the absence of a conductive middle crust there. Otherwise, the Intermontane Belt has a generally conductive lower crust (Fig. 3c). The phase map at 1 Hz (Fig. 3b), representing the middle crust, shows generally lower conductivities, with a large resistive block in the northwest corner.

High conductivities, indicated by high phase values, occur at frequencies of 0.1–0.01 Hz at the west ends of MT lines 2–5 (Figs. 2, 3c). These periods correspond to lower crustal depths. This highly conductive lower crust under the Coast Belt can probably be ascribed to silicate melt and (or) hot saline water. In the upper crust, very low phases, corresponding to highly resistive rocks, are seen in phase sections for lines 3–5 (Figs. 2, 3a). For line 5, the phase section suggests that the resistive Coast Belt upper crust extends eastward under a thin conductive skin, about halfway across the Intermontane Belt (Majorowicz and Gough 1991). Gough and Majorowicz (1992) consider that the resistive mass consists of granodiorites or similar acid crystalline rocks of the Coast Belt, that these are resistive because they have low fracture densities, and that they extend under a thin skin of basalt across the western half of the Intermontane Belt. These structures have been better defined in a 2D model (Majorowicz and Gough 1994). The phase map at 33 Hz (Fig. 3a) shows low phases (cold colours) along the Coast Belt, but interrupted by higher phases, which may indicate local hotspots in the melt under recently active volcanoes. The resistivity maps of Fig. 4, from Niblett–Bostick inversions, show very high resistivities, in the range 500–10 000  $\Omega \cdot m$ , through most of the Coast Belt at depths 5 and 10 km, but at 20 and 30 km, the highly conductive lower crust is reached, and has values in the range 5–300  $\Omega \cdot m$ . These low values are confirmed by the 2D modelling undertaken by Jones and Dumas (1993), which shows a lower crust of  $\sim 100 \Omega \cdot m$  within which is a highly conductive body, of  $\sim 10 \Omega \cdot m$ , interpreted to be a magma chamber.

### A major east–west-trending upper crustal discontinuity

A remarkable and unexpected result of this compilation of magnetotelluric data is the trace of a major crustal discontinuity running east–west between MT lines 2 and 3. The phase map at 33 Hz (Fig. 3) and the resistivity maps at depths 5 and 10 km (Fig. 4) all show an abrupt transition from high resistivities (500–10 000  $\Omega \cdot m$ ) south of this discontinuity, to much lower values (30–300  $\Omega \cdot m$ ) north of it. Two-dimensional models of the MT data from lines 2 and 3 confirm that the upper crust, to about 15 km, is more resistive on line 2 than it is on line 3 (Marquis et al. 1995). Two-dimensional modelling of the data from a line to the west of Okanagan Lake in the Intermontane Belt suggests that this north–south difference may persist even into the lower crust (Marquis et al. 1995).

This discontinuity cuts across the regional geological south-southeast–north-northwest trend. However, there is also a strong change in exposed rocks on either side of around 50°N in the compilation map of Wheeler and McFeely (1991). The rocks to the north are predominantly Eagle Bay clastics and volcanics of the pericratonic Kootenay Terrane, whereas to the south Jurassic plutonic rocks and early Terti-

ary granites (Ladybird suite, Coyell syenites) dominate. There are exposures of Jurassic plutons and Ladybird suite granites north of this line, but not of Coryell syenites.

The southern Cordillera seismic reflection transect of Lithoprobe crosses the Omineca and Intermontane belts along MT line 3, and so is on the less resistive upper crust. We note that the refraction model for refraction line 1 also displays north–south lateral variation in the upper crust, with higher velocity (6.35–6.40 km/s) upper crust beneath MT line 2 compared with lower velocity material (6.15–6.25 km/s) beneath MT line 3 (B.C. Zelt et al. 1992).

To substantiate this discontinuity in other geophysical parameters, we have mapped residual Bouguer gravity for the region, after removal of a quadratic regional trend (Fig. 5a). There is a step of about 20 mGal (1 mGal =  $10^{-3} \text{ cm/s}^2$ ) across the discontinuity, high on the north side. The aeromagnetic data for the region are shown in Fig. 5b, and a map of filtered aeromagnetic data for the region is shown in Fig. 5c. The filter used was a median filter with a wavelength of 10 km, so small features are suppressed. Once again, the discontinuity appears between MT lines 2 and 3 and gives a step, high to the south, of about 200 nT.

The upper crust is more magnetic, less dense, seismically faster, and more resistive south of the discontinuity, than the upper crust north of it. As noted earlier, one significant geological difference between line 2 and line 3 is the exposure of the Coryell syenites. If these syenites are more competent than the surrounding host rock, that might explain the higher resistivity. Syenites could have lower densities and higher magnetite content than the rocks north of the discontinuity, though it is more common to find higher densities associated with higher magnetization. Also, syenites have, on average, a somewhat higher compressional-wave velocity than do granites (Carmichael 1982).

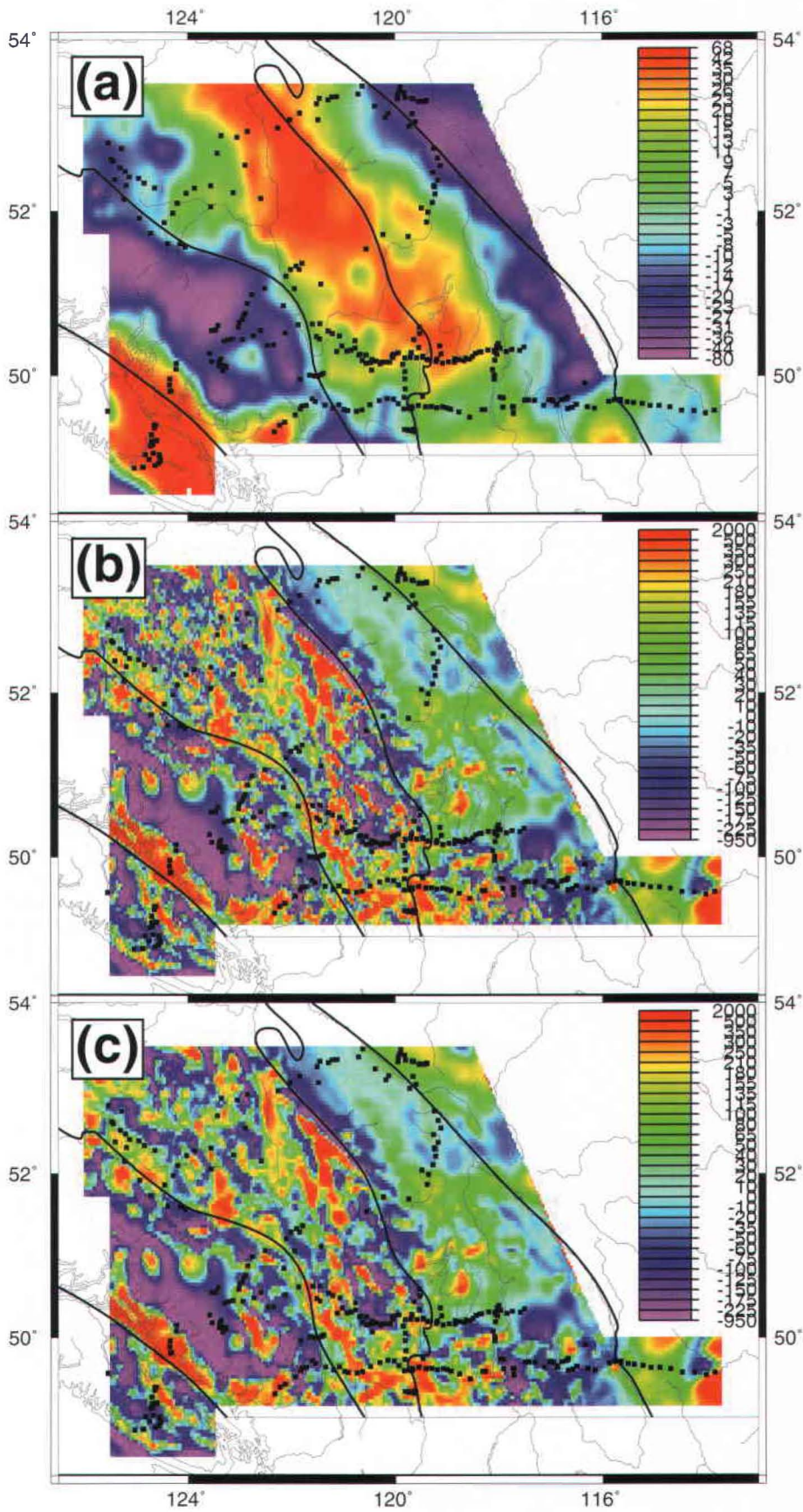
If the highly resistive crust to the south is imaging these intrusives, then we conclude that the syenites are not present in the tectonic stack to the north of about 50°N, in agreement with the surface exposures of the Coryell suite. Given that extension is observed as far north as 52°N, why are these syenites limited to the south? The answer may lie in the differences in extension: the syenites exist where extension exceeds  $\sim 50\%$  but not to the north where extension decreases. One counterargument to this proposal is that both the discontinuity and the northward cessation of exposure of syenite appear to be relatively sharp, whereas extension appears to decrease continuously farther north.

This major cross-strike discontinuity may be related to the cross-strike discontinuity near the Crownsnest Pass, reported by Price (1994), which he estimates to produce a right-hand offset in the cratonic margin of 200 km. While the upper crustal structure is far from two-dimensional, the conductivity, gravity, and magnetic maps suggest that the Lithoprobe seismic transect along MT line 3 may reasonably represent structure in the upper crust along strike to the northwest. Structures may be quite different along MT line 2.

### Imaging faults

The Canadian Cordillera contains a number of major faults crossed by the MT lines. From east to west: the northeastern part of line 5 crosses the Rocky Mountain Trench where it is close to the southern termination of the major Tintina –

**Fig. 5.** (a) Residual gravity map. (b) Aeromagnetic magnetic map. (c) Filtered aeromagnetic map.



northern Rocky Mountain Trench 2000 km long transform fault feature (Price and Carmichael 1986), rather than a valley controlled by normal faults, which it is to the south. The Eocene extensional Slocan Lake Fault is crossed by line 1. Between the Intermontane and Omineca belts is the Okanagan Valley Fault on lines 2 and 3, a regional extensional detachment of Eocene age. Perhaps the most dominant faults are those of the Fraser fault system, comprising the Late Eocene Fraser Fault itself, crossed by lines 2, 3, 5, and 6, and the Earlier, Mid Eocene, offset Yalakom–Hozameen Fault and Pasayten Fault.

#### Slocan Lake Fault

A surface dip of 30° to the east, with a dip-slip displacement of the order of 10–20 km, is attributed to the Slocan Lake Fault (Carr et al. 1987; Parrish et al. 1988). Most of the movement on it is dated at 59–46 Ma (Parrish et al. 1988). The Slocan Lake Fault passes beneath the Kootenay Arc at mid-crustal depths (~15 km), and interpretation of vertical-incidence seismic reflection data and wide-angle seismic refraction data from lines crossing Kootenay Lake and the Nelson batholith suggests that it extends throughout the whole crust, and perhaps projects to the position of Moho depth increase (Cook et al. 1987, 1988; C.A. Zelt and White 1995) from the Omineca Belt to the Foreland Belt. As stated above, in southeastern British Columbia the Kootenay Arc marks the eastern limit of Eocene magmatism, the eastern limit of exposed metamorphic core complexes, and the eastern limit of the allochthonous terranes. Geophysically, the Slocan Lake Fault demarks a lower crust of very low seismic velocity to the east, ~6.4 km/s, compared with a higher velocity to the west, ~6.7–6.8 km/s (C.A. Zelt and White 1995). These compressional-wave values immediately to the east of the trace of the Slocan Lake Fault in the lower crust are much lower than typical cratonic values of 6.7–7.5 km/s (Holbrook et al. 1991; Durrheim and Mooney 1994), and 7.1 km/s reported for the North American craton farther eastwards beneath Alberta (Kanasewich et al. 1994).

The MT data are unequivocal in their assertion of a significant change in conductivity on either side of the suggested lower crustal location of the fault (Fig. 2a). This concurs with observations since the 1960's that in southeastern British Columbia the Kootenay Arc is at the eastern limit of the Canadian Cordilleran Regional lower crustal conductor. Modelling in progress indicates that the lower crust beneath the Foreland Belt is in excess of 300  $\Omega \cdot m$  (Gupta and Jones<sup>3</sup>), and beneath the Purcell Anticlinorium, the lower crust is modelled at 100  $\Omega \cdot m$  (Jones et al. 1993) to 300  $\Omega \cdot m$  (Eisel and Bahr 1993). In comparison, west of the fault, beneath the Valhalla Gneiss Complex, the lower crust is of the order of 5  $\Omega \cdot m$  (Jones et al. 1988; Eisel and Bahr 1993) to 30  $\Omega \cdot m$  (Jones et al. 1993). Obviously, by geophysical measures, the lower crust changes dramatically in physical parameters at the Kootenay Arc – Slocan Lake Fault boundary.

#### Okanagan Valley Fault

The Okanagan Valley Fault (OVF) marks the boundary between the Intermontane and Omineca belts, and, from surface geological mapping and guided by the seismic reflection data, it dips shallowly westward. It is a detachment fault of

regional scale associated with the west-dipping phase of Eocene extension, dated at 50 Ma (Parrish et al. 1988). Based on the sections and maps, one could speculate about a resistivity variation associated with the Okanagan Valley Fault, but this is ill-defined. True-amplitude processing of the seismic data from the reflection lines crossing the Okanagan Valley Fault, together with inversion of the MT data for lines 2 and 3, shows that there is no significant change in either reflectivity or conductivity across the boundary (Marquis et al. 1995). Also, the geoelectrical strike direction remains the same across the fault at periods sampling the middle crust, and this direction is different from the direction of shorter period data sampling the upper crustal rocks. Accordingly, we must conclude that at mid-crustal depths the Okanagan Valley Fault has no perceptible conductivity enhancement and juxtaposes rocks of similar physical characteristics. Furthermore, this implies that the rocks beneath the Okanagan Valley Fault are of North American affinity (Marquis et al. 1995), rather than exotic allochthonous material as suggested in the interpretations of Cook et al. (1992).

However, there is a significant variation in lower crustal conductivity on either side of this boundary, with resistivities much higher beneath the Intermontane Belt (100–1000  $\Omega \cdot m$ ) than beneath the Omineca Belt (30  $\Omega \cdot m$ ) (Jones et al. 1992a, 1992b; Marquis et al. 1995). As with the Slocan Lake Fault above, this would argue against a simple geometric extension of unmodified North American basement all the way to the Fraser Fault.

#### Fraser Fault

The Fraser Fault (FF), known as the Straight Creek Fault in northern Washington State, is an important feature that lies within the eastern margin of the Coast Belt in the south, is at the boundary between the Intermontane and Coast belts farther to the north, and lies within the Intermontane Belt yet farther north. Price and Carmichael (1986) suggest that it is a component of a 2000 km long intracontinental transform fault that joins the Straight Creek fault to the northern Rocky Mountain Trench – Tintina fault system. The initial interpretation of the Fraser Fault from the vertical-incidence seismic reflection profiling preferred a listric geometry with the trace of the fault soling westwards into a mid-crustal zone of ductile shear (Varsek et al. 1993).

Magnetotelluric lines 2 and 3 cross the Fraser Fault between 121° and 122°W (Fig. 1b). The phase sections (Fig. 2) there show high phases, indicating high conductivities, through the upper crust to the surface. Additional short MT lines were placed to cross the Fraser Fault in latitudes 50 and 51.3°N and all four lines were used in a study of conductive structure in the fault zone (Jones et al. 1992b). Inversions show resistivities as low as 10–100  $\Omega \cdot m$  joining the lower crustal conductor west of the fault zone, under the Coast Belt, to the surface. The Fraser Fault has not been mapped geologically north of 52.1°N, but where MT line 5 crosses its northward extrapolation, near 52.7°N, 123°W, very low resistivities are again found from mid-crustal depths to the surface (Gough and Majorowicz 1992; Majorowicz and Gough 1994). Assuming that this marks the unmapped Fraser Fault there, a zone of high conductivities several kilometres wide has been observed at four crossings of this fault zone. We

have little doubt that the high conductivity is associated with fracturing, with either saline water or graphite or both in the fractures. Jones et al. (1992b) suggested, on the basis of the observed  $\delta D$  values and reduced  $\delta^{13}C$  ratios close to the fault (Nesbitt and Muehlenbachs 1991), that the enhanced conductivity at mid-crustal depths within the fault zone itself may be due to the emplacement of graphite or organic carbon deposited during the upwelling of deeply penetrating meteoric waters.

Also of note is that the MT data are unequivocal in showing that the Fraser Fault must penetrate the entire crust, rather than have the listric geometry preferred by the reflection interpretation. More recent processing of the seismic data, taking into account the crooked acquisition line, is now interpreted as being consistent with the MT interpretation (R.M. Clowes, personal communication, 1994).

### Causes of high conductivity in the Canadian Cordillera

The major silicates composing the Earth's crust have resistivities of order  $10^5 \Omega \cdot m$  or more when dry. Resistivities of order  $100 \Omega \cdot m$  are found at mid-crustal depths in continental shields, and in regions like the Canadian Cordillera, a zone of recent tectonic activity, values of  $1-10 \Omega \cdot m$  are observed. As the silicate grains are, for practical purposes, insulators, the bulk resistivity of the rock is controlled by conductive material in fractures and interstitial spaces between the grains. In the lower crust of old continental shields, graphite coatings on the silicate grains may be a major cause of enhanced conductivity. Such films have been detected in anorthosite from Wyoming (Frost et al. 1989), and in rocks from the Kapuskasing zone in Ontario (M. Mareschal et al. 1992). Calculations of bulk resistivity, using Hashin-Shtrikman bounds (Waff 1974) appropriate for a thin interconnected grain-boundary film of graphite, show that resistivities in the range  $1-100 \Omega \cdot m$  are theoretically possible (Frost et al. 1989), with values at the upper end, of some  $100s \Omega \cdot m$ , being suggested as likely given the observed film thicknesses (M. Mareschal et al. 1992; Katsube and Mareschal 1993). This hypothesis prompted Katsube and Mareschal (1993) to suggest that the continental lower crust is generally of much lower resistivity than expected from laboratory studies because of conducting pathways along grain-boundary films, but that these films become "broken" on uplift into the upper crust. Duba et al. (1994), using samples exhumed from deep in the upper crust as part of the German deep drilling (KTB) project, suggested that enhanced conductivity is likely to be due to a mixture of contributions from fluids that connect films or isolated conducting phases.

Mineral deposits such as graphite and sulphide ores are often highly conductive, a fact much used in exploration for metals. Gupta and Jones (1995) and Cook and Jones (1995) suggest that such mineralization could produce the observed high conductivities and contribute to some of the seismic reflectivity in the upper crust of the Purcell Anticlinorium. Boerner et al. (1995) suggest that a strong conductor in Alberta, which correlates spatially with the south Alberta - British Columbia (SABC) conductor and which is apparent in our maps (Figs. 4), is caused by graphitic metasedimentary rocks. It is unlikely, however, that the pervasive

enhanced conductivity in the lower crust beneath the whole of the Cordilleran region is due to mineralization.

The very low resistivities, of a few ohm metres, observed in a million or more cubic kilometres of the lower crust are far more probably associated with hot saline water and silicate melt in fractures and other interconnected spaces. This is not a new proposal; Caner (1970) suggested this as an explanation one quarter of a century ago, and many others have advanced it since. The Canadian Cordillera forms a region of new crust recently accreted from predominantly oceanic and island-arc material, with an active subduction offshore, and, possibly, a planetary-scale mantle upflow inland of the subduction (see below). In such a region, ionically conductive fluids are to be expected. Such fluids might include molten silicate mixtures such as basalt, water with salt and (or)  $CO_2$ , and pure water, which is a strong acid at lower crustal pressures and temperatures.

In many regions, the onset of increased conductivity, in the middle crust, appears to coincide with the appearance of high reflectivity in seismic reflection data (Gough 1986b; Jones 1987; Hyndman and Shearer 1989; Marquis and Hyndman 1992; Marquis et al. 1995), and with temperatures of order  $400-450^\circ C$  (Jones 1987, 1992; Hyndman and Lewis 1995) thought to represent the brittle-ductile transition. In particular, for the southern part of the region of study, Marquis et al. (1995) showed that reflectivity energy was much stronger in the lower crust than in the upper crust, and that this strong increase correlated with an increase in conductivity and with thermal models of the depth to the  $450^\circ C$  isotherm. This correlation of conductivity and reflectivity, together with the restriction of intraplate earthquakes to the upper crust in many continental areas, led Gough (1986b) to suggest that the change in conductivity might arise from a small content of water between the silicate grains, in separate inclusions in the upper crust but interconnected in the lower crust. This could produce rheological changes to cause reflective flow-like structures and restrict brittle failure, and earthquakes, to the upper crust. Bailey (1990) has modelled fluid residence times, and shown that the residence time of fluids in the crust is highly dependent on the ambient temperature. He suggests that fluids will rise rapidly to the brittle-ductile transition zone, and accumulate there in extensive reservoirs with high horizontal permeability but low vertical permeability. Jones (1987, 1992) proposed alternatively that the fluids are kept at depth because of the precipitation of minerals on the grain boundaries as the fluids cool below  $\sim 400^\circ C$ , which leads to the formation of an impermeable layer (Etheridge et al. 1983). The few boreholes that penetrate to these temperatures indeed show a zone of high pore fluid pressure beneath a zone chemically sealed by deposition of minerals in veins (Fournier 1991). Whichever effect is operating, there is a mechanism for keeping fluids at depth in the crust for reasonably long geological times. Evidence for fluids being retained at depth for times in excess of 70 Ma was presented by Goldfarb et al. (1991), who showed rapid, widespread fluid release at 56-55 Ma in southeastern Alaska from fluids interpreted to have been trapped in the deep crust during subduction of the Kula plate from 110 to 56 Ma. Trapped fluid, from the descending Juan de Fuca plate, has been imaged beneath Vancouver Island using both MT and seismic methods by Kurtz et al. (1986, 1990; see



also Jones 1987 and Hyndman 1988), Calvert and Clowes (1990), and Cassidy and Ellis (1991, 1993); and from MT alone, beneath the Coast Belt by Jones and Dumas (1993) and beneath the Coast Range of Oregon and Washington by Wannamaker et al. (1989).

If it is accepted that the lower crust of the Canadian Cordillera is highly conductive because it is wet, it follows that variations of fracture density with position must give rise to variations of conductivity. Majorowicz and Gough (1994), discussing conductive structures along MT line 5 (Fig. 1*b*), suggested that the high upper crustal resistivities in the Coast Belt result from low fracture density in these strong, acid crystalline rocks, and that the basalt of the Intermontane Belt was less resistive because it is more densely fractured. Structures in the maps of Figs. 3 and 4 and the phase–frequency sections of Fig. 2 may be related to variations in fracture density in the crust, in addition to temperature and nonuniform sources of fluids from below. The low resistivities, rising from the middle crust to the surface along the Fraser Fault, are probably related to high fracture densities in the fault zone. While water is probably present in the fault zone, graphite may also be present (see above).

### **Mantle upflow beneath the southern Omineca Belt: active or passive?**

As shown above, although there are some lateral variations, much of the crust of the southern Canadian Cordillera, from the Kootenay Arc in the east to the Fraser Fault, exhibits low electrical resistivity (high conductivity). The lower crust shows minimum values ( $<30 \Omega \cdot \text{m}$ ) beneath the eastern part of the Omineca belt, and, following Lewis (1990) and Lewis et al. (1992), who presented a geothermal model for the region with  $730^\circ\text{C}$  being reached at 21 km, Jones (1992) and Jones et al. (1992*a*) suggested that the enhanced conductivity is due to the lower crust there being partially molten.

The crust is also thin in the southern Omineca and Intermontane belts, with a virtually flat Moho at  $\sim 32$  km. This is anomalous compared with global averages of 50 and 42 km for young and older orogens, respectively (Christensen and Mooney 1995). Furthermore, both deep seismic refraction experiments and teleseismic studies have shown since the 1960's that the mantle lithosphere is also anomalously thin and slow (W.R.H. White and Savage 1965; Kanasewich 1966; Wickens 1971; Fulton and Walcott 1975), and recent rheological modelling places the lithosphere–asthenosphere boundary at 40 km beneath the Omineca Belt, compared with 130 km beneath the Foreland Belt (Lowe and Ranalli 1993). The seismicity associated with the offshore subduction is restricted to the Insular and Coast belts, and to depths less than 40 km (Milne et al. 1978). There is no currently active Benioff zone beneath the Intermontane and Omineca belts recognized from seismicity data.

These facts, together with the crustal extension and basaltic extrusions of Tertiary time, led Gough (1986*a*) to propose that an elongated active upflow in the mantle now lies beneath the Intermontane and Omineca belts, behind the offshore subduction, in a northward continuation of a larger upflow beneath the Basin and Range tectonic province of the U.S.A. (Gough 1984). Gough (1984) suggested that the cause of this upflow is an upper mantle convection cell,

which had its rising edge beneath the East Pacific Rise during the Tertiary, but which has since moved eastwards, relative to North America, and now lies beneath the Basin and Range and extends into southern British Columbia. Wilson (1991) supports Gough's general thesis, but postulates rather than the upflow is caused by North America overriding a mid-ocean ridge associated with the now-vanished Kula plate. These proposals are variations on Menard's (1960) original suggestion that the East Pacific Rise extends beneath western North America. Menard thought that the East Pacific Rise extended from the Gulf of California northwestwards, exiting at about the middle of the state of Oregon; that is, he connected the East Pacific Rise with the Juan de Fuca ridge. Kanasewich (1966) also suggested, on the basis of low sub-Moho P-wave velocities (7.7–8.0 km/s) compared with crustal values (8.1 km/s), that the East Pacific Rise currently extends beneath western North America.

Could an upwelling associated with the ridge of an upper mantle convection cell be the cause of the Eocene extension event? As discussed above, the onset of extension was marked by emplacement at high levels in the crust of anatectic leucogranites of the Ladybird suite. Crustal-derived anatectic magmas are atypical of both back-arc and magma plume settings (Bardoux and Mareschal 1994), and would argue against a causative connection between an active mantle event and the onset of extension. In addition, whereas the extension in the Basin and Range is still taking place today, the extension event in the southern Omineca Belt was very short-lived, only  $\sim 10$  Ma, from  $\sim 59$  to  $\sim 46$  Ma (Bardoux and Mareschal 1994). We note that Engerbretson et al.'s (1985) plate tectonic reconstruction has the Yellowstone hot spot just on the paleocoast at the peak of magmatism in the southern Canadian Cordillera ( $\sim 50$  Ma, Armstrong and Ward 1991), and that the spreading centre between the Farallon and Pacific plates was well offshore at this time. The nature of the boundary between the Farallon and Kula plates is still unknown, but if it were a spreading centre, then North America would have begun to override it at about 60–80 Ma, and its trace would place it beneath the southern Omineca Belt at about the period of the Kamloops–Challis–Absaroka magmatic episode.

Alternatively, did extension start as a consequence of mantle upwelling caused by a passive event, namely, the detachment of overthickened mantle lithosphere at the culmination of compression from the docking of the Insular Superterrane (Ranalli et al. 1989)? Whether the resulting thinned lithosphere observed today is a consequence of a passive or an active dynamic mechanism is still a debatable issue, but numerical modelling suggests that removal of the mantle lithosphere will always trigger extension (J.C. Mareschal 1994).

Others have suggested that extension was a result of plate reorganization within the Pacific at around 55 Ma (Bardoux and Mareschal 1994). While this may be an attractive theory, it implies that plate kinematics, rather than mantle dynamics, are the driving force. This is clearly untenable – the plates respond to mantle dynamics, not the other way around!

### **Conclusions**

We have compiled MT data from the largest number of sites

ever recorded for the study of accretionary tectonics, and shown the lateral and depth variation of MT parameters within the southern and central Canadian Cordillera. From these sections and maps, we have drawn mainly qualitative conclusions that support prior observations, dating back to the early 1960's, and previous reports of analyses of subsets of our data. The bulk of these MT data were acquired as part of the Lithoprobe southern Cordilleran transect activities as a complement to the vertical-incidence seismic reflection and the seismic refraction studies. However, whereas in the main the interpretations of the reflection data have emphasized structural elements, mostly in two dimensions, the interpretations of the MT data have highlighted the current physical state of the crust by mapping the brittle–ductile transition as manifest by the rapid increase in conductivity within the middle crust. Also, the MT data testify that the simple 2D structure implied by the morphogeological belts belie actual lateral and vertical variations in three dimensions — the strike variation can be as significant as the variation across strike.

Regionally, much of the crust of the Canadian Cordillera exhibits electrical conductivities two to four orders of magnitude greater than what one would expect from dry, competent rocks, which make up the crust. Volumetrically, the conductive region we have mapped, named the Canadian Cordilleran Regional conductor by Gough (1986a), is in excess of one million cubic kilometres. The depth to its top varies laterally; it is shallower beneath the Omineca Belt than beneath the Intermontane Belt. This contrast correlates with lateral variations in the depth to the 450°C isotherm, and to the top of high-amplitude reflectivity. Internally, the electrical conductivity of the Canadian Cordilleran Regional conductor varies, principally with belt boundaries.

We interpret the Canadian Cordilleran Regional conductor as saline fluids trapped at depth below the brittle–ductile transition, and that internal conductivity variations are a consequence of fracture density variation. There are three potential sources of these fluids: (1) metamorphic fluids released by devolatilization reactions as the Juan de Fuca plate subducts beneath Vancouver Island (Kurtz et al. 1986, 1990; Hyndman 1988) and the western Coast Belt (Jones and Dumas 1993); (2) deeply penetrating surface-derived meteoric waters in the Intermontane and Omineca belts (Nesbitt and Muehlenbachs 1991), and (3) fluids coming into the crust from the mantle (Beaudoin et al. 1991).

Superimposed on the regionally extensive Canadian Cordilleran Regional conductor are local anomalies due to either partial melt or mineralization, either graphitic or sulphitic. A magma chamber at 10–12 km was imaged beneath the Garibaldi volcanic zone (Jones and Dumas 1993). The lowermost crust beneath the Valhalla Complex in the southern part of the Omineca Belt exhibits high conductivity (low resistivity of  $\sim 5 \Omega \cdot \text{m}$ , Jones et al. 1988), and temperatures in excess of 730°C are inferred on the basis of the very high heat flow observed (Hyndman and Lewis 1995).

Recent observations in Alberta by Boerner et al. (1995) show that the southern Alberta – British Columbia conductor is in the upper crust, and those authors favour an interpretation in terms of graphitic metasediments. Highly conducting sulphides observed in a drill hole in the Purcell Anticlinorium (Cook and Jones 1995) give proof that the

enhanced conductivity observed in the upper crust in south-eastern British Columbia and northwestern Montana (triangles in Fig. 1a) are associated with electronic conduction, rather than ionic conduction in fluids (Gupta and Jones 1995). A narrow mid-crustal vertical conductor imaged within the Fraser fault zone was interpreted as graphitic (Jones et al. 1992b), although fluids have been proposed here also as an explanation (Gough and Majorowicz 1992; Majorowicz and Gough 1994).

Some of the results from these MT investigations also have relevance to structural studies. From the reflection data, the Fraser Fault was suggested to have a westward listric geometry into a mid-crustal shear zone. The MT data, and their modelling, are unequivocal in their imaging of a subvertical crustal-penetrating feature. It is clear that MT data are complementary to seismic reflection data in that they can image vertical features much more robustly (Jones 1992).

The location of the edge of North American cratonic basement is a fundamental question. While deductive arguments based on isotope ratios and palinspastic reconstruction geometries for the continuation of North American basement west of the Kootenay Arc – Slocan Lake Fault to at least Penticton, if not to the Fraser Fault, may be convincing, what cannot be denied is that geophysical data, seismic P-wave velocities, heat flow and electrical conductivity show that the state and nature of the lower crust is fundamentally different on either side of the Kootenay Arc – Slocan Lake Fault. This difference may be a consequence of magmatic underplating since the Jurassic (Armstrong and Ward 1991, 1993), or that cratonic material to the west is highly modified with respect to the east by compressive heating and subsequent extension. From seismic refraction velocity evidence, C.A. Zelt and White (1995) interpret the observed transition in lower crustal velocities across this boundary as North American craton being attenuated and modified (injected with mafic intrusives) to the west during Middle and Late Proterozoic rifting. Alternatively, perhaps the Intermontane Superterrane lay on top of its own Archean-aged cratonic fragment, distinct from North American craton, and that our seismic refraction and electrical observations indicate differences between cratons. As an ancient analogue, a fragment of Archean crust is conjectured to lie within the Paleoproterozoic Trans-Hudson Orogen between the Superior craton to the east and cratons to the west (Wyoming and Rae–Hearne) (Nelson et al. 1993; Lucas et al. 1993). However, there is no geological evidence for such an exotic fragment.

One new result from this compilation is the recognition of a major east–west geophysical boundary within the southern Intermontane and Omineca belts at about 50°N. To the south of this boundary, the upper crust is highly resistive, faster, less dense, and contains more magnetic material, compared with the upper crust farther north. This boundary is also strongly apparent in the compilation map of Wheeler and McFeely (1991), where it demarks the northern limit of exposures of the Coryell syenite suite. These geophysical and geological observations are presumably related in some manner. Competent syenitic plutons could be highly resistive and have lower density, but it is difficult to explain the observed higher velocities and the higher magnetism. A basic question that needs to be addressed concerning these syenites is why they are not emplaced in the crust farther

north. Coeval extension continued to Mica Dam at 52°N.

With regard to the cause for the onset of extension, one must keep in mind that the seismic and electromagnetic observations are responding to the state of the lithosphere that exists today, and that the heat flow on the surface of the Earth is giving information about the thermal state at the top of the mantle lithosphere ~10 Ma ago. In contrast, geology and geochemistry are reporting events that occurred in the past. The challenge is to bring these observations together in order to understand the complete dynamic, kinematic, and thermal history of the southern Canadian Cordillera. It is quite clear that the crust and lithosphere of western North America are highly anomalous at the present time — they are both hot and both thin. Thus, we surmise that there currently exists an upwelling of the mantle beneath this region. This hypothesis is not new, it was advanced as early as the mid-1960's by W.R.H. White and Savage (1965) to explain slow mantle seismic velocity observation in western North America. What is still a point of debate is whether this upwelling is responding to an active or a passive process. Some preferentially associate this upwelling with the East Pacific Rise (Gough 1986a) or an overridden paleoridge of the Kula plate (Wilson 1991), whereas others prefer passive detachment of overthickened mantle lithosphere (Ranalli et al. 1989). Whether this upwelling was responsible for the initiation of the Eocene extension, or whether the extension is associated with a possible spreading centre between the Kula and Farallon plates overridden by the North American craton (Wilson 1991) or with the reorganization of the plates that took place in the Pacific (Bardoux and Mareschal 1994) at about 55 Ma after the collision of India with Asia are questions that remain to be resolved. Coupled with this is the question why extension ceased after only 10 Ma in the southern Canadian Cordillera, whereas in comparison it is still taking place in the Basin and Range today.

Clearly, interpretations of this vast amount of magnetotelluric data have not only added considerably to the body of knowledge about the southern Canadian Cordillera but have also raised questions about some of the generally accepted models based on surface geological mapping coupled with seismic reflection geometries.

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