Highly conductive horizons in the Mesoproterozoic Belt-Purcell Basin: Sulfidic early basin strata as key markers of Cordilleran shortening and Eocene extension

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ABSTRACT

We investigated the crustal structure of the central Mesoproterozoic Belt Basin in northwestern Montana and northern Idaho using a crustal resistivity section derived from a transect of new short- and long-period magnetotelluric (MT) stations. Two- and three-dimensional resistivity models were generated from these data in combination with data collected previously along three parallel short-period MT profiles and from EarthScope MT stations. The models were interpreted together with coincident deep seismic-reflection data collected during the Consortium for Continental Reflection Profiling (COCORP) program. The upper-crustal portion of the resistivity model correlates well with the mapped surface geology and reveals a three-layer resistivity stratigraphy, best expressed beneath the axis of the Libby syncline. Prominent features in the resistivity models are thick conductive horizons that serve as markers in reconstructing the disrupted basin stratigraphy. The uppermost unit (up to 5 km thick), consisting of all of the Belt Supergroup above the Prichard Formation, is highly resistive (1000–10,000 Ω m) and has relatively low seismic layer velocities. The intermediate unit (up to 7 km thick) consists of the exposed Prichard Formation and 3+ km of stratigraphy below the deepest stratigraphic exposures of the unit. The intermediate unit has low to moderate resistivity (30–200 Ω m), relatively high seismic velocities, and high seismic reflectivity, with the latter two characteristics resulting from an abundance of thick syndepositional mafic sills. The lowest unit (5–10 km thick) is nowhere exposed but underlies the intermediate unit and has very high conductivity (4–8 Ω m) and intermediate seismic velocities. This 17–22-km-thick three-layer stratigraphy is repeated below the Libby syncline, with a base at ~37 km depth. Seismic layer velocities indicate high mantle-like velocities below 37 km beneath the Libby syncline. The continuous high-conductivity layer in the lower repeated section

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INTRODUCTION

The Cordilleran thrust belt in the southern Canadian Rockies and Glacier National Park of the northern U.S. Rockies (Fig. 1) is a classic area for viewing the structure of a foreland fold-and-thrust belt (Bally et al., 1966; Dahlstrom, 1969, 1970; Price, 1981; Boyer and Elliott, 1982). These and other studies have documented significant Jurassic to Paleocene shortening (50–150 km) of Phanerozoic strata in the frontal ranges of the eastward-verging thrust belt above a gradually westward-deepening basal décollement that occurs at or above the contact with basal Precambrian crystalline basement. Additional, but less well-constrained, shortening occurs west of the frontal ranges, where a thick sequence of Mesoproterozoic sedimentary rocks (Belt and equivalent Purcell Supergroups) occurs structurally above the imbricated Phanerozoic strata along the Lewis thrust. Spectacular exposures for over 200 km along strike of the Lewis thrust in Glacier and Waterton National Parks (Willis, 1902; Dahlstrom, 1970; Mudge and Earheart, 1980; Boyer and Elliott, 1982; Fermor and Price, 1987; Yin et al., 1989; Fermor and Moffat, 1992; Whipple, 1992) have provided clear, well-exposed examples of thrust belt geometries over a broad area of the foreland region.

The fate of the Lewis thrust as it deepens farther west and presumably joins the westward-deepening basal décollement of the Phanerozoic front ranges was beyond the reach of early industry seismic-reflection data (Bally et al., 1966), but it has since been pursued by government-funded deep seismic-reflection programs in the United States (Consortium for Continental Reflection Profiling [COCORP]; Potter et al., 1986; Yoos et al., 1991) and in Canada (LITHOPROBE; Cook et al., 1992; van der Velden and Cook, 1994; Cook and van der Velden, 1995). These studies have had variable success in tracking the basal thrust belt décollement into the deeper crust to the west and in elucidating overlying thrust belt geometries. Estimates of total shortening from these and other studies of the entire width of the thrust belt near the international border are variable, ranging from 100 km (Price and Sears, 2000) to over 200 km (Price, 1981), with most estimates being somewhere between.

Farther west, in the hinterland of the Jurassic to Paleocene Cordilleran thrust belt, exposures of midcrustal metamorphic core complexes in the United States and Canada, which were uplifted in early Eocene time, are bounded to the east by Eocene east-dipping low-angle normal faults. Significant displacement (tens of kilometers) on the bounding low-angle normal faults is required to account for uplift and exposure of these midcrustal rocks. These eastward-deepening faults must displace features within the Cordilleran thrust belt, but they have not been consistently imaged in the seismic-reflection data (Yoos et al., 1991; Cook and van der Velden, 1995).

In order to provide new information relevant to the crustal structure in the hinterland of the Cordilleran thrust belt, we imaged the crustal conductivity structure along a 143 km transect from just east of the Rocky Mountain Trench near Whitefish, Montana, to the Priest River metamorphic core complex (herein Priest River complex) near Bonners Ferry, Idaho (Figs. 1 and 2). Parts of this transect coincide with legs of the Idaho-Montana COCORP deep seismic-reflection profile (Yoos et al., 1991). We deployed paired long- and short-period magnetotelluric (MT) instruments at 34 stations spaced roughly 5 km apart along the transect that allowed us to generate a continuous model of electrical resistivity along the transect down to a depth of ~50 km. We combined that data with seismic-reflection profiles and seismic interval velocities derived from the COCORP data, as well as preexisting MT profiles to the north and south and recent EarthScope MT data. We used this new data set to model and interpret the two- and three-dimensional crustal resistivity structure over
Figure 1. Geological map of Mesoproterozoic strata of the Belt-Purcell Basin, showing the locations of Figures 2 and 9. Locations of two major sedimentary exhalative massive sulfide deposits (Sullivan and Sheep Creek) within the Prichard and equivalent Aldridge Formations are shown, along with two deep drill holes. Figure is modified from Box et al. (2012). ALB—Alberta; BC—British Columbia; ID—Idaho; MT—Montana; OR—Oregon; WA—Washington; WY—Wyoming.
Figure 2. Location of magnetotelluric (MT) stations from this study shown on a base of (A) geologic map (simplified from Miller et al., 1999; Harrison et al., 1992; Harrison and Cressman, 1993), and (B) isostatic residual gravity map (from Mankinen et al., 2004), shown along with MT stations from Phoenix USA (KCU-1, KCU-5, and KCU-9 from Gupta and Jones, 1995) and from EarthScope (http://www.earthscope.org), and seismic reflection lines ID-2 and MT-2 from the COCORP program (Yoos et al., 1991), and line WTF-82-1 from Signal of Montana (Yoos et al., 1991). Also shown is the location of the deep oil exploratory well Gibbs #1 (Boberg et al., 1989). COCORP—Consortium for Continental Reflection Profiling.
REGIONAL SETTING

Belt Supergroup

The transect area is underlain primarily by folded and faulted strata of the Mesoproterozoic Belt (Purcell in Canada) Supergroup, fill of the Belt (or Belt-Purcell) basin (Figs. 1 and 2). The Belt Supergroup has an estimated stratigraphic thickness of between 15 and 20 km in the central part of its exposure belt (Cressman, 1989; Lydon, 2000; Hoy et al., 2000), including 5–7 km of section below the stratigraphically deepest exposures, which have only been imaged seismically (Cook and van der Velden, 1995). The basin is considered to have developed in an aborted intracontinental rift setting (Chandler, 2000), with basin fill deposited between 1500 and 1400 Ma (Anderson and Davis, 1995; Evans et al., 2000; Lydon, 2000). Basin strata (from bottom upward) consist of (1) the basal 5–7-km-thick unit of unknown lithology that is seismically imaged as weakly reflective, (2) the Prichard Formation, a southwest-derived, deep-water, generally argillaceous turbidite section with locally abundant syndepositional mafic sills (5–7 km thick) that produce prominent seismic reflectivity, (3) the Ravalli Group, a southwest-derived fluvial-deltaic complex (2.5 km), (4) the Piegan Group, a cyclic carbonate and siliciclastic sequence of either lacustrine or marine setting (formerly the middle Belt carbonate) capped by a sequence of mafic to felsic lavas (2 km), and (5) the Missoula Group (2.5 km), a south- and east-derived fluvial succession (Fig. 1; Winston and Link, 1993; Link et al., 2007; Winston, 2007; for correlation of the U.S. unit nomenclature with Canadian unit nomenclature to the north across the border, see Box et al., 2012). The Prichard Formation argillaceous strata are characterized by abundant carbonaceous seams and by abundant iron sulfide (Huebschman, 1973; Cressman, 1989), which can result in higher electrical conductivities if physically interconnected. Typical thicknesses indicated for these units are quite variable, owing in part to internal thickness variations across the geographic outcrop belt as well as unseen structural thickening or thinning. The depositional base of the Belt Supergroup is only exposed along the eastern flank of the basin in central Montana. There, the thin Neihart Quartzite rests nonconformably on Paleoproterozoic basement and grades upward through shelf and slope strata into a greatly compressed section of Prichard Formation–equivalent strata, with the entire Belt section ~5 km thick.

The Belt Supergroup experienced a number of deformational and thermal episodes through its long postdepositional history. Deformation, magmatism, and metamorphism have been documented at 1.37 Ga ~120 km to the northwest of our transect in British Columbia (East Kootenay orogeny of McMechan and Price, 1982) and in east-central Idaho, ~350 km to the southeast (Doughty and Chamberlain, 1996). Although McMechan and Price (1982) suggested the East Kootenay orogeny was compressive in nature, Anderson and Parrish (2000) suggested it may have been an extensional event akin to that in east-central Idaho documented by Doughty and Chamberlain (1996). A widespread static thermal event between 1.05 and 1.15 Ga has been documented 100 km to the north near the Sullivan mine (Anderson and Parrish, 2000), within the Priest River complex in northern Idaho (Doughty and Chamberlain, 2008), and ~170 km to the south in north-central Idaho (Zirakparvar et al., 2010). Multiple late Neoproterozoic to Late Cambrian mafic alkalic intrusive events (780, 660, and 500 Ma) are recorded by small intrusive bodies throughout the Belt Supergroup outcrop area, and these are associated with the late Neoproterozoic rifting of western North America (Cecile et al., 1997; Harlan et al., 2003; Lund et al., 2010). This was followed by deposition of a thin (<2 km) sequence of Paleozoic carbonate and elastic rocks. We infer that any deformation of Belt strata in the study area prior to Paleozoic deposition was local and minor, as Cambrian strata are only known to depositionally overlie the youngest formation in the Belt Supergroup (Harrison et al., 1992), indicating little if any folding and erosion prior to Paleozoic deposition. Significant deformation of Belt strata occurred during Late Jurassic–Paleocene activity of the Cordilleran thrust belt and during major Eocene extensional faulting. Granitic plutonism occurred locally in Cretaceous and Eocene time.

Structural Framework from Surface Geology

Northwestern Montana and northern Idaho are part of the Sevier fold-and-thrust belt (DeCelles, 2004; Dickinson, 2004; Burchfiel et al., 1992), the eastern front of the Late Jurassic–Paleocene part of the Cordilleran thrust belt that extends from Canada to Mexico. This reach of the Sevier belt differs from other parts of the belt to the north and south by the presence of the thick Mesoproterozoic Belt strata in the hanging wall west of the front ranges of the belt. The Sevier belt here consists of the frontal “Montana disturbed belt” (Mudge, 1970) of imbricated Paleozoic and Mesozoic strata, structurally overlain by the Lewis and associated thrusts with Proterozoic strata in their hanging walls. From east to west, these thrusts are the Hefty, Whitefish, Pinkham, Libby, and Moyie thrust systems (Fig. 2; Harrison et al., 1992; Harrison and Cressman, 1993; DeCelles, 2004). These thrust systems and intervening major folds are discussed in the following.

Regional outcrop patterns (Fig. 2) are controlled by broad north-northwest–trending folds. From east to west, these are the synclinal Rocky Mountain Trench, the Purcell anticlinorium, the Libby syncline, and the Sylvanite anticline. Most workers interpret folding to be thrust-related (Yoos et al., 1991; Cook and van der Velden, 1995; DeCelles, 2004). However Harrison et al. (1992) and Harrison and Cressman (1993) suggested that the
major folds are Precambrian in age and are unrelated to, and cut by, Cordilleran thrusts. Precambrian folding with slaty cleavage has been documented near the Sullivan Pb-Zn-Ag deposit (Leech, 1962; McMechan and Price, 1982), and an unconformity between Belt and Cambrian strata in western Montana is locally observable (Deiss, 1935). However, in the area of Figure 2, all of the very small occurrences of Cambrian strata rest on the stratigraphically highest formation of the Missoula Group (Harrison et al., 1992), suggesting to us that any folding and erosion prior to Cambrian deposition was minor and is not reflected in the strongly folded outcrop pattern.

Beginning on the east of our transect (Fig. 2A), the Whitefish Range consists of middle and upper Belt Supergroup units (Ravalli, Piegan, and Missoula Groups), cut by several southwest-dipping thrust faults (Harrison et al., 1992). The gently southwest-dipping Hefty thrust system occurs on the east flank of the Whitefish Range near the Canadian border, placing Missoula Group over Paleozoic carbonates and quartzose sandstones. Fritts and Klipping (1987a, 1987b) used seismic data to image other southwest-dipping thrust systems of lesser stratigraphic offset within the Belt Supergroup (Wigwam, Whitefish, Ten Lakes thrust faults, simplified in Figure 2 as Whitefish thrust) that are mapped within the western Whitefish Range. The western flank of the Whitefish Range is marked by a mostly infilled set of down-to-the-west Tertiary normal faults under Quaternary cover of the Rocky Mountain Trench, though inferred stratigraphic offsets are small.

The Rocky Mountain Trench is a broad (15-km-wide) topographic low exposing mostly Quaternary glacial gravels with scattered bedrock knobs of Belt strata. The trench is manifest as a discontinuous, possibly en-echelon series of Tertiary half grabens above a discontinuous set of down-to-the-west listric normal faults. Across the border in Canada, the Rocky Mountain Trench fault has up to 10 km of displacement (van der Velden and Cook, 1994), while to the south near Whitefish (Fig. 2), a well with 500 m of Tertiary sediment suggests a correlatable fault with ~3 km of displacement (Fritts and Klipping, 1987a, 1987b). However, no Tertiary sediment is known at the latitude of our MT transect, nor does the geologic mapping suggest significant normal displacement of stratigraphy. A gentle syncline of Piegan Group is exposed in the western part of the Rocky Mountain Trench.

The southwest-dipping Pinkham thrust separates exposures of Prichard Formation and Ravalli Group strata within the Purcell anticlinorium to the west from Piegan Group exposures in the Rocky Mountain Trench. The Gibbs #1 well (Fig. 2) was drilled into the crest of the Purcell anticlinorium to test for possible petroleum-bearing Paleozoic strata beneath the Pinkham thrust. The well is interpreted to have penetrated the gently dipping Pinkham thrust at ~5.2 km below the surface (calculated dip = 18°), but Belt Supergroup quartzite, not Paleozoic strata, underlie the thrust (Boberg et al., 1989). The top of the anticlinorium is a very gentle and broad structure. However, the western flank of the Purcell anticlinorium steepens sharply into the Libby syncline.

The Libby syncline is narrow and steep sided, with maximum dips greater than 70° on each flank. Harrison et al. (1992) and Harrison and Cressman (1993) showed several west-dipping thrust faults within the axial portion of the Libby syncline, which are simplified in Figure 2. As mapped, the thrust faults have limited stratigraphic throw, although south of our transect, Cambrian strata depositionaly overlying upper Missoula Group are preserved beneath upper Missoula Group in the footwall of the Libby thrust. West of the town of Libby, the east-dipping, west-directed Snowshoe thrust has a displacement of ~4 km (Fillipone and Yin, 1994). Fillipone and Yin suggested that the Snowshoe thrust is most likely a back thrust within the upper plate of the Libby thrust.

The Sylvanite anticline (Movie anticline in Canada) is a broad, doubly plunging anticline (northern nose in Canada) with a wide gentle crest and sharp, steeply dipping flanks. The gentle crest of the anticline exposes Prichard Formation, while the more steeply dipping flanks expose strata from higher in the Belt Supergroup stratigraphy. A well on the crest of the anticline in Canada (Duncan Energy Movie #1) penetrated 3500 m of Prichard Formation strata heavily intruded by basaltic sills. The parallelism between the axis of the Sylvanite anticline and the Libby thrust system suggests it is a hanging-wall anticline developed during east-directed thrust motion.

The Movie thrust is a west-dipping, east-directed thrust with Prichard Formation in its hanging wall and upper Missoula Group in its footwall, which represents the greatest stratigraphic separation of any thrust fault in the area of Figure 2. At the latitude of the MT transect, the footwall strata dip steeply (70°−80°) under the fault, while the hanging-wall strata dip steeply toward the fault. As the Movie thrust is followed south of the MT transect, it bends to the west before it intersects the subvertical right-lateral Hope fault. In that bend area, a north-northwest−trending syncline of Piegan and Missoula Groups projects under the Movie thrust, while a north-northeast−trending syncline parallels the fault in the upper plate.

The hanging wall of the Movie thrust consists of Prichard Formation riddled with mafic sills, dipping moderately eastward toward the thrust (Miller and Burmester, 2004). It is intruded by mid-Cretaceous granitic plutons of the Kaniksu batholith (Gaschnig et al., 2012), with U-Pb zircon ages of 105–120 Ma and biotite and hornblende K-Ar (cooling) ages of 90–100 Ma. These plutons predate at least some of the displacement on the Movie thrust, indicated by 70 Ma mylonite formation along the fault (Fillipone and Yin, 1994).

The only major Eocene and younger normal fault exposed within the MT transect area is the east-dipping Purcell Trench detachment fault (Rhodes and Hyndman, 1984; Rehrig et al., 1987; Harms and Price, 1992; Doughty and Price, 1999, 2000), bounding the east side of the Eocene Priest River metamorphic core complex. The Priest River complex consists of interleaved foliated Cretaceous granitic rocks and Belt metasedimentary rocks. Peak metamorphism occurred at 15–30 km depths at ca. 72 Ma (Doughty et al., 1998). The fault marks a sharp break to high
metamorphic grade and younger metamorphic cooling ages in the footwall below the fault (Miller and Engels, 1975; Doughty and Price, 1999, 2000). The actual fault trace is mostly hidden within a wide valley beneath Quaternary glacial deposits. Doughty and Price (2000) described an exposure of the fault just north of our transect, consisting of mylonitic rocks dipping 57° to the east and capped by a 3–6 m zone of chloritic microbreccia. Normal offset across the combined Purcell Trench–Newport fault system is estimated at 35–68 km (Harms and Price, 1992), while Doughty and Price (1999) estimated 66 km of displacement across the southern Purcell Trench.

The Priest River complex has exposures of Archean orthogneiss ~45 km to the south-southwest of the westernmost MT station in our profile (Doughty and Price, 1999). Those Archean rocks occur beneath the gently north-dipping Spokane Dome mylonite zone, which is overlain by the Selkirk Crest portion of the Priest River complex. The Spokane Dome mylonite zone has top-to-the-east kinematics and has been interpreted to have formed in the deep crust either along an east-vergent thrust of the Sevier orogenic belt (Rhodes and Hyndman, 1984) or along an east-vergent detachment fault during Eocene extension, possibly overprinting an earlier thrust-related mylonite (Doughty and Price, 1999).

A major top-to-the-east detachment fault (Kettle fault) is exposed along the flooded Columbia River ~115 km west of the west end of the MT profile (Rhodes and Cheney, 1981; Parrish et al., 1988), but it may be present in the deep crust in the western part of the MT profile. Parrish et al. (1988) suggested that the Kettle and related faults to the north in British Columbia may have offsets on the order of 30–40 km. Hurich et al. (1985) used seismic reflection to track the mylonitic rocks along the fault for 10 km east of their surface exposure, dipping at ~15.5° eastward toward the MT profile. Potter et al. (1986) used the COCORP deep seismic-reflection profiles to track the gently dipping (12°) Kettle fault to depths of 12–15 km at a position ~55 km west of the west end of our MT profile. If the Kettle fault maintains that gentle dip farther east, it would enter the west end of our profile at a depth between 23 and 27 km.

**Crustal Structure: Previous Work**

Overthrusting of Belt Supergroup over Mesozoic and Paleozoic rocks has long been recognized in the geologic mapping of the front ranges of the northern Rockies (Willis, 1902; Campbell, 1914), and it was recognized that the thrust surfaces dipped below the surface to the west. Seismic-reflection studies in support of petroleum exploration in the foothills of the Canadian Rockies began in the 1940s (Link, 1949; Bally et al., 1966) and greatly clarified the crustal structure of the upper 5–10 km of the frontal part of the thrust belt east of our study area and in southern Canada. Industry seismic surveys were extended west into our study area in the 1980s with recording depths down to 15 km (Fritts and Klipping, 1987a, 1987b; Yoos et al., 1991), and deep (5.2 and 3.5 km, respectively) wildcat petroleum exploration wells (Gibb #1 in Montana—Boberg et al., 1989; Moyie #1 in southern Canada—Cook and van der Velden, 1995) were drilled on major lower Belt Supergroup anticlines (Purcell anticlinorium and Sylvanite anticline, respectively). Geologic mapping at the scale of 1:250,000 was completed over most of our study area in the 1980s (Harrison et al., 1992; Miller et al., 1999). Deep seismic-reflection surveys penetrating the entire crust were completed soon after as part of the COCORP program in western Montana and northern Idaho (Yoos et al., 1991) and as part of the LITHOPROBE program in adjacent southern Canada (Cook and van der Velden, 1995; van der Velden and Cook, 1994). Upper-crustal (<15 km) MT studies in western Montana (Fig. 2, lines KCU-1, KCU-5, and KCU-9) and southern Canada occurred during the same period (Gupta and Jones, 1995). Significant interpretations of crustal structure in western Montana have occurred since that time utilizing some or all of the previous data (Kleinkopf, 1997; Price and Sears, 2000; Fuentes et al., 2012).

The most comprehensive studies of crustal structure in the region have been the deep seismic-reflection studies by the LITHOPROBE program along strike in southern British Columbia (Cook and van der Velden, 1995; van der Velden and Cook, 1994). These studies presented 11 detailed interpretations of full crustal seismic-reflection profiles over an area from 50 to 225 km along strike to the northwest of our MT transect; one of these interpreted sections is reproduced here as Figure 3A. Broadly, these profiles show a 50-km-thick crust to the east of the Rocky Mountain Trench, with ~12 km of thrust-imbricated Proterozoic and Paleozoic strata overlying about a 38-km-thick crystalline Archean and Paleoproterozoic basement. Their seismic profile array shows a dramatic thinning of the crystalline basement to less than 20 km over a 40 km swath centered on the Rocky Mountain Trench, coinciding with an overall crustal thinning down to 35–40 km to the west. East of the Rocky Mountain Trench, the basal décollement of the Cordilleran thrust belt is interpreted to be either atop crystalline basement or within the thin (1–2 km) Paleozoic sequence that rests depositionally on the crystalline basement. West of the Rocky Mountain Trench, the profiles are interpreted to show the basal décollement atop crystalline basement or atop a thin veneer of lower Belt strata that rests depositionally on crystalline basement.

The structure of the thrust belt above the basal décollement and west of the Rocky Mountain Trench was interpreted relatively simply by Cook and van der Velden (1995). The thick Belt Supergroup sequence is basically shown to have been transported bodily eastward with an attached 5–10-km-thick slab of crystalline basement. The Belt strata are shown as mildly deformed in a broad, gentle anticline (northern continuation of the Sylvanite anticline) draped over a similarly shaped crystalline basement thrust slab, with a gentle syncline to the east. Later, more detailed analysis by Ainsworth (2009) reinterpreted the crystalline basement slab to consist of Aldridge and equivalent Prichard Formations strata with abundant mafic sills. The Moyie thrust west of the Sylvanite anticline is shown to have moderate (20 km) top-to-the-east displacement.
Figure 3. Comparative crustal-scale cross sections through the Belt-Purcell Basin in southern Canada and northern Montana-Idaho. (A) Cross section in southern Canada at ~49°15′N from Cook and van der Velden (1995). (B) Cross section along seismic lines ID-2, MT-2, and WTF-82-1 from Yoos et al. (1991). (C) Cross section from this study along the western two thirds of same corridor as part B. Abbreviations: FF—Flathead fault; HT—Hefty thrust; KF—Kettle fault; LT—Lewis thrust; LIT—Libby thrust; MT—Moyie thrust; NAB—North American basement; PA—Purcell anticlinorium; PT—Pinkham thrust; PRC—Priest River metamorphic core complex; PTF—Purcell Trench fault; RMT—Rocky Mountain Trench topographic low; RMTF—Rocky Mountain Trench fault; SA—Sylvanite anticline; WT—Whitefish thrust.
Cook and van der Velden (1995) recognized several west-dipping normal faults, but they did not recognize an east-dipping Purcell Trench low-angle normal fault cutting the thrust belt structures. The Rocky Mountain Trench fault, on the east side of the physiographic trough, was interpreted to be a listric west-dipping normal fault with ~10 km of post-thrust displacement. Several west-dipping normal faults were interpreted west of the Moyie fault with relatively small (<2 km) displacements.

Interpretation of the COCORP deep seismic-reflection data in western Montana and northern Idaho (coincident with our central MT profile) by Yoos et al. (1991) is broadly similar to that of Cook and van der Velden (1995); their interpreted cross section is reproduced in Figure 3B. The crust thins from ~45 km to 35 km over 40 km horizontally, centered on the Rocky Mountain Trench. Crystalline basement thins from 35 km to 10 km in thickness over that same interval. The basal thrust décollement occurs within the lower Belt Supergroup section west of the Rocky Mountain Trench, overriding 3–7 km of basal Belt strata in depositional contact with the thinned crystalline basement. They also interpreted an allochthonous slab of crystalline basement to override autochthonous Belt strata at depth under the core of the Sylvanite anticline. The deep structure of the thrust belt east of the Sylvanite anticline is significantly imbricated in the interpretation of Yoos et al. (1991), in contrast to essentially preserved coherent stratigraphy 50 km to the north in Canada, in the interpretation of Cook and van der Velden (1995). Yoos et al. (1991) also interpreted the Purcell Trench low-angle normal fault to cut thrust belt structure for up to 50 km east of its surface expression, penetrating the crust to a depth of 10–12 km.

We present a structural interpretation of that portion of the Belt Basin centered on the Montana and Idaho COCORP profiles (Fig. 3C for comparison). Our interpretation relies primarily on magnetotelluric data but draws heavily upon the existing seismic data. In contrast to previous studies, our interpretation gives more significance to Eocene extension along the Purcell Trench and Kettle faults.

METHODS

MT and Electrical Conductivity

Electrical conductivity within Earth spans many orders of magnitude, reflecting variations in lithology, mineralogy, and hydrology (Guéguen and Palciauskas, 1994; Palacky, 1987; Haak and Hutton, 1986; Keller, 1987). As such, spatial variations in conductivity record structural information inherited from past tectonic events and the current physico-chemical state. Because the conductivity of crustal materials is typically quite low (<<1 S/m), it is common to work interchangeably between conductivity and its reciprocal, resistivity. On average, the Phanerozoic section in the western United States averages 100 Ω-m, while crystalline continental upper crust is quite resistive (105 Ω-m), reflecting the low conductivity of silicate and carbonate rocks (Gough, 1986). In contrast, continental lower crust is commonly less resistive (20–30 Ω-m), due to the presence of a minor conducting phase (Hyndman and Shearer, 1989; Haak and Hutton, 1986; Shankland and Ander, 1983).

MT studies seek physical explanations for regions of anomalous conductivity. Certain mineralogies such as metallic sulfides and graphite are highly conductive (σ ≥ 1 S/m), giving rise to enhanced bulk conductivity when these minerals are present even in small percentages. Within regions of active or recent tectonics, both aqueous fluids and partial melts are common causes of elevated crustal conductivity (e.g., Hyndman and Hyndman, 1968;Wei et al., 2001; Li et al., 2003; Wannamaker et al., 2008). In both these cases, rock conductivity is dominated by ionic conduction within the fluid phase.

MT is a natural-source electromagnetic technique that involves surface measurements of orthogonal electric- and magnetic-field variations (Tikhonov, 1950; Cagniard, 1953). Interaction of the solar wind (a stream of charged particles ejected from the Sun) with Earth’s magnetic field distorts Earth’s ambient magnetic field and produces low-frequency electromagnetic energy that penetrates Earth. Relative to the incident magnetic field, the amplitude, phase, and polarization of the measured electric field depend upon the conductivity of the medium it travels through. At the frequencies of interest (10−3–10−4 Hz), the propagation of electromagnetic energy in Earth is described by a diffusion equation, of the same form as that governing heat diffusion through a solid. Within this frequency range, lower-frequency energy penetrates deeper within Earth and averages over a larger volume. Bulk resistivity also affects the diffusion of electromagnetic energy, with greater subsurface resistivity giving rise to a greater depth of penetration for a given frequency. Typical depths of exploration in MT range from hundreds of meters to tens or even hundreds of kilometers.

MT data can be inverted, with varying degrees of complexity for one-, two-, or three-dimensional (1-D, 2-D, or 3-D) resistivity models. The data used in inversion of MT data are frequency-dependent transfer functions between the horizontal electric (E) and magnetic (H) field vectors. This complex, four-element impedance tensor, \( Z = E^2/H \), contains information about the heterogeneous and sometimes anisotropic subsurface resistivity structure. Vertical magnetic-field transfer functions, \( T_z \), are also commonly estimated and inverted alongside MT impedance data to help further constrain subsurface resistivity structure. \( T_z \) data are particularly sensitive to lateral changes in resistivity. Further background on the MT method can be found in Chave and Jones (2012, and references therein).

MT Data Acquisition

This paper describes the analysis and interpretation of MT measurements along a series of subparallel profiles within the Belt Basin (Fig. 2). Broadband (10−3–10−3 Hz) and long-period (1–10−3 Hz) MT data were collected at a total of 34 sites along a 150 km transect crossing the central Belt Basin from the Priest River metamorphic core complex (Selkirk Crest) in the west to
the Whitefish Range in the east. MT data collected along this central profile (hereafter the main profile) closely followed COCORP deep seismic lines ID-2 and MT-2, both collected in the mid-1980s (Yoons et al., 1991). Additionally, industry broadband MT data collected in 1984 by Phoenix USA along three subparallel profiles (KCU-1, KCU-5, and KCU-9; Fig. 2) were reanalyzed. The central of the three profiles overlays the east end of the main profile, and these data were incorporated into our analysis of this profile (Fig. 2). 1-D and limited 2-D analysis of this industry data was carried out by Gupta and Jones (1995). Regional data used in this study include six long-period MT sites recorded in 2008 as part of the EarthScope MT Transportable Array (http://ds.iris.edu/spud/emtf).

Together, this rich data set permits detailed characterization along the main transect while maintaining the wide aperture needed to look at basin-scale structural variability. Figure 2B shows station locations overlain upon the intermediate-wavelength isostatic residual gravity map of Mankinen et al. (2004) and illuminates major structural features including gravity highs over the Purcell and Sylvanite anticlines and lows over the Purcell and Rocky Mountain Trenches and the Libby syncline. At the 23 western sites, both long-period and broadband data were recorded. At the 11 eastern sites, which overlap with broadband stations in Phoenix USA line KCU-5, only long-period data were collected. All sensors were oriented in geomagnetic north and east directions, and electric dipoles were on average 100 m in length. The broadband data along the main profile were collected at a sampling rate of 500 (8) Hz using induction coil (fluxgate) magnetometers and nonpolarizable lead-lead chloride electrodes. Broadband stations recorded for 1–2 d, while long-period stations recorded for up to 2 wk; the resulting MT and vertical magnetic-field transfer functions are at most sites well defined to 10^{-4} Hz. Time-series processing was carried out using the multistation remote-reference transfer-function estimation program of Egbert (1997). The MT and vertical magnetic field transfer functions for all stations collected along the main profile, including time series for the long-period data, are archived and available through the Incorporated Research Institutions for Seismology (IRIS) Data Management Center (http://www.iris.edu/dms/products).

Previous MT Studies

Early electrical studies identified high conductivity within or below the lowest exposed Belt Supergroup in western Montana and southern British Columbia. Audio-frequency magnetotelluric (AMT) data, sensitive to the upper few kilometers, were collected by Wynn et al. (1977) along a 45 km transect crossing the Purcell anticlinorium at a location between the central and southern MT profiles (Fig 2). Those authors identified a 10–15-km-wide zone of high conductivity just east of the hinge line, the top of which was between 1 and 2 km below the surface exposure of Prichard Formation. AMT studies atop the Sylvanite anticline just west of the northern MT profile revealed similar high conductivity at a depth of just over 1 km (Long, 1988). Neither of these studies had sufficient penetration to image the base of these high-conductivity zones.

The most encompassing previous study of the Belt Basin was that of Gupta and Jones (1995), who interpreted commercial MT data across the Purcell anticlinorium between 48°N and 49.5°N. As part of the current study, we reanalyzed and interpreted the southern three profiles from Gupta and Jones (1995). Through primarily 1-D inversion, the authors imaged a resistive upper crust of 2–6 km thickness above a widespread electrically conductive “basement” extending to their depth of investigation at midcrustal depths. They further concluded that rocks of the upper and middle Belt Supergroup are more electrically conductive than the underlying Prichard Formation, an interpretation that we revisit in the section on “Identifying Belt Stratigraphy in the 2-D Resistivity Section.”

COCORP Deep Seismic Data

Two 96 channel, deep seismic-reflection profiles (ID-2, MT-2 in Fig. 2) were previously collected (Yoons et al., 1991) along two segments of the central MT profile. The data acquisition parameters for those profiles were given in Yoons et al. (1991). A depth-migrated version of one of those profiles (MT-2) and a line interpretation of it was illustrated in Yoons et al. (1991). We acquired the original seismic data package for both lines from COCORP. The data for ID-2 were processed by the U.S. Geological Survey (USGS) Seismic Data Processing (SDP) group, and a depth-migrated section was generated for ID-2. We subsequently developed a line interpretation of that depth-migrated profile. Yoons et al. (1991) also presented a line interpretation of an unmigrated seismic-reflection profile previously collected by the petroleum industry (Fig. 2; Signal of Montana WTF-82-1).

In addition to the reflection sections, seismic interval velocities were calculated by the USGS SDP group from the velocity models picked by COCORP and used to stack the data. The stacking velocities were first smoothed using a horizontal average over 50 common depth points (CDPs) and a vertical average over 500 ms of two-way traveltime; the resulting average stacking velocities were calculated at an interval of 50 CDPs. The Dix equation (Dix, 1955) was finally used to convert average stacking velocities into interval velocities, which can be more directly related to subsurface velocity variations. These seismic interval velocities, as well as line interpretations of reflection profiles ID-2, MT-2, and WTF 82-1, are discussed further in the section on “Geologic Interpretation of the MT Model.”

MT Data Dimensionality

The dimensionality of MT data provides important constraints on the complexity of the subsurface. This information is reflected in the symmetry of the impedance tensor, Z, as well as in the vertical magnetic-field transfer function, T. Dimensionality is important because it determines the level of complexity required in modeling and inversion, which in turn determines
the computational resources needed. Defining the dimensionality of MT data and determination of so-called geoelectric strike go hand in hand and are commonly determined statistically by examining phase tensors (Booker, 2014; Caldwell et al., 2004) and induction vectors (Schmucker, 1970) at multiple sites and frequency bands within the survey area.

In the case of a 1-D or layer-cake Earth model, where conductivity varies only with depth, phase tensors are circular, $\mathbf{T}$, vanishes, and there exists no well-defined geoelectric strike. In a 2-D Earth with a well-defined geologic strike, $\mathbf{Z}$ can be decomposed into transverse magnetic (TM) and transverse electric (TE) modes in which electric currents flow across and along strike.

Over a 2-D Earth, phase tensors plot as ellipses with geoelectric strike parallel to one of the ellipse axes and the normalized skew angle $\psi$ is, to within measurement errors, zero. The real and imaginary parts of the complex vector $\mathbf{T}$, can be plotted as induction vectors, and, in the case of 2-D structure, they are oriented perpendicular to geoelectric strike. In 3-D, the phase tensor skew parameter is statistically different from zero, and the induction vectors do not exhibit a consistent direction either across the survey or as a function of frequency; in such cases, the concept of geoelectric strike breaks down, and a 3-D analysis becomes necessary.

Figure 4 shows phase tensors (PT) and real induction vectors (IV) for all long-period stations along the main profile. This information is shown in pseudosection format, in which period serves as a (nonlinear) proxy for depth. The ellipses are colored according to the maximum principal phase, which is related to subsurface resistivity. Phase values approaching 90° indicate the presence of anomalous conductive material in the subsurface. We used the information contained in Figure 4 both qualitatively and quantitatively to examine structure, dimensionality, and geoelectric strike.

Taken as a whole, the real IVs, which as displayed point toward regions of high electrical conductivity (Parkinson, 1962), show a strong east-west grain, particularly from 10 to 1000 s period, and point toward the center of the basin. This, together with a broad region where the maximum principal phase is above 80°, indicates a large region of high conductivity beneath the center of the profile. High phases extending to periods of 1000 s or more indicate that a portion of this high conductivity is deep, likely within the middle to lower crust. A decrease in the maximum principle phase at the longest periods, particularly near the ends of the profile, may reflect igneous basement beneath the Belt Supergroup metasediments.

Abrupt changes in the IVs and PTs provide indications of structural changes along the profile. In this way, absent any modeling assumptions, we directly interpret several such boundaries to mapped faults or folds. Changes in IV direction, many accompanied as well by changes in the PT ellipticity, orientation, and maximum principal phase, occur at or near the surface trace of the Purcell Trench fault, the Moyie thrust, and the Pinkham thrust. Additional changes in the PT parameters are seen along the hinge lines of the Sylivanite and Purcell anticlines.

The PT and IV information can finally be used to extract quantitative information about dimensionality and geoelectric strike direction. As illustrated in the inset (Fig. 4), each individual PT and IV diagram, pertaining to a single period at a single station, contains directional information in map view. If sufficiently 2-D, the direction of geoelectric strike is both perpendicular to the IV direction and parallel to one of the PT ellipse axes. Whether each diagram is considered 2-D is a function of $\psi$, the normalized skew angle (not shown), which is a measure of the asymmetry of the phase tensor. In determining geoelectric strike, we take the recommendation of Booker (2014), and consider $|\psi| < 6^\circ$ to be a necessary condition for a 2-D phase tensor. We also require the maximum and minimum principal phase angles, which are related to the lengths of the major and minor ellipse axes, respectively, to differ by at least 5°. This latter condition ensures that circular, or 1-D, phase tensors do not bias any statistical determination of geoelectric strike. Finally, we apply these conditions to only a subset of the data, ignoring the four shortest and four longest periods, as considerable scatter is evident in the PTs and IVs at these periods (Fig. 4).

Using this subset of the data, a histogram of $\psi$ (Fig. 5A) shows more than half the data fall outside the 2-D criteria of $\pm 6^\circ$. Based on this and the phase split criteria, less than one quarter of the data are consistent with a 2-D resistivity structure; 1/10 are consistent with 1-D structure ($|\psi| < 6^\circ$ and $\Phi_{\text{max}} - \Phi_{\text{min}} < 5^\circ$); and the majority of the data, including almost all data at periods greater than 100 s, are considered 3-D. Of those data considered 2-D, a geoelectric strike direction between 0° and 10° east of north (not shown) is found at short periods, while longer periods reveal no distinct geoelectric strike direction.

The IVs, however, reveal a more consistent picture (Fig. 5B). At all but the longest periods (1000–10,000 s), rose histograms of IV direction, taking only IVs with magnitude greater than 0.1, suggest a geoelectric strike direction between N30°W and N50°W. This direction is consistent with both the surface geologic strike, as defined by faults and folds, and the trend of potential field anomalies, as shown in Figure 2. At longer periods, the IVs rotate to N15°E. The smooth counterclockwise rotation of IVs with increasing period toward this direction suggests volumetric averaging at the longest-period data between a region of high conductivity beneath the profile and a separate region of anomalous conductivity to the west or southwest of the profile. We do not speculate on the source of this latter high-conductivity region, but we note that 3-D inversion of EarthScope MT data imaged a zone of high conductivity in the lower crust beneath the Coeur d’Alene mining district, 100–150 km southwest of the central profile (Bedrosian and Feucht, 2014). Given our analysis, is 2-D modeling and inversion appropriate? The induction vectors suggest that a 2-D approach, with a geoelectric strike between N30°W and N50°W, is valid at all but the longest periods. The phase tensors, however, indicate a predominantly 3-D data set. Undoubtedly, this data set exhibits 3-D behavior; we are faced, however, with the oft-encountered question of whether the 3-D character is strong.
enough to invalidate a 2-D inversion. We have chosen to take a “trust, but verify” approach, in which we proceed with 2-D inversion assuming a N30°W strike, but we take certain steps to ensure that off-profile structure is not being mapped onto our 2-D model section. First, we do not apply tensor decomposition methods to this data set. The prevalence of data with $|\psi| > 6^\circ$ is inconsistent with these methods, which are based upon a 2-D distortion model with $\psi = 0$ (Booker, 2014). Second, in inverting the data, we focus on the TM-mode data and have excluded the TE-mode data, which are more sensitive to off-profile structure in the case of quasi–2-D resistivity models (Wannamaker et al., 1984). Finally, we carried out a series of 3-D inversions on this data, including data from profiles KCU-1 and KCU-9 and regional EarthScope MT data, in order to verify that basic model features are consistent between the 2-D and 3-D inversions. Note that computational limitations and the nonuniform...
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distribution of MT stations limit the spatial resolution of our 3-D inversions; it is for this reason that we invest in 2-D inversion of this data set.

MT Modeling and Inversion

2D Inversion

We carried out 2-D inversion of the data along each profile following rotation to N30°W, the predominant geologic strike of the region (e.g., Fig. 2). We focus our discussion upon our new MT profile, but the same inversion approach and final settings were also applied to profiles KCU-1 and KCU-9. The 2-D finite-difference conjugate-gradient inversion RLM2DI (Rodi and Mackie, 2001), as incorporated within the Winglink software package, was employed. Within the profile area, an average cell width of 800 m was used. Topography was included in the inversion mesh; cell thickness started at 50 m at the land surface and increased logarithmically with depth. Where present, six decades of data from 100 Hz to 10,000 s were inverted. Regularization was included using a Laplacian model structure function with a trade-off parameter of 30. All inversions were run until convergence, defined as the point at which the iterative change in data misfit fell below a threshold value.

The statistically derived data errors show a clear dependence on period, with larger errors at long periods, reflecting the small number of transfer-function estimates available for averaging. Direct application of these errors results in a very coarse inverse model and a highly nonuniform data fit due to limitations in the data set and our ability to model it. Measured MT data, for example, are sensitive to small-scale resistivity structure that we cannot model due to mesh size limitations, cannot constrain due to limited data bandwidth and site spacing, and cannot recover due to the regularization required to solve the overparameterized 2-D inverse problem. Furthermore, assumptions about the MT source fields are sometimes violated, giving rise to precise but inaccurate data. Following a series of tests with various error schemes, 5% relative error was applied to the TM-mode apparent resistivity data, and 1.4° error was applied to the TM-mode phase data. An absolute error of 0.01 was applied independently to the real and imaginary parts of $T_z$. To suppress 3-D effects during the inversion, the TE-mode apparent resistivity and phase data were excluded from the inversion.

Over 100 individual inversions were run with various subsets of data, applied errors, regularization, and inversion parameters. The final 2-D resistivity model was generated along the main profile from the combined broadband and long-period MT data as follows: An initial inversion was started from a 100 $\Omega\cdot$m half-space using the aforementioned data and applied errors. The resulting inverse model was fed into a subsequent inversion in which static-shift parameters were allowed to vary during the inversion. Finally, this inversion model was modified to include a 1000 $\Omega\cdot$m resistor at 35–45 km depth. The modified model was used to start a final inversion to test whether the measured data are consistent with a resistive upper mantle or

Figure 5. (A) Histogram of normalized skew angle, $\psi$, determined from phase tensors in Figure 4. Red lines indicate $|\psi| < 6°$, which is taken as an indication of 2-D resistivity structure. Much of the data have $|\psi| > 6°$, suggesting 3-D character. (B) Rose histograms of induction vector direction at four different period ranges. At all but the longest periods, induction vectors align along a northeast/southwest corridor (gray bars), consistent with a geologic strike of N30°W to N50°W (green bars).
whether high conductivity is required at these depths in order to fit the measured data. The a priori model was in all cases a homogeneous half-space. This final 2-D inverse model is given in Figure 8B.

**Model Assessment**

Pseudosections of the measured data and corresponding model response are shown in Figure 6. Note that even though the TE-mode data were excluded from the inversion, the model response is shown. The final model represents a 78% reduction in normalized root-mean-square (n.r.m.s.) relative to the starting 100 Ωm half-space model (n.r.m.s. = 29.1). The model does not, however, fit the data to within the specified errors. Given the applied errors, the n.r.m.s. statistic taken over all the inverted data (static-shifted TM-mode apparent resistivity, $\rho_{a}$, and TM-mode phase, $T_z$) is 6.4, split evenly between TM-mode and the $T_z$ data. It is our experience, however, with this implementation of the RLM2D inversion, that artificially small errors must be applied, particularly in $T_z$, to produce models with appreciable structure.

Rather than discuss the n.r.m.s. statistic at these artificially low error levels, we examine the “whiteness” of the fit, that is, how uniformly the data are fit as a function of period and distance along the profile. For this, we examine the n.r.m.s. assuming more realistic errors of 20% in apparent resistivity ($0.2 \ln(p_a)$), 5.7° in phase, and 0.05 in $T_z$. A plot of n.r.m.s. versus site (Fig. 7A) illustrates a uniform misfit across the profile. A few sites are characterized by elevated n.r.m.s. in TM-mode apparent resistivity but not in phase. This is the result of static shifts in the measured data that are not well estimated during the inversion. Similarly, small numbers of sites show elevated n.r.m.s in TM-mode phase but not in apparent resistivity. These correspond to sites where phases are greater than 90° at longer periods; as TM-mode phases are constrained to fall within the range of 0° to 90° (Weidelt and Kaikkonen, 1994), these data violate 2-D assumptions and are hence poorly fit. Looking at n.r.m.s. as a function of period (Fig. 7B), we see that misfit is largely uniform with respect to frequency. An exception to this is TM-mode phase at the longest periods. We attribute the elevated n.r.m.s. to 3-D effects in the data, as evidenced by high values of $\psi$ and out-of-quadrant phases at the longest periods (Fig. 6).

We note that our choice of error levels result in n.r.m.s. values that are close to 1 for the different data subsets. Thus, we conclude that our model fits the measured data to 20% in TM-mode $\rho_{a}$, 5.7° in TM-mode phase, and 0.05 in $T_z$. It is worth noting that even though we have excluded the TE-mode data from the inversion, the calculated TE phase response of the final model fits the TE-mode phase data to within 5.7° errors over much of the frequency range and at approximately half of the 54 sites (Fig. 7). The largest mismatch between the measured and modeled TE-mode response is at long periods, where the high observed TE-mode phase data are underfit by the 2-D model response. This is consistent with a finite strike length to the conductive basin (Wannamaker et al., 1984).

**3-D Inversion**

The 3-D inversion was carried out to assess whether structures identified within the 2-D model section can be attributed to artifacts that stem from the assumptions inherent in 2-D inversion. We furthermore examined the 3-D inversion model to determine the spatial extent of high conductivity within the Belt Basin.

We inverted data using the 3-D method of Siripunvaraporn et al. (2005), which seeks the smoothest, minimum structure models subject to an appropriate fit to the data. A data-space approach, where matrix dimensions depend on the size of the data set rather than the number of model parameters, surmounts the computational demands of construction and inversion of model-space matrices. The data used for inversion include the full impedance tensor at 12 periods ranging from 0.01 s to 6000 s; TE data were not inverted. In total, 91 sites were used, including wide-band data from profiles KCU-1, KCU-5, and KCU-9 and six long-period EarthScope stations (Fig. 2). Computational resources limited the size of the model mesh to 70 × 70 × 40 cells (model extent is shown in Fig. 1), resulting in a horizontal cell size of 3 × 3 km. The thickness of cells increases logarithmically with depth starting from 75 m at the surface. Both the data and model grid were rotated to N30°W to ensure that regularization was applied in a quasi–2-D coordinate system, as described by Tietze and Ritter (2013). We applied errors of 2.5% to the magnitude of the off-diagonal (principal) impedance elements and 5% to the magnitude of the diagonal tensor elements. Applying larger errors to the diagonal impedance elements is common practice, given their generally smaller amplitude in comparison to the off-diagonal principal impedances.

Based on the specified errors, the final 3-D inversion model has a n.r.m.s. misfit of 11, which represents a 93% reduction in data misfit over the 100 Ωm homogeneous half-space starting model (n.r.m.s. of 152). The same half-space model was also used as a prior model. In this manner, the inversion seeks deviations from the prior model only where required by the data, while regions poorly constrained by the data tend toward 100 Ωm. Practically, this leads to model structure only in the vicinity of measurement locations.

**DISCUSSION**

The 2-D model is characterized by strong variations in electrical resistivity both laterally and vertically. Two subhorizontal, highly conductive horizons (Fig. 8B and C2) stand out in the model section. A deep highly conductive, concave-up layer (C1) occurs at 25–35 km depth from just west of the crest of the Purcell anticlinorium and abruptly ends ~20 km east of the Purcell Trench. From this point west to the Selkirk Range, the crust is very resistive (R1), save for some dipping conductive structures within the upper 5 km. A compact resistive feature (R4) is evident from the surface to 5 km depth at ~20 km distance along line; this feature is in the hanging wall of the Purcell Trench fault. Atop C1, there is a pronounced east-dipping resistor (R2) that reaches the surface along the western limb of the Sylvanite anticline and
Figure 6. Pseudosections of measured data (left) and forward response of final central line model (Fig. 8B), shown as distance east from westernmost station. White regions denote data excluded from the inversion. Only long-period data were collected east of km 120. Note that transverse electric (TE) apparent resistivity and phase data were not inverted; the TE-mode response of the final model is shown for comparison. TM—transverse magnetic mode; Tz—vertical magnetic-field transfer function.
extends to 25 km depth. R2 shows some internal variation, with higher resistivity from the surface to ~10 km depth. A midcrustal horizon at 10 km depth (C2) begins west of the Whitefish Range front (120 km along line) and continues to the west for ~60 km, where it ultimately fades away. A more compact conductive feature (C3) is imaged at similar depth beneath the Libby syncline, which at the surface is highly resistive (R3). A more subdued and possibly discontinuous conductive layer (C4) is imaged at ~5 km depth along the eastern third of the model section and is overlain by resistor R5 along the Rocky Mountain Trench. The lower crust beneath the eastern half of the model section is moderately resistive (R6), though significantly less resistive than R1–R5.

Comparison of the 2-D and 3-D Resistivity Models

The 3-D inverse model both confirms and builds upon the 2-D resistivity model. The 3-D model cross section B–B′ (Fig. 9C) is coincident with the 2-D model profile and can be directly compared to the 2-D model (Fig. 8B). Other cross sections and depth slices through the 3-D model reveal the minimum 3-D extent of the prominent conductive features C1 and C2 revealed by the 2-D model (Fig. 9). Our modeling cannot address whether these conductive features extend beyond the stations shown in Figure 2; however, Gupta and Jones (1995) identified high crustal conductivity in MT data to the north of our study area. We will first discuss the 3-D extent of the major conductive features C1 and C2 and then look in more detail at the comparison between the 2-D and 3-D model profiles.

C2 is prominent along all three cross sections, but its along-section extent narrows to the northwest. We interpret C2 to be continuous between the station profiles; the disconnected nature of C2 in the along-strike direction is a result of nonuniform station spacing. At upper-crustal depths, those portions of the model domain far from stations are unconstrained by the data and remain at or near the starting model (100 Ω·m). This is confirmed from a series of 3-D inversions with different starting models, where the resistivity between profiles is observed to scale with the resistivity of the start model.

C2 as imaged in the 3-D inverse model is subhorizontal, with the highest conductivity imaged between 10 and 20 km depth. The top of C2 traces out the broad synclines and anticlines, coming closest to the surface beneath the hinge line of the Purcell anticline. An analysis of the mean conductivity of C2 is presented in a later section, but we note here that the conductivity of C2 as recovered from the 3-D model is significantly higher than that recovered in the 2-D model. This is to be expected, however,
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Figure 8. Geologic, resistivity, and seismic data along the magnetotelluric (MT) profile of this report (located in Fig. 2). (A) Geologic cross section down to ~5 km, adapted from Miller et al. (1999), Harrison et al. (1992), and Harrison and Cressman (1993), with units colored as in Figure 2A. Abbreviations above cross section are as in Figure 3 caption; ST—Snowshoe thrust. (B) Two-dimensional resistivity model down to 50 km depth (relative to sea level), with labeled conductors (C1–C4) and resistors (R1–R6) discussed in the text; dashed outlines with lowercase letter labels identify locations of examples of resistivity stratigraphy shown in Figure 10. MT stations are shown by small circles over profile: red—this report; white—Phoenix USA data from line KCU-5 of Gupta and Jones (1995). (C) Seismic constraints from Consortium for Continental Reflection Profiling (COCORP) lines ID-2 and MT-2, and Signal of Montana line WTF-82-1. Colors depict contoured seismic interval velocities obtained during depth migration calculations from original COCORP data. Short lines are reflection interpretations for depth-migrated seismic section of upper 20 km of line ID-2 (migration by U.S. Geological Survey; line interpretation, this report), depth-migrated seismic section of line MT-2 (migration and interpretation from Yoos et al., 1991), and unmigrated line drawing for the eastern half of line WTF 82-1 (interpretation from Yoos et al., 1991).
as TE-mode data are most sensitive to the conductivity of such elongate structures and were included in the 3-D inversion but excluded from the 2-D inversion.

C1, as imaged in the 2D model, is distinctly deeper than C2 and lies to the southwest of it. In the 2-D model, C1 has a subhorizontal, concave-up geometry with a bulbous, very high-conductivity “knot” at its western end. In the 3-D model section (B–B′), C1 has a similar though broader subhorizontal, concave-up geometry, as well as a similar though shallower high-conductivity knot at its western end. The 3-D model section B–B′ has
an additional bulbous high-conductivity knot in its central part, which lies at roughly 25 km depth, but may extend downward close to the base of the crust. C1 is elongated in the along-strike direction and is most prominent along cross sections A–A’ and B–B’ (both with two very high-conductivity knots), while fading somewhat to the south. This variation may reflect the absence of long-period data along the southern line. The northern profile does not extend far enough to constrain the northern limit of C1; however, data from an isolated EarthScope MT station (Fig. 9B, northernmost station) suggest C1 may continue to the north.

The connection between C1 and C2 varies along strike. C1 and C2 are distinct within the northern part of the model and are separated from one another by roughly 20 km along profile A–A’. Along profile B–B’, C1 and C2 are connected; however, the loci of high conductivity (<5 \( \Omega \)m) remain distinctively separated. Along profile C–C’, C1 is located beneath and partly overlapping C2. This along-strike variation may reflect differences in the amount of shortening and/or extension that has occurred throughout the region. Both the extent and configuration of C1 and C2 in the 3-D model are consistent with 3-D resistivity models derived from EarthScope MT data. The studies of Bedrosian and Feucht (2014) and Meqbel et al. (2014) imaged two highly conductive features R2 is imaged in both the 2-D model and 3-D cross section B–B’. It is somewhat more pronounced in the 2-D model, extending to 70 km along-line distance and to 25 km depth; in the 3-D model, it extends to 60 km along-line distance and between 15 and 20 km depth. The greatest discrepancy between the 2-D and 3-D models, however, is observed at the west end of cross section B–B’, where resistor R1 extends throughout the crustal column in the 2-D model. In the 3-D model, high resistivity (>1000 \( \Omega \)m) is limited to the upper 15 km, below which resistivity drops to ~100 \( \Omega \)m, comparable to the resistivity of R6. Independent support for the validity of the 3-D model in this area is discussed later in the section “Projection of Surface Faults to Depth.”

### Geologic Interpretation of the MT Model

#### Identifying Belt Stratigraphy in the 2-D Resistivity Section

A geologic cross section based on surface geology (Miller et al., 1999; Miller and Burmester, 2004; Harrison et al., 1992; Harrison and Cressman, 1993) is shown above the 2-D resistivity profile in Figure 8A. Profiles of contoured seismic interval velocities from COCORP lines ID-2 and MT-2 are shown in Figure 8C, along with interpreted seismic reflections from those lines and from Signal of Montana line WTF82-1. Several features from the geologic section have recognizable expressions in the MT and seismic data. In the following discussion, specific features are referred to by their depth (with “km” following the number; e.g., 10 km) and horizontal location as measured on the x-axis scale in Figure 8 (with “km” listed first; e.g., km 27). The resistivity stratigraphy developed in this section is shown in Figure 10, while the interpreted geology of the 2-D resistivity section is presented in Figure 11. The discussion in this and the following section can best be understood by referring to both Figures 8 and 11.

Units within the Belt Supergroup stratigraphy can be correlated with distinctive resistivities, seismic velocities, and reflectivities in Figure 8. Areas with Prichard Formation strata exposed at the surface are underlain by intermediate resistivities (~100 \( \Omega \)m), while exposures of Belt units stratigraphically higher than the Prichard Formation are underlain by very high resistivities (1000–10,000 \( \Omega \)m). The stratigraphic relationship between these two units is best seen in the upper 10–15 km below the Libby syncline (km 70) and the Rocky Mountain Trench (km 115). In each case, the surface syncline of middle and upper Belt Supergroup strata (Yb3—all formations above the Prichard Formation) is reflected with an apparent syncline of highly resistive (1000–10,000 \( \Omega \)m) material. Exposures of Prichard Formation in the Purcell anticlinorium and the Sylvanite anticline are underlain by more intermediate-resistivity material (100–200 \( \Omega \)m). Where seismic interval velocities are available across the Sylvanite anticline and the Libby syncline, this intermediate-resistivity layer (Yb2) also correlates with relatively high seismic velocity and seismic reflectivity (Figs. 8A and 11A), reflecting the abundance of mafic sills within the exposed Prichard Formation (Cook and van der Velden, 1995). Yb2 is interpreted to pass under Yb3 in the Libby syncline as a continuous layer. At 10 km below the Libby syncline and the Rocky Mountain Trench, highly conductive layers (C3 and C2, respectively in Fig. 8B) are encountered. Similar conductive layers (between 5 and 10 km thick) are imaged beneath the Rocky Mountain Trench and Purcell anticlinorium on cross sections near the northern and southern MT lines KCU-1 and KCU-9 (Fig. 9C). We interpret these deep high-conductivity layers to represent strata in the lower part of the Belt stratigraphy (Yb1—strata below the mapped Prichard Formation). Possible causes of their high conductivity will be discussed in a later section. This three-layer resistivity stratigraphy and its apparent correlation with established Belt Supergroup stratigraphy are highlighted in Figure 10.

We note that our resistivity stratigraphy is in disagreement with that of Gupta and Jones (1995), who concluded that the upper and middle Belt rocks (Yb3) are more electrically conductive than the underlying Prichard Formation (Yb2). Those authors based their conclusion upon an apparent correlation between surface geology and determinant phase pseudosections. We take issue with this correlation for several reasons: First, in all but a 1-D environment, the determinant phase mixes TE- and
TM-mode phase data, confounding any simple relation between, for example, high phase and low resistivity. Second, the regions of high and low determinant phase, upon which those authors based their correlation, are at periods sensitive to ~5 km depth, and thus cannot be directly related to surface geology, particularly in an area where significant tectonic transport has occurred, as our 2-D and 3-D models show. Finally, given the dimensionality of the MT data (Figs. 3 and 4), our explicit 2-D and 3-D inversions are taken to be more realistic than heuristic displays of measured data.

At the east end of our MT profile, a broad area of the crust below 8–10 km shows intermediate resistivities (R6, 50–500 Ω·m) down to the projected base of the crust (~39 km). We interpret that to represent Archean plutonic and metasedimentary rocks of the Medicine Hat Province (Xmh; Ross, 2002). That intermediate-resistivity body is bounded to the west at km 120 by the eastern flank of a high-conductivity body (C2), with the contact dipping west ~45° from 10 to 25 km depths. We interpret that contact to be the depositional nonconformity of high-conductivity basal Mesoproterozoic Belt strata (Yb1) on a ramp of Archean crystalline basement (Xmh). Cook and van der Velden (1995) observed a similar but less steep basement ramp under the Rocky Mountain Trench in seismic data 50 km to the north in Canada. The more bulbous eastern half of C2 is interpreted to represent a doubled-up section of unit Yb1—an upper Yb1 layer continuous with the western “tail” of feature C2, faulted over a lower, autochthonous Yb1 layer in the lower part of the bulbous part of C2.

In the deep crust (below 20 km) along the central part of the MT model (between km 30 and km 80), a very distinct 5–10-km-thick synclinal layer of very high conductivity (C1) and intermediate seismic velocity is interpreted to be Yb1. It is overlain by a 5–6-km-thick layer of intermediate resistivity and high seismic velocities characteristic of Yb2. That in turn is overlain by a 5-km-thick layer (deeper part of R2) with very high resistivity and intermediate seismic velocities, characteristic of Yb3. This 15–20-km-thick three-layer sequence thus has similar resistivities and seismic velocities to those of the three-layer Belt Supergroup sequence beneath the Libby syncline (Fig. 10). On its east flank near km 80, it contacts a broad intermediate-resistivity block, similar to Archean basement (R6) as interpreted east of km 120. We interpret this block similarly as Archean basement nonconformably beneath the basal very conductive layer of the three-layer Belt Supergroup.

**Projection of Surface Faults to Depth**

Several faults of major (>4 km) interpreted dip-slip displacement are mapped at the surface along our MT profile (Fig. 2). In
this section, we discuss the constraints provided by the resistivity and seismic data on their apparent projections to depth. Since we associate prominent resistivity layers with particular intervals within the Belt Supergroup stratigraphy (units Yb1, Yb2, and Yb3), their terminations may represent fault offsets. Seismic interval velocity lows may also represent highly fractured zones along major faults. Discontinuities in patterns of seismic reflections can further indicate fault displacements, and in some cases, prominent seismic reflections can be produced by faults or by mylonitic fabrics produced by the faults at depth. This discussion is based primarily on the data given in Figure 8, including the shallow geologic cross section, the 2-D resistivity profile, the contoured seismic interval velocities, and the seismic reflectivity. The geologic interpretation of the main profile is presented as line work over the resistivity model together with the contoured seismic interval velocities in Figure 11A. For clarity, the same interpreted model is depicted in Figure 11B as a cross section of colored geologic strata.
The Pinkham thrust crops out at km 113, and, based on the intercept in the Gibbs #1 well (Boberg et al., 1989), it dips gently (~20°) to the west. It has no obvious expression near the surface in the resistivity model but is marked by changes in phase tensor and induction vector orientation (Fig. 4). We interpret it to project above the very conductive feature at the west end of conductive feature C4 (Fig. 8B), and to continue with a gentle dip to the western end of the long synclinal tail of feature C2, intercepting a pronounced low in the seismic interval velocities at ~12 km depth at km 75. Similar relationships are seen in cross sections through the 3-D MT model along the northern and southern MT profiles. We infer the Pinkham thrust to continue farther west under the C3 conductor to ~20 km depth at km 55, where it seems to end against the deep continuous eastern tail of the resistive R2 feature.

The Libby thrust surfaces at km 70 in the middle of resistive feature R3, which we have identified as the Libby syncline. We infer that the fault projects through the prominent seismic interval velocity low at 10 km depth at km 64, passing slightly west of the prominent C3 conductor and its gently west-dipping tail. The moderately dipping east limb and near-vertical west limb of the Sylvanite anticline are interpreted to be abruptly cut by the Libby thrust dipping 30° to the west. The Libby thrust is interpreted to project to 17 km depth at km 49, where it abuts against the deep continuous tail of the resistive R2 feature.

The Moyie thrust occurs at the surface at about km 40, with Prichard Formation strata thrust eastward over upper Missoula Group units (Miller et al., 1999; Harrison and Cressman, 1993). The resistivity profile shows a prominent contact between intermediate-resistivity and very high-resistivity (R2) rock bodies dipping to the west at 30° down to at least 7 km, which we interpret as the subsurface expression of the Moyie thrust. Projection to greater depths is uncertain. The wide section of high resistivity (R2) below the Moyie thrust and west of the Yb2-Yb3 contact on the western flank of the Sylvanite anticline is thought to consist entirely of Yb3. At the surface, where last seen beneath the Moyie thrust, bedding dips are near vertical with tops to the west. If bedding is subvertical throughout that high-resistivity block, then unit Yb3 has about twice the thickness that it does in the Libby syncline, or in the deep synclinal fold 20 km below the Sylvanite anticline. We suggest that the Yb3 unit is folded into a relatively tight syncline, which is paired with a similarly tight Sylvanite anticline to the east. Such a syncline projects toward this point below the Moyie thrust ~20 km south of our transect (Fig. 2A).

The Eocene east-dipping Purcell Trench detachment fault occurs at the surface at about km 11, dropping upper-plate subgreenschist-facies Prichard Formation strata on the east, intruded by unfoliated granitic rocks, against lower-plate foliated granitic rocks and amphibolite-facies Belt Supergroup metasedimentary rocks on the west (Doughty et al., 1998). The resistivity profile shows an east-dipping contact at that location between very high-resistivity rocks (R1) to the west against rocks of intermediate resistivity (Yb2) to the east. The contact is listric (downward-flattening) with a dip of near 60° in the upper 3 km and ~15° at 8 km below sea level. We interpret this contact as the subsurface projection of the Eocene Purcell Trench detachment fault. We would project this fault east to at least km 25, where the sharp vertical resistivity contrast ends at 10 km below sea level. At this point, the Eocene Purcell Trench detachment fault is interpreted to cut and abruptly end the subsurface trace of the older Late Cretaceous–Paleocene Moyie thrust. The trend of the gently east-dipping resistivity gradient continues into a similarly dipping seismic velocity gradient, continuing east to km 33.

Interpretation of the eastward continuation of the Purcell Trench fault east of km 33 is less certain. The fault appears to trend into, but not offset, the vertically continuous R2 resistive feature. However, we suggest the Purcell Trench fault does continue across the narrow neck region of the R2 feature at 13–16 km depth, following a subtle resistivity gradient between the broad, very resistive upper half of the R2 feature and its narrower, less-resistive lower half. At km 50, the fault projects along the sharp, east-dipping resistivity gradient at 17 km depth and follows it eastward to ~20 km depth at km 70, where the R2 feature pinches out. This position of the Eocene Purcell Trench fault would explain the apparent downdip terminations of the Pinkham and Libby thrusts against the lower R2 resistive feature, offset by the younger detachment fault. The Purcell Trench fault trend continues eastward along a zone of strong seismic reflectivity and through a seismic interval velocity low between km 70 and 78 (Fig. 8C) to ~20 km depth at km 80, where the fault projects across the abrupt updip termination of the east limb of the synclinally folded layer C1. Continuing that trend eastward, the fault projects to km 110 at 24 km depth across the abrupt down-dip termination of the highly conductive layer (part of C2) that we interpreted to represent a layer dipping 45° to the west of lower Belt strata nonconformably overlying Archean basement. The Purcell Trench fault can thus be traced in the subsurface for ~100 km east of its surface exposure to a depth of ~24 km in the middle-lower crust.

Above the Purcell Trench fault at 15–20 km depth between km 60 and 100, an intermediate-resistivity horizon occurs below features C2 and C3 that are interpreted as Yb1 horizons. Since it sits below Yb1, could it be a displaced sliver of its depositional Precambrian crystalline basement (Xmh)? Cook and van der Velden (1995) interpreted an allochthonous slab of crystalline basement in a similar position in their seismic line 100 km to the north (Fig. 3A), although later, more detailed work by Ainsworth (2009) reinterpreted those rocks to be equivalent to our Yb2 unit (Prichard-Aldridge Formations). Along the COCORP seismic lines coincident with parts of our MT profile, Yoos et al. (1991) interpreted a thinner basement slab at similar depth but slightly to the west of the sub-Yb1 slab cited here (Fig. 3C).

How do the resistivity and seismic-reflection data compare between this sub-Yb1 region between km 60 and 100 and regions we interpret as Archean crystalline basement? Their intermediate resistivities are characteristic of both the shallow Yb2 in our profile and of the deeper Archean crystalline basement (R6) below the Belt Supergroup along the east side of our profile. However,
the sub-Yb1 region has seismic interval velocities similar to those of Yb2 on the east limb of the Sylvanite anticline, and somewhat higher than those in the interpreted wedge of crystalline basement below the Yb1 layer at 30–37 km depth between km 70 and 80 (Fig. 8C). The western end of the sub-Yb1 region shows prominent, closely spaced reflectors (Fig. 8C), a characteristic of Yb2 here and in the work of Cook and van der Velden (1995). On balance, the data support interpretation of this sub-Yb1 slab as Prichard Formation (Yb2). Since it is overlain by the highly conductive layer (Yb1), we infer a thrust fault along that overlying resistivity gradient, apparently connecting the subsurface traces of the Pinkham and Lewis thrusts (discussed later).

A broad area under the Purcell Trench fault below 10 km and west of km 25 consists of very resistive crust (R1) with little internal resistivity variation. Are the rocks there different than the layered strata to the east (e.g., plutonic?), or are they the same strata that have somehow lost their conductive character? We suggest the latter, as crust there does maintain some seismic velocity layering (Fig. 8C) that can be interpreted as a westward continuation of the deep Belt stratigraphy below the Purcell Trench fault farther east. The intermediate-resistivity Yb2 layer above the western end of the deep Yb1 layer at km 30 appears to continue to the west as a similar layer thickness of high-seismic-interval-velocity material (20–26 km depth) overlying a lower-velocity layer with thickness similar to Yb1. The Yb2 layer is arched into an anticline, the western limb of which continues to almost 30 km depth at km 10, where it either ends or bends upward farther west. We interpret it to end against the east-dipping Kettle detachment fault, which, based on COCORP data to the west discussed earlier, projects into our profile at about that position. The high-seismic-velocity Yb2 layer is overlain by a wedge of lower-velocity material at 17–20 km depth between km 27 and 33 that is interpreted as the westward continuation of layer Yb3. That layer is overlain by higher-seismic-velocity material along a contact that dips west and continues along the seismic velocity gradient along the top of the westward continuation of the Yb2 layer. We interpret this contact as a west-dipping fault with Yb2 above it, ending downdip against the east-dipping Kettle fault. A wedge of lower-seismic-velocity material at ~20 km depth west of km 8 is interpreted as a wedge of Yb1 above the fault and below Yb2. Supportive evidence for the continuation of Yb2 layers in the lower crust west of 25 km is given by the coincident 3-D profile (B–B’ in Fig. 9C), which, while showing a significant increase in resistivity west of km 25 and below 10 km, does suggest material of intermediate resistivity like Yb2 below 15 km.

Within the shallower Priest River complex, a prominent band of seismic reflectors occurs at 7 km depth at the west end of the profile (Fig. 8C), gradually deepening east to 10 km at about km 25, where it intersects the resistivity gradient that we interpret to reflect the Purcell Trench fault. Hauser Lake Gneiss, the metamorphosed equivalent of the Prichard-Aldridge Formations of the Belt Supergroup (Doughty et al., 1998), occurs at the surface within the Selkirk Crest portion of the Priest River complex. The strong reflector band is interpreted to represent the Spokane Dome mylonite zone, overlain by Hauser Lake Gneiss and underlain by a 5-km-thick zone of few seismic reflectors and intermediate seismic velocities that we interpret as Archean crystalline basement. The Spokane Dome mylonite zone and underlying Archean rocks crop out ~45 km to the south, dipping gently northward toward the MT and COCORP lines. Whether the Spokane Dome mylonite zone represents an exhumed deep thrust fault (Rhodes and Hyndman, 1984) or a deep-seated Eocene detachment fault (Doughty et al., 1998) is not resolved. The contact at ~15 km depth between the nonreflective Archean crystalline basement rocks above and more reflective, higher-velocity rocks below (interpreted above as Prichard-equivalent Yb2) is interpreted to be a gently west-dipping thrust fault.

**Restoration of the Deformed Cross Section**

In this section, we attempt to restore the interpreted crustal-scale geologic cross section (Fig. 11B), derived from interpretation of the 2-D resistivity model and seismic-reflection data (Fig. 11A), to its predeformation state in two steps: (1) to its state at the end of the Paleocene, after Jurassic to Paleocene compressional fold-and-thrust deformation, but prior to Eocene and younger extensional deformation, and (2) to its state in Jurassic time prior to the onset of compressional fold-and-thrust deformation. Figure 12A shows the restoration at end of Paleocene time, while Figure 12B shows the restoration at the start of Jurassic time. For ease of discussion, all the blocks separated by pre-Eocene faults are given labels in Figures 12A and 12B, from B1 (west) to B10 (east).

**Reversal of Eocene and Younger Extensional Faulting**

As discussed already, the geophysical expression of the Purcell Trench fault is interpreted for at least 100 km to the east to a depth of ~24 km at km 110 (Fig. 11). At that point, the fault is inferred to mark the downdip cutoff of the west-dipping, highly conductive lowermost Belt unit (Yb1), interpreted as being in depositional contact with Archean crystalline basement farther east. At about km 80, the fault is inferred to mark the updip cutoff of a similar, deeper, west-dipping Yb1 layer. If we infer that the interpreted Archean–lower Belt contact above the deep Purcell Trench fault is offset from the similar contact below the fault, then the Purcell Trench fault records top-to-the-east displacement of ~26 km, which is significantly less than that inferred previously (Harms and Price, 1992; Doughty and Price, 1999). In Figure 12A, we reverse the 26 km of top-to-the-east displacement along the Purcell Trench fault and realign Yb1 below the fault with Yb1 above the fault. This reconstruction approximately aligns two thrust faults above the Purcell Trench fault with the inferred deep continuations below the fault: the Libby and the Pinkham thrusts with interpreted faults above and below the sliver of Archean basement within the Priest River complex. The reconstruction moves the Yb2 layer in block B7 above the deep layer Yb3 in the core of the deep Belt Supergroup syncline,
suggesting that a thrust fault is still required to account for this older-over-younger contact. The east end of this thrust fault strand would connect into the downdip projection of the Lewis thrust above the Purcell Trench fault. The west end of the thrust fault below this Yb2 layer connects to the thrust fault below block B4, allowing us to identify that deep fault as the continuation of the Lewis thrust.

Reversal of 26 km of top-to-the-east displacement on the Purcell Trench fault also returns a considerable thickness (10–12 km) of crust over the top of the Selkirk Crest region of the Priest River complex prior to Eocene detachment faulting. This is compatible with peak metamorphic pressures of 500 MPa (~15 km burial) during peak metamorphism at ca. 72 Ma as indicated by metamorphic mineral assemblages in metapelites under the Purcell Trench fault ~15 km south of our MT transect (Doughty et al., 1998; Doughty and Price, 1999). Rapid uplift and cooling of the Priest River complex at ca. 50 Ma have been known for some time (Miller and Engels, 1975; Harms and Price, 1992; Doughty and Price, 2000) and document the age of detachment faulting on the Purcell Trench fault.

As mentioned earlier, the gently east-dipping Kettle fault projects into the west end of our profile at a depth between 23 and 27 km (Fig. 11). We added the gently east-dipping (12°) Kettle detachment fault to our interpretation beneath the Priest River complex, suggesting it intersects the base of the crust at a depth of 34 km at km 40. We infer that Archean crystalline basement underlies the Kettle fault at these depths. Reversal of top-to-the-east motion on the Kettle detachment near the base of the crust would put Archean crystalline crust below the deep Yb1 layer that now sits directly on the mantle between km 40 and 65 (Fig. 11). In Figure 12A, we reverse ~55 km displacement on the deep Kettle detachment fault to put at least 3 km thickness of Archean crystalline crust below the deep Yb1 layer.

A major Tertiary west-side-down listric normal fault (Flathead fault) bounds the Tertiary Kishenehn Basin between the Whitefish Range and Glacier National Park ~35 km east of the MT profile (Constenius, 1996), and these features are shown in Figure 12A. The fault was active from 49 to 20 Ma and is interpreted to record ~15 km of normal offset. Some authors (Dahlstrom, 1970; Yoos et al., 1991; Constenius, 1996) have interpreted the fault to flatten into the preexisting Lewis thrust plane at 7–8 km below the surface dipping 3°–6° west. However, other authors (Mudge and Earheart, 1980; Whipple, 1992) have interpreted the Flathead fault to cut and displace the Lewis thrust. Both interpretations result in upward displacement of the Lewis fault east of the Flathead fault. We leave that relationship unresolved in Figure 12A. We assume that the Flathead fault flattens into the Lewis thrust plane and continues downdip into the Purcell Trench detachment fault. Then, of the 26 km of offset suggested for the deep Purcell Trench fault, only ~13 km of top-to-the-east displacement would be required on the upper end of the Purcell Trench fault west of that junction. This alternate
Interpretation would also require us to account for an additional 11 km of shortening in the prethrusting palinspastic reconstruction discussed in the following section.

**Reversal of Jurassic to Paleocene Folding and Thrusting**

Once we have reversed the offset of the Eocene and younger normal faulting from our interpreted geophysical profile (Fig. 12A), the crustal section is restored approximately to its configuration prior to Eocene extension (although post-Eocene erosion is not restored). Pre-Eocene crustal thickness was over 50 km west of the ramp in the crystalline basement at about km 100 in Figure 12A, resulting from Jurassic–Paleocene folding and thrust duplication during Cordilleran shortening. In this section, we attempt to remove the effects of this Jurassic to Paleocene shortening to generate a palinspastically restored cross section, which assumes that the area of the section prior to shortening was equal to that area after shortening (Dahlstrom, 1969; Hossack, 1979). In practical terms, this presumes that original bed lengths have also been preserved. Since the structural styles of folds and thrust faults are well known in the adjacent Canadian Rockies (Bally et al., 1966; Dahlstrom, 1970; Price, 1981; Boyer and Elliott, 1982), we presume that the downlap continuation of this thrust system is deformed in a similar style. In this case, we have interpreted that deformation to have deformed the three-layer resistivity stratigraphy of the Belt Supergroup and its nonconformably underlying Archean crystalline depositional basement.

We restore the Belt Supergroup to its prethrusting/prefolding geometry by reversing top-to-the-east displacement on the interconnected Lewis, Whitefish, Pinkham, Libby, and Moyie thrusts and by unfolding folded strata to a flat, subhorizontal geometry with correct stratigraphic stacking. Figure 12B was constructed by slicing up Figure 12A along faults or along fold axial planes, and rotating and moving them as rigid blocks (numbered B1, B2, etc., in Fig. 12) into predeformation positions as geometrically tight as possible with minimal overlap of blocks. It can be seen that the pieces cannot be fit tightly together, and we have not tried to deform the pieces to produce a better fit. We review the restoration process and rationale moving from east to west, from foreland to hinterland. As with most thrust belts (Boyer and Elliott, 1982), it is presumed here that thrusts propagated to the east through time, younging from the hinterland to the foreland.

At the east end of the MT profile, the block (B8) beneath the Pinkham thrust (combined with the area east of our profile in the Whitefish Range [block B9] and Glacier National Park [block B10] above the Lewis thrust) must be moved westward along the Lewis thrust so that the Yb2 layer in eastern block B10 is aligned against the west side of the deep autochthonous Yb2 layer (minimum of 160 km of displacement). The Yb2 of block B7 must also be moved back to the west (along the Lewis thrust) but less than block B10 (~70 km), fitting nicely under the thin Yb2 layer of block B10 and also against the deep autochthonous Yb2 layer. Block B4 is interpreted to have been contiguous with block B7 prior to Eocene displacement on the Purcell Trench fault (Fig. 12A), so it also moves 70 km to the west, overlain by Yb2 of B9. Block B8 is restored west of block B4, such that their Yb1 and Yb2 layers align. The restored separation of blocks B8 and B9 results from reversal of movement on thrust faults in the Whitefish Range (Fig. 2A).

The block overlying the Pinkham thrust (B6) restores westward along the Pinkham thrust until its Yb2-Yb1 contact is aligned with that along the west side of block B8 (~24 km). Archean crystalline rocks at the east end of block B3 are interpreted to be nonconformable beneath the west end of Yb1 of block B6, so it is restored westward with the B6 block.

West of the Libby thrust, Belt strata are folded into west-vergent folds of increasing tightness to the west within block B5. Unfolding and realigning of the Yb2 layer of the Libby syncline, Sylvanite anticline, and the syncline under the Moyie thrust west of the Pinkham thrust—Fuentes et al., 2012). It also compares well with shortening estimates in the Sevier thrust belt farther south in central Utah-Nevada (DeCelles, 2004).

Several features of the palinspastically restored basin and subsequent thrust belt are worth mentioning. The Belt Basin at this latitude was originally at least 410 km wide west to east, adding Belt exposures west of the MT profile to the 345 km of restored width of the MT profile. This is roughly the size of the present-day Black Sea. Prior to shortening, the Belt strata are restored westward roughly to the present edge of continental basement, marked by the present-day initial Sr = 0.706 line 20 km west of the Okanagan valley near Omak (Carlson et al., 1991). Archean rocks represented by those in the Priest River complex are allochthonous, and they may have originated ~190 km to the west (at a crustal ramp in the Lewis thrust?) beneath the present-day Okanagan metamorphic core complex.

The Lewis thrust system is dominated by a large-displacement basé décollement (Lewis thrust) from which rise imbricate faults (from west to east, the Moyie, Libby, Pinkham, and several Whitefish Range thrust faults) of lesser displacement that shorten the upper plate. The basal décollement climbs up section in the
direction of transport, as illustrated in Figure 12B. In the west, the block of Archean crystalline rocks in the Priest River complex (B3) is interpreted to overlie the westernmost segment of the Lewis thrust (km 35, Fig. 12B). The thrust appears to climb above the Archean basement into Yb1 at the base of block B6 (km 65, Fig. 12B) and continues to propagate within Yb1 under blocks B8 and the western part of B4. The thrust climbs into Yb2 under the eastern part of block B4 (km 200, Fig. 12B) and continues within Yb2 under block B7. West of block B4, a second branch of the fault climbs up section off the lower thrust into a décollement higher in Yb2, above blocks B4 and B7 and below blocks B9 and B10. The relative timing of motion on those two branches is not constrained. Those branches reunite at the east end of block B7 at km 265 (Fig. 12B), where the thrust climbs out of Yb2 and through Yb3, into or above the basal contact with the overlying Paleozoic section.

Comparison with Previous Regional Crustal-Scale Cross Sections

A comparison of our geologic cross section with that of Yoos et al. (1991) along the same transect, and with that of Cook and van der Velden (1995) along a parallel transect 50 km north in Canada is given in Figure 3. Several differences between the interpreted geologic section in this report from those earlier ones are readily apparent.

(1) All three profiles show autochthonous lowermost Belt strata below the basal Cordilleran thrust at and somewhat west of a pronounced thinning of the crystalline basement (at ~125 km in each section of Fig. 3), resting nonconformably on Archean crystalline basement. In the other two profiles, those autochthonous strata consist of 3–4 km of Yb1, while in our profile, this autochthonous section consists of the full 20 km Belt section consisting of units Yb1, Yb2, and Yb3.

(2) The other two sections show an allochthonous slab of Archean crystalline basement of variable thickness ~15–20 km below the Sylvanite anticline, while we find no evidence for that in our resistivity model. However, Ainsworth (2009) provided evidence that the interpreted Archean slab in the Cook and van der Velden profile is actually imbricated Aldridge Formation with mafic sills (Yb2). We do interpret a thrust slab of Archean basement farther west in the Priest River complex, where outcropping Archean rocks south of our line appear to project into our profile (Doughty et al., 1998).

(3) Archean crystalline basement rocks are interpreted to continue in the lower crust across both of the other two sections, while we have strong evidence that the basal strongly conductive Belt unit (Yb1) sits directly atop the seismically inferred mantle with no intervening Archean basement in the middle of our transect. We interpret that to have resulted from low-angle extensional displacement of that Belt section from the west off of its original Archean depositional basement.

(4) While Cook and van der Velden (1995) did not image the east-dipping Purcell Trench normal fault less than 50 km north of our line, and Yoos et al. (1991) imaged that fault continuing at depth for ~30 km east of its surface exposure, we see evidence for offsets along the projected subsurface trend of this fault for more than 100 km east of its surface exposure.

Source of Enhanced Conductivity

We turn now to the source of high conductivity within unit Yb1. Enhanced conductivity within the Belt Basin was recognized early by Wynn et al. (1977) and Long (1988), which they attributed to metallic sulfides within the thick Belt sequence. However, these audio-frequency MT studies were quite limited in areal and vertical extent, sensitive only to a depth of a few kilometers. Gupta and Jones (1995) used broadband MT data covering a wider area than the earlier studies and with deeper response (down to ~10 km) but provided only 1-D modeling and inversion save for a rudimentary 2-D inversion of profile KCU-1 (Fig. 2). In light of the long-period MT data presented here and of our 3-D MT inversion model, we revisit the issue of the source of high electrical conductivity within the Belt Basin. We start by characterizing the high-conductivity zones imaged and then examine potential conductivity mechanisms in relation to the geologic and tectonic setting. In the following section, we speculate on the implications of our preferred interpretation regarding the early evolution of the Belt Basin.

Three aspects of the resistivity model shed light on the cause of the high conductivity—the geometry of C1 and C2, their mean and minimum conductivity, and their vertically integrated conductivity, or conductance. The high conductivity of C1 and C2 continues for ~150 km along geologic strike; this represents a minimum extent given our station coverage (Fig. 9). C2 is between 40 and 60 km wide in the across-strike direction and between 5 and 10 km thick. Note that the segmented nature of C1 and C2, in particular between the MT profiles, stems from our limited ability to resolve structure away from stations. We presume the high conductivity of C1 and C2 is in fact continuous throughout the dashed regions defined in Figure 9. In the 2-D model, C1 is comparable in geometry to C2 but is more tube-like in the 3-D model. Regardless of which model is chosen, C1 is at least 40 km wide in the across-strike direction and ~10 km thick. These estimates suggest that C1 and C2 cover minimal spatial extents of 3500 km2 and 8000 km2, respectively. As we interpret C1 and C2 to be offset segments of an originally continuous horizon, this suggests a minimum original extent of the conductive horizon of nearly 12,000 km2.

Given the range of layer thickness estimates, this constitutes a volume of 60,000 km3 to 120,000 km3 of electrically conductive metasediments at mid- to lower-crustal levels. By means of comparison, this volume is equivalent to 10%–20% of the water in the present-day Black Sea.

The estimated resistivity of C1 and C2 provides additional constraints on the cause of high conductivity. We used the resistivity model to calculate the mean and minimum resistivity of C1 and C2. For this calculation, we chose the 3-D model over its 2-D counterpart because the 2-D model does not use the TE-mode
data in the inversion, so it will not accurately reflect the true subsurface resistivity. We must, however, take care when averaging model resistivity, as not all parts of the 3-D model are equally well constrained. We limited our calculation to those model cells that fall within the broad confines of C1 and C2 (black dashed lines in Fig. 9), and that are also within 5 km of an MT station. We further restricted this calculation to stations along our MT profile, where long-period data provide the most reliable constraints on deep-crustal structure. Finally, in order to examine the resistivity of C1 and C2 separately, we attributed that portion of the model from 5 km to 15 km depth to C2 and from 15 km to 25 km depth to C1. The spatial extents of averaging for C2 and C1 are shown in white on Figures 9A and 9B, respectively; the depth intervals chosen to separate C1 and C2 are shown schematically on profile B–B’ (Fig. 9C).

The mean resistivity of C2, averaged over 655 model cells, is 4.6 Ω·m, and, at one standard deviation, ranges from 3.3 Ω·m to 6.4 Ω·m. The minimum resistivity within C2 is two orders of magnitude less (0.04 Ω·m), corresponding to a maximum conductivity of 24 S/m. C1, in contrast has a mean resistivity of 14.4 Ω·m averaged over 245 model cells, and ranges between 8.2 Ω·m to 25.3 Ω·m at the one standard deviation level. The minimum resistivity within C1 is again orders of magnitude lower than the mean (0.08 Ω·m), with a peak conductivity of 12.6 S/m. In addition to the mean and maximum conductivity, we calculated the conductance of C1 and C2. In contrast to conductivity, which is an intrinsic physical property, conductance is an extrinsic property and scales with the amount of conductive material present and its distribution. Conductance is furthermore one of the best-resolved parameters in MT models (Jiracek et al., 1995) and permits comparison of this study to MT modeling results from other geologic settings. Using the same geometric boundaries to define C1 and C2, we calculated a mean conductance for C2 at 12,100 S, with over 70% of the area defined in Figure 9A (825 km²) having >1000 S conductance. Similarly, the mean conductance of C1 is 5200 S, with over 60% of the area defined in Figure 9B (625 km²) having >1000 S conductance. We note that the estimated conductance likely underrepresents the true conductance, as we have limited our calculation to 5–15 km for C2 and 15–35 km for C1, and thus may not capture the full conductance of these features.

While we are confident in attributing C1 and C2 to metasediments of deep-basin origin, the estimated conductance values are far greater than that of typical continental-basin deposits. A global conductance map assembled by Everett et al. (2003) suggests that conductance of the sedimentary section in continental regions is almost always less than 500 S. While this map is based only on sediment thickness data (and assumes a conductivity of 0.03 S/m for continental sediments), it highlights the difference between our conductance estimates and more typical values. Our conductance estimates are also high in comparison to average lower-crustal conductance, which a compilation of MT models from Precambrian crust suggests is on the order of 20 S (Hyndman and Shearer, 1989). Notable exceptions are found in active tectonic environments. For example, lower-crustal conductance within the extending Basin and Range province falls between 750 and 3000 S (Wannamaker et al., 1997, 2008), while the lower crust beneath the Tibetan and Altiplano Plateaus has been estimated at 6000 S (Li et al., 2003) and 20,000 S (Brasse et al., 2002), respectively. The high conductance in all these cases is attributed to some combination of partial melt and saline fluids exsolved from partial melt.

With these constraints on the geometry, conductivity, and conductance of C1 and C2, we examined the source of high conductivity within the Belt Basin. Within continental crust, the most common causes of anomalous bulk conductivity are aqueous fluids, partial melt, graphite, and metallic sulfides. We examine each of these in turn.

### Aqueous Fluids

In considering aqueous fluids as an explanation for high crustal conductivity, the conductivity of the fluids, the distribution of fluids, and the stability of fluids over geologic time scales must all be considered. For aqueous fluids, the conductivity of the fluid phase depends primarily on temperature and ionic strength. For salt solutions, fluid conductivity increases with concentration until very high concentrations (30–40 wt%) are reached. At a fixed concentration and (upper-crustal) pressure, fluid conductivity also increases with temperature until 300–400 °C (Quist and Marshall, 1968). As the onset of wet partial melting occurs around 650 °C, this is likely the highest equilibrium temperature at which free aqueous fluids can exist. These data can be combined to estimate peak fluid conductivities of 1–30 S/m at upper- and midcrustal conditions. Archie’s law is commonly used to relate fluid conductivity to bulk conductivity. For porosities expected at mid- to lower-crustal depths, bulk conductivity is likely 1–2 orders of magnitude less than the pure fluid conductivity. For example, to explain a 20 Ω·m lower crust, Hyndman and Shearer (1989) estimated between 0.5% and 3% pore waters, assuming seawater salinity. In the case of C1 and C2, with a mean resistivity of 4–14 Ω·m, the required porosity would significantly higher.

The source and stability of such fluids at depth are other considerations. Within active tectonic regimes, the explanation of conductivity anomalies by way of aqueous fluids is tenable. Fluids may, for example, derive from amphibolite to granulite metamorphism in continental crust or directly from the mantle in an extensional environment (Glover and Vine, 1995). Following the cessation of metamorphism or tectonic activity, however, the residence time of deep-crustal water is on the order of 100 m.y. (Bailey, 1990; Thompson and Connelly, 1990). An additional argument against aqueous fluids in stable lower continental crust comes from petrology. Yardley and Valley (1997) argued that retrograde mineral reactions, occurring as rocks cool from their original equilibrium temperatures, will consume any traces of pore fluids remaining from peak metamorphic recrystallization. These arguments suggest a dry lower crust outside of active or recently active tectonic regimes.
Taken together, the mean conductivity of C1 and C2, as constrained by the 3-D resistivity model, would require several volume percent of hypersaline fluids distributed in crack-like pore geometry at depths in excess of 25 km. Any less-extreme fluid conductivity or pore geometry would only serve to increase the required porosity. This is at odds with petrology, which suggests that beyond 0.1%, fluid storage in the lower crust over geologic time spans is implausible (Yardley and Valley, 2000).

Partial Melt

Neglecting compositional differences, the onset of dry melting requires temperatures of ~1200 °C, while water-saturated rocks begin to melt around 650 °C (Lebedev and Khatarov, 1964). The conductivity of the melt phase varies slightly with rock composition but typically falls within the range of 1–10 S/m (Tyburczy and Waft, 1983). This represents the maximum conductivity to expect for a pure melt phase. The bulk conductivity of partially molten rock additionally depends on its distribution within the supporting rock matrix. Numerous mixing models exist (e.g., Hashin and Shtrikman, 1962); however, it is difficult to attain bulk conductivity in excess of 1 S/m at lower-crustal pressures and temperatures without invoking water-saturated melt fractions in excess of 30% (Li et al., 2003). Attributing C1 and C2 to partial melt would further suggest high heat flow and a reduction in seismic velocity, neither of which is observed. It would also give rise to a rheologically weakened crust, a prediction at odds with the supported topography. Finally, the lack of volcanism since the Eocene is inconsistent with a present-day magmatic system; the shear extent of C1 and C2, if attributed to even a few percent partial melt, would put it on par with the largest magmatic systems on Earth. Therefore, we rule out partial melt as the cause of high conductivity within the Belt Basin.

Graphite

Graphite is a highly conductive mineral; if interconnected, less than 1% graphite can give rise to bulk conductivity in the range of 1–10 S/m. Graphite is common in metasedimentary rocks, where it forms through the conversion of organic carbon via metamorphism. It is not uncommon for marine basin sediments to have a significant amount of total organic carbon (TOC). While the TOC content of the sub-Prichard (Yb1) layer is unknown, cuttings from the Gibbs well (Yb2) show TOC at levels of 1%–3% (Boberg et al., 1989). If this carbon were converted to graphite, C1 and C2 are interpreted to be stratigraphically bound and of broad regional extent. We therefore find it unlikely that metamorphic graphite is the sole source of their high conductivity.

In addition to metamorphic graphite, fluid-deposited graphite may form from precipitation of reducing, carbon-bearing (C-O-H) fluids. Fluid-deposited graphite requires a source of carbon, a means of mobilizing it, and a mechanism for precipitating it. Possible carbon sources include carbon-bearing metamorphic fluids, mantle-derived fluids, or incorporation of sedimentary carbon within igneous fluids; cooling is the most efficient means of graphite precipitation (Luque et al., 1995, 1998). With few exceptions, fluid-deposited graphite is thought to form in high-temperature (>550 °C) environments, where it occurs as vein-type deposits in granulite-facies metamorphic rocks or mafic-ultramafic rocks (Luque et al., 1998). Fluid-rock interactions may also facilitate graphite formation by the replacement of silicate minerals, typically in igneous rocks. While exposed rocks of the Prichard Formation outside of the Priest River complex are typically greenschist to amphibolite facies, the older sub-Prichard rocks (Yb1) have likely experienced pressure-temperature conditions similar to metamorphosed Prichard Formation strata (Hauser Lake Gneiss; 675–930 °C and 600–1100 MPa) within the Priest River complex (Doughty and Price, 1999). Reduced C-O-H fluids may have existed early in the basin history. Submarine-vent fluids, such as those that fed the Sullivan massive sulfide sedimentary exhalative deposit in Prichard time (Yb2), may have deposited graphite in older sub-Prichard Formation sediments. Fluids derived from the intrusion of mafic sills into the Prichard Formation may also have deposited graphite at deeper stratigraphic levels. Whether such fluid systems could deposit graphite over the basinwide scales suggested by the resistivity models is unknown. While we cannot rule out fluid-deposited graphite as an explanation for C1 and C2, there remains no direct evidence of graphite within the Belt Supergroup.

Metallic Sulfides

Similar to graphite, trace amounts of metallic sulfides can dramatically lower bulk resistivity. The bulk conductivity of sulfide-bearing rocks is highly variable, with pyrite ores ranging from 0.1 S/m to 10° S/m and pyrrhotite ores ranging from 10° S/m to 107° S/m (Parasnis, 1956). Additionally, under reducing conditions, laboratory measurements on low-grade metamorphic black shales at simulated midcrustal conductions (300 °C and 250 MPa or ~10 km depth) show an order-of-magnitude enhancement in conductivity over measurements at ambient temperature and pressure (Raab et al., 1998). We note that C1, C2, and C3 are all at depths at or below 10 km, similar to conditions where laboratory studies showed conductivity enhancement in black shale. Raab et al. (1998) found the increase in conductivity at elevated temperature and pressure (~1 Ω·m) to be irreversible and attributed it primarily to elongation of the newly formed...
sulfides and to a lesser extent from the transformation of pyrite to pyrrhotite. Fabric is also speculated to play an important role in the conductivity of carbon- and sulfide-bearing rocks. Foliation developed during burial metamorphism increases carbon and sulfide connectivity through the alignment, flattening, and elongation of minerals. The presence of metallic sulfides requires a method of emplacement (sources of metals and sulfur) and a reducing environment for their preservation.

In a marine deep-water environment, sulfides and carbon are often found together, where anoxic conditions can lead to large-scale deposition of both organic carbon and sulfides from blue-green algae hosting sulfate-reducing bacteria (Goodfellow, 2000). The Prichard Formation was laid down in such an environment, most likely in a stratified water column with anoxic bottom conditions, as indicated by both sulfur/carbon ratios and sulfur isotope values (Lyons et al., 2000; Goodfellow, 2000). Disseminated sulfides and carbon are present in the exposed Prichard Formation (Yb2), typically at the 1–3 wt% level based upon cuttings from the Gibbs well (Boberg et al., 1989). In the Canadian Belt, laminated sulfides, predominantly pyrrhotite, are described within the Prichard and Aldridge Formations (Yb2). Hoy et al. (2000) described a series of thin-bedded to laminated pyrrhotite intervals within the Prichard Formation. One of these intervals, the so-called Sullivan horizon, correlates temporally with the Sullivan sedimentary exhalative massive sulfide deposit and is up to 100 m thick proximal to the deposit. At a distance of 5 km from the Sullivan deposit, Goodfellow (2000) examined this interval and found 5–15 wt% pyrrhotite aligned along bedding planes and with a laminated appearance. The surrounding sedimentary rock, in contrast, contained between 1% and 6% pyrrhotite, largely disseminated in the rock mass. We conclude that there is ample evidence throughout the basin for both disseminated and laminated pyrite and pyrrhotite within the Prichard (Yb2) strata. The mean resistivity of C1 and C2 (sub-Prichard; Yb1), however, is significantly lower than that of the Prichard Formation (Yb2) strata. The mean resistivity of C1 and C2 (sub-Prichard; Yb1), however, is significantly lower than that of the Prichard Formation (10–100 \( \Omega \cdot m \)), leading us to speculate that the sulfide content of the lowest (unexposed) Belt Supergroup strata is either significantly higher or has a greater degree of connectivity than the sulfides within Yb2.

In summary, the mean bulk conductivity values of C1 and C2 are considerably higher than can be produced by aqueous fluids given reasonable mid- to lower-crustal porosities. Likewise, we dismiss partial melt as an explanation for the high conductivity given the high melt fractions required and the lack of elevated heat flow or recent magmatism expected to accompany such an extensive region of crustal melt. We cannot rule out interconnected graphite seams as a cause, and some conductivity enhancement, particularly in proximity to fault zones, may be attributed to metamorphic graphite. Metamorphic temperatures may have been high enough to form graphite in the deep crust after Cordilleran deformation had thickened the crust, but conductive bodies now in the middle or upper crust (C2, C3, and C4) are unlikely to have ever experienced temperatures that high. We favor laminated metallic sulfides, predominantly pyrite and pyrrhotite, laid down in a reducing marine environment along a series of discrete stratigraphic horizons, as the primary source of high conductivity in the sub-Prichard metasediments (Yb1).

**Implications of High Conductivity for Early Basin Evolution**

While the lack of exposure of Yb1 limits our understanding of the basin in pre-Prichard time, studies of the Sullivan Zn-Pb-Ag sedimentary exhalative massive sulfide deposit shed light on conditions within the Prichard Formation (Yb2) at the time of this synsedimentary mineralization. The Sullivan deposit formed in and around the vent field of a submarine hydrothermal system around 1470 Ma. Emplacement of sills forced the expulsion of large volumes of heated formation waters along synsedimentary faults within the rift basin. Expelled from below, the ore-forming fluids were both chloride-rich and metalliferous. Most of the sulfur within the Sullivan deposit is thought to have been sourced from local reduction of seawater sulfate (Goodfellow, 2000).

Before examining C1 and C2, it is instructive to consider conductor C4 within Yb2 higher in the section and its association, if any, with metallic sulfides. Cuttings from the Gibbs #1 well (Fig. 2) reveal 1–3 wt% disseminated pyrite throughout the sampled Prichard Formation (Boberg et al., 1989). This corresponds to unit Yb2 in our resistivity stratigraphy, which is generally characterized by resistivities of 10–100 \( \Omega \cdot m \), but which also contains the thin highly conductive interval C4 east of km 100 (Fig. 8B).

The position of conductor C4 in the 3-D resistivity model along the southern MT profile (line KCU-9 on Fig. 2) can be compared to resistivity at depth from geophysical logs of the nearby Gibbs #1 well. Within the 5.4-km-deep well, the lowest-resistivity values recorded occur over a 150 m interval centered at 3.1 km depth, where the resistivity logs drop to 3 \( \Omega \cdot m \). This interval corresponds closely with the depth of C4 along cross section C–C′ (Fig. 9D), which passes through the location of the Gibbs #1 well. Conductor C4 can be traced in the upper 5 km through the eastern part of the 3-D model area (below the Pinkham thrust in the northern MT lines; above the Pinkham thrust in the southern MT lines), suggesting that, like C1 and C2, it represents a stratigraphic horizon. Can the position of C4 be related to stratigraphy established elsewhere in the basin? Boberg et al. (1989) tied the stratigraphy penetrated by the Gibbs #1 well to the regional stratigraphy of Finch and Baldwin (1984), based on drill cuttings and borehole logs. This low-resistivity interval corresponds to the top of the lower Prichard Formation, which is also the stratigraphic position of the “Sullivan horizon.” The Sullivan horizon is a stratigraphic interval rich in laminated pyrrhotite that correlates with the synsedimentary Sullivan massive sulfide deposit and has been identified up to 30 km south of the Sullivan deposit (~120 km NW of the Gibbs #1 well). The association between the thin conductor within Yb2 and the Sullivan deposit is intriguing, and it suggests that C4 may reflect a distal (laminated?) sulfide blanket deposit associated with the Sullivan
or a related exhalative event. Subsequent burial metamorphism is likely to have increased sulfide connectivity, forming an electrically conductive horizon. The apparent folded and segmented nature of C4 (Figs. 8B, 9C, and 9D) is attributed to subsequent tectonic activity recorded by this horizon. We note, however, that although low resistivities are recorded in the geophysical logs, no increase in sulfide content is noted in the Gibbs well cuttings at this interval. This apparent discrepancy may reflect the sampling method or coarse sampling interval (10 ft); given the continuous sampling and larger averaging volume of the resistivity log, we interpret the low resistivity at the Sullivan interval to reflect an increase in sulfide content or connectivity relative to the surrounding formation.

We posit that an electrically conductive Sullivan horizon may extend more widely across the northern part of the Belt-Purcell Basin than previously known. Support for this interpretation comes from the 3477-m-deep Duncan Energy Moyie #1 well located north of the study area near the hinge of the Sylvanite (Moyie) anticline (Fig. 1). A resistivity log from this well reveals at least five discrete high-conductivity zones between 1 and 2 km depth. Similar to the association noted in the Gibbs #1 well, this depth interval is coincident with the Sullivan horizon, as defined by marker sills identified within the Moyie #1 well. Cook and Jones (1995) further correlated this horizon and the interval of high borehole conductivity with a conductor at 1 km depth in an MT resistivity model that intersects the Moyie #1 well. We did not, however, image C4 within the western part of the 3-D resistivity model (Fig. 9), suggesting that the Sullivan horizon is somewhat limited in its westward extent. This observation is consistent with that of Hoy et al. (2000), who suggested that the Sullivan horizon may be limited to, or thicker within, basin-parallel grabens within the broader Belt Basin.

We conclude that the moderately conductive Yb2 (10–100 Ω·m) contains a few percent disseminated sulfides, but that the more conductive C4 is attributed to laminated sulfide stringers at the Sullivan horizon (Hoy et al., 2000). This interpretation is consistent with Goodfellow (2000), who noted the difference in sulfide content between laminated pyrrhotite at the Sullivan horizon and a lesser degree of disseminated sulfide in the surrounding rock. Stratiform or stringer sulfides are recognized at several intervals within the Prichard Formation and are believed to be concentrated at the transitions between cycles of tectonic extension and subsequent basin infilling (Hoy et al., 2000). Four such periods during Prichard Formation deposition are recognized and are thought to reflect times of elevated fluid flow and geothermal gradients, accompanied by increases in up flow of structurally controlled fluids into basin waters. Within this framework, we speculate that C1 and C2 reflect a series of laminated pyrrhotite intervals deposited by a series of pre-Sullivan exhalative events during early extensional episodes. We emphasize that the 5–10-km-thick conductive zones defined as C1 and C2 need not be uniformly conductive; the measured MT response can be produced by a series of thin (meter scale) discrete sulfide horizons within otherwise resistive metasediments.

The conductivity and geometry of C1 and C2, in comparison to C4, suggest either proximity to one or more source vent fields or much larger or longer-lived exhalative events. Within the study area, the shallower of the two conductive layers (C2) remains beyond the reach of exploration, but its upper horizon comes within 5 km of the surface throughout parts of the 3-D model area (Fig. 9). A more regional MT survey may potentially enlarge the mapped extent of C1 and C2, particularly in the north and south directions. We suggest that future MT investigations of the deepest Belt Supergroup strata and their underlying basement should be focused on regions where C2 comes closest to the surface. Such a location, if within the reach of a deep drill hole, would permit further in situ investigation into the cause of high conductivity.

In suggesting that C1 and C2 are the signatures of massive sulfide exhalative events in the early basin history, we ask the question of whether this is consistent with age constraints on basin formation and estimated accumulation rates. As C2 is higher in the crustal column, we are more confident in its thickness estimate (5–10 km) than that of the deeper conductive horizon. Lydon (2000) extrapolated pre-Sullivan accumulation rates of between 25 cm/k.y. and 57 cm/k.y. based on a series of stratigraphically controlled age dates—the oldest constraints being the 1468 Ma age of the Moyie sills (Anderson and Parrish, 2000) and the 1470 Ma age of the Sullivan deposit (Jiang et al., 2000). From Cook and van der Velden (1995), the base of highly reflective crust (which we interpret as the transition from Yb2 to Yb1) is ~3 km below the Sullivan horizon, as inferred by those authors based upon correlation of their seismic-reflection section and stratigraphy in the Moyie #1 well. Using the accumulation rates from Lydon (2000) and the minimum thickness of 5 km for C2, we extrapolated minimum estimates of the age of the base of Yb1 (the onset of Belt sedimentation) between 1502 and 1484 Ma. The higher C2 thickness estimate of 10 km results in an older estimate for the beginning of Belt Supergroup deposition at 1522–1493 Ma. These dates are compatible with other estimates for the onset of Belt-Purcell Basin sedimentation (1543–1497 Ma—Anderson and Parrish, 2000; 1510–1485 Ma—Lydon, 2000) and consistent with the maximum age for onset for Belt sedimentation at 1576 Ma, as constrained by basement gneiss of the Priest River complex (Evans and Fischer, 1986).

Regional MT studies (Bedrosian and Feucht, 2014; Meqbel et al., 2014) provide a coarse image of crustal conductivity and suggest that the highest crustal conductivity, of which C1 and C2 are part, occurs only west of the Rocky Mountain Trench and the ramp in Archean basement below it (Xmh, Fig. 3). We speculate that the present location of the Rocky Mountain Trench was the eastern edge of Belt deposition prior to deformation and that sub-Prichard deposition occurred over an attenuated crustal section prior to significant magmatism, as evidenced by the low reflectivity of Yb1. Basaltic underplating during this time may have served to thermally weaken the crust, facilitating crustal attenuation and feeding hydrothermal fluids to submarine vent systems that disseminated metallic sulfides within the early basin.
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deposits (Yb1). Within this framework, high reflectivity within the Prichard (Yb2) marks the onset of magmatic addition to the crust through a then-hot and mechanically weakened crust.

CONCLUSIONS

We combined new broadband and long-period MT data collected from 34 sites in northern Idaho and northwestern Montana with previous broadband and EarthScope long-period MT data to generate 2-D and 3-D resistivity models of the 35–40-km-thick crust in the region. These data were combined with previous seismic-reflection profiles to constrain a geologic interpretation of the crustal section. The shallow portion of the resistivity model correlates well with the mapped surface geology of the Belt Supergroup. The uppermost unit (up to 5 km thick) consists of all of the Belt Supergroup above the Prichard Formation, is highly resistive (typically 1000–10,000 Ω-m), and has relatively low seismic layer velocities. The intermediate unit (up to 7 km thick) consists of the exposed Prichard Formation and 3+ km of stratigraphy below the deepest stratigraphic exposures of the unit. The intermediate unit has low to moderate resistivity (30–200 Ω-m), relatively high seismic velocities, and high seismic reflectivity, with the latter two characteristics resulting from an abundance of thick syndepositional mafic sills. Locally, a narrow, highly conductive horizon occurs within this intermediate unit and correlates in stratigraphic position with the Sullivan horizon, a distal sulfide-rich blanket deposit broadly dispersed across much of the northern Belt-Purcell Basin. This horizon is considered to be the result of hydrothermal venting that formed the world-class Sullivan sedimentary exhalative massive sulfide deposit in southern British Columbia.

Thick conductive horizons occur deeper in the resistivity section and are interpreted to represent a single, structurally disrupted stratigraphic unit (5–10 km thick) with very low resistivity (4–8 Ω-m), intermediate seismic velocities, and little seismic reflectivity. This unit underlies the intermediate unit (but is nowhere exposed) and is interpreted to sit nonconformably atop Archean crystalline basement of the Medicine Hat Province. By analogy with the Sullivan horizon, the thick conductive sub-Prichard unit may record repeated sulfide depositional events within the early rifting history of the basin. The lack of seismic indications of mafic sills in this sub-Prichard unit suggests that, unlike at the Sullivan deposit, causative hydrothermal fluids were not associated with interlayered mafic sills but may instead have been associated with high heat flow due to basaltic underplating of attenuated continental crust.

Identification of this three-layer Belt Supergroup resistivity stratigraphy allows us to recognize the structurally disrupted stratigraphy in the resistivity section and map out the subsurface continuations of surface faults, as well as positions of unexposed faults. The downdip continuation of the Eocene, east-dipping Purcell Trench detachment fault displaces the high-conductivity sub-Prichard layer ~26 km to the east in the deep (24 km) subsurface over 100 km east of its surface exposure. We interpret the east-dipping continuation of the Eocene Kettle detachment fault to be present in the deep crust beneath the Priest River metamorphic core complex and to continue to the base of the crust to the east. Farther east, the Kettle fault drops the entire Belt stratigraphic section directly onto the underlying mantle, excising its original depositional basement of Archean crystalline rocks.

Reversal of motion on these and other Eocene normal faults reveals a crustal section greater than 50 km thick at the end of Paleocene time and returns the Cordilleran thrust geometries to their positions at the end of Jurassic–Paleocene shortening. Further reversal of apparent thrust displacements of the three-layer stratigraphy along the Lewis, Whitefish, Pinkham, Libby, and Moyie thrusts allows construction of a restored section prior to the onset of Cordilleran thrusting in the Jurassic. A total of ~220 km of Jurassic–Paleocene shortening along these faults is indicated. Allochthonous Belt Supergroup strata restore west of underlying autochthonous Belt strata that depositionally overlie thinned crystalline basement along and west of a prominent ramp in crystalline basement thickness beneath the present-day Rocky Mountain Trench. Given the thickness of the sub-Prichard unit and Prichard sedimentation rates, we suggest that crustal thinning and the onset of Belt Basin sedimentation occurred at roughly 1503 ± 19 Ma.

ACKNOWLEDGMENTS

Louise Pellerin (Green Geophysics) and Victor Labson (U.S. Geological Survey [USGS]) were instrumental in conceiving and championing this work. Bedrosian was initially funded as a USGS Mendenhall Fellow. Long-period MT data were collected with instrumentation made available by John Booker (University of Washington). Regional MT data were collected and made available as part of the National Science Foundation–funded EarthScope program. Reprocessing of COCORP seismic lines MT-2 and ID-2 was carried out by John Miller (USGS). We gratefully acknowledge the support of numerous land owners, both public and private, who provided access to their land. This work would not have been possible without enthusiastic field support of Louise Pellerin, Chrissy Ricks, Rick West, and Jay Sampson. This manuscript was improved as a result of constructive reviews from Jared Peacock, Phil Wannamaker, and an anonymous reviewer.

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Highly conductive horizons in the Mesoproterozoic Belt-Purcell Basin


Manuscript accepted by the Society 8 December 2015
Manuscript published online 6 April 2016