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Flexural rigidity of the northern Rocky Mountains: Relationship to crustal domains and deformational style

Diane Friend

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Flexural Rigidity of the Northern Rocky Mountains:
Relationship to Crustal Domains and Deformational Style

by

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B.A., San Diego State University, 1977

presented in partial fulfillment of the requirements for the degree of

Master of Science

The University of Montana

2000

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Flexural Rigidity of the Northern Rocky Mountains: Relationship to Crustal Domains and Deformational Style

Director: Steven D. Sheriff

The hypothesis that lithospheric structure inherited from the time of continental assembly profoundly influences subsequent tectonism has been strongly supported by recent studies in the middle and southern Rocky Mountains. Compositional differences between tectonic provinces and continued reactivation of the boundaries between them provide significant and long-lived control on the location and style of subsequent deformation. This study investigates correlations between inherited crustal properties, the flexural rigidity of the crust, and observed patterns of deformation in the northern Rocky Mountains. The study area includes most of Montana, Idaho, eastern Washington, eastern Oregon, and southern portions of British Columbia, Alberta, and Saskatchewan. Flexural rigidities are estimated using the Maximum Entropy Spectral Estimation coherence method and range from a low of $8.1 \times 10^{20}$ Nm in the western Snake River Plain to a high of $1.38 \times 10^{24}$ Nm in the Columbia Mountains of British Columbia. Flexural rigidity values show significant correlation with crustal age and the location of extensional features. Correlations between flexural rigidity and deformational style along the fold and thrust belt are suggestive but less certain. Inherited crustal boundaries within the area are delineated by zones of crustal weakness with the significant exception of the Vulcan Low. A finger of relatively low flexural rigidities extending northeastward through the Montana Great Plains corresponds closely to the inferred position of the Great Falls Tectonic Zone and supports the hypothesis that a zone of crustal weakness defines this often reactivated boundary between the Archean Wyoming and Hearne provinces. To the south of the area examined in this study, the flexural rigidity estimates of Lowry and Smith (1995) also show a correlation between periodically reactivated, Proterozoic terrane boundaries and areas of crustal weakness. Correlations between these boundaries and low mantle velocities at 100 km depth suggest that these areas may currently be undergoing reactivation by the injection of hot, young mantle material from the west along these preexisting zones of crustal weakness. When used in concert with other geophysical data sets, flexural rigidity mapping can provide a powerful, predictive tool for the investigation of both inherited structure and deformational processes.
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Introduction

Although compelling and logical, the argument that inherited crustal structure influences subsequent tectonic events is often difficult to support. Crustal processes are dynamic, complex, and often transformative, and they make it difficult to disentangle causality through time. But like a picture slowly emerging from a puzzle with lost and scattered pieces, modern integrated studies, advancing technologies, higher resolution data sets, and increasingly sophisticated models continue to provide new evidence which may eventually elucidate the mechanisms of inheritance. In the Rocky Mountains, the interdisciplinary Canadian Lithoprobe project, the Continental Dynamics-Rocky Mountain project (CDROM), and the Canadian-U.S. Deep Probe experiment are all addressing questions of importance to this hypothesis. Unfortunately, most of these studies concentrate on the middle and southern Rocky Mountains (southern Wyoming south to Mexico) and the northern Rockies north of the United States-Canadian border. While some studies have overlapped into the Rocky Mountains of Montana, there is a noticeable dearth of large-scale crustal studies in this area. This study examines crustal strengths in this overlap area to aid in the search for correlations between crustal structure, inherited tectonic domains, and deformational patterns currently observed.

Crustal strength, as parameterized by flexural rigidity ($D$), or effective elastic thickness ($T_e$), has been found to correlate with deformational style in mountain belts elsewhere throughout the world (Stewart and Watts, 1997), with the spatial and temporal evolution of foreland basins (Waschbusch and Royden, 1992; Watts, 1992), and with large-scale heterogeneities in continental crust (Burov et al., 1998). Recent evidence is also accumulating that lithospheric structure inherited from the time of continental amalgamation continues to influence the physical and chemical modification of the continental lithosphere through subsequent tectonic events (O’Neill and Lopez, 1985; Karlstrom, 1998; Karlstrom and Humphreys, 1998; Pazzaglia and Kelley, 1998). Ancient tectonic boundaries, especially, seem to control large-scale deformation through continued reactivation over very long periods of time (Karlstrom, 1999; Sutherland et al., 2000). Might these ancient tectonic boundaries correlate with crustal strength through time?

Along mountain belts, changes in deformational style have been widely recognized and correlated with changes in lithology or paleogeographic setting (e.g. Marshak and Wilkerson, 1992; Mitra, 1997; Corrado et al., 1998; Macedo and Marshak, 1999; Price and Sears, 2000). Plate motions, and the location and form of plate boundaries impact these characteristics as well. Stewart and Watts (1997) noted that relatively low $T_e$ segments along the strike of the Appalachians correspond to areas where rift structures are thought to occur in the Grenvillian
basement. They suggest that the formation of these structures lowered the strength of the crust in these areas and caused a strongly segmented, transform-dominated Paleozoic passive margin to form. This may have played a role in the location of the present-day salients and recesses of the eastern passive margin which has, in turn, fundamentally controlled sediment deposition along the eastern coast of North America.

Might similar correlations exist within the Rocky Mountains? In the northern Rocky Mountains, both the intracratonic rift that led to the creation of the Belt-Purcell Basin (Winston, 1986) and the Great Falls Tectonic Zone, which appears to have developed along the boundary separating the Archean Hearne and Wyoming provinces (O’Neill and Lopez, 1985, Boerner et al., 1998), have exerted an ongoing and profound influence on subsequent structural deformation in Montana (e.g. O’Neill and Lopez, 1985; Price and Sears, 2000). Previous studies have not searched for connections to crustal strength in these areas.

This study uses the methodology of Lowry and Smith (1994) to estimate crustal strength throughout the Montana Rockies, northern Idaho, eastern Oregon and Washington, and southern portions of British Columbia, Alberta, and Saskatchewan. Spatial variations found are compared to the distribution and boundaries of inherited tectonic domains thought to comprise the area, and to changes in the style and amount of deformation observed along the fold and thrust belt and within selected areas of the adjacent foreland basin.

Data

Gridded topographic, Bouguer gravity anomaly, and magnetic anomaly data were acquired from the Geophysical Data Center of the Geological Survey of Canada and from the United States Geological Survey’s National Geophysical Data Grids. Original grid spacing is 1 km x 1 km for the Canadian topographic data, 2 km x 2 km for the Canadian Bouguer and magnetic anomaly data and the United States magnetic anomaly data, and 4 km x 4 km for the United States Bouguer anomaly data. Topography was obtained exclusively from the high resolution Canadian topographic data set since it extended into the United States well south of the study area.

All original grids are referenced to the Clarke 1866 ellipsoid, with an equatorial radius of 6378.2064 km and an eccentricity of 0.082271. Both Canadian and United States gravity measurements were reduced to the International Gravity Standardization Network 1971 (IGSN71) datum and theoretical gravity values were calculated from the Geodetic Reference System 1967 (GRS67) gravity formula. Bouguer values were computed using a reduction density of 2670 kg/m³ and were terrain corrected in areas of substantial relief. To meet the requirements of this
Inherited crustal domains

Overview

Hidden beneath the spectacular topography and modern geologic surface features of the northern Rocky Mountains is a complex patchwork of distinct tectonic terranes stitched together through billions of years of plate motions. The pattern which has evolved may not only reveal the history of this region, but may also influence its future.

Unfortunately, identifying and delineating these inherited structural terranes is a complex and challenging problem. Much of the evidence has been buried, metamorphosed, deformed dissected and/or eroded over time. Still, terranes accreted to the continent relatively recently may retain visible traces of a unique stratigraphic record, differing igneous or metamorphic histories, surface traces of bounding faults, and paleontological or paleomagnetic evidence which differentiates them from surrounding areas (Monger and Berg, 1987; Siberling et al., 1987).

Gravity and magnetic anomaly data can further refine province boundaries and are especially important for identifying Precambrian terranes of the continental shield (e.g. Thomas et al., 1987; Thomas et al., 1988; Kane and Godson, 1989). Although the non-uniqueness of potential field solutions implies that any boundaries determined from the gravity and magnetic data will be subject to differing interpretations, radiometric age determinations from isolated exposures of crystalline basement rocks (Hoffman, 1989; Ross, 1991) can greatly increase confidence in the results obtained. Tectonic domains and province boundaries can also often be delineated from magnetic field data based on the texture and orientation of the magnetic fabric coupled with abrupt changes, or terminations in the magnetic trend (Gibson, 1987; Kane and Godson, 1989; Ross et al., 1991). In this study, this is especially useful in the Archean provinces east of the fold and thrust belt where the magnetic fabric is exquisitely defined. Figure 4 provides a generalized summary of terranes interpreted to underlie the area of this study (Monger and Berg 1987; Siberling et al., 1987; Hoffman, 1989; Oldow et al., 1989; and Ross et al., 1991).
Figure 1. Topography. Dashed box outlines the study area. Grid spacing is 4 km x 4 km. Gridpoints are at the same geographical coordinates as gridpoints in figure 2.

Figure 2. Complete Bouguer anomaly. Dashed box outlines the study area. Grid spacing is 4 km x 4 km. Gridpoints are at the geographical coordinates as gridpoints in figure 1.
**Inferred crustal terranes**

The western edge of the study area is comprised of exotic and pericratonic terranes accreted to the continental margin during the Phanerozoic. Since the time of amalgamation, this area has been subject to displacement from transpression, igneous intrusions, significant extension, and the creation of one of the largest continental flood basalt provinces in the world. Physical expressions of the terrane boundaries are often discontinuous and regional scale gravity and magnetic data (figures 2 and 3) provide little help in distinguishing individual terranes. Generalized boundaries for these terranes (figure 4C) were adapted from the lithotectonic terrane maps of Siberling *et al.* (1987) and Monger and Berg (1987) which are based on detailed, local geologic evidence.

Magmatic arcs and accretionary prisms accreted to the continent during the Mesozoic lie...
Figure 4. Inferred tectonic domains of the northern Rocky Mountains region (C). Dashed line shows the approximate western edge of Archean crust (Reed, 1993). Archean domain boundaries to the east of this line are delineated primarily by potential field data (A and B) and U-Pb geochronology of crystalline basement outcrops (Ross et al., 1991; Hoffman, 1989). Numbers give approximate crystallization ages in billions of years. RI=Rimbey magmatic arc, TH=Trans-Hudson orogen, VL=Vulcan Low, HL=Hearne province, Loverna Block, HMH=Hearne province, Medicine Hat Block, GFTZ=Great Falls Tectonic Zone, WY=Wyoming province. Proterozoic to mid-Paleozoic crust is inferred to extend westward to the $^{87}\text{Sr}/^{86}\text{Sr} = 0.706$ isopleth (dot-dash line) (Oldow et al., 1989). Miocene Columbia River basalts obscure basement age, terrane boundaries, and crustal isotope data through much of eastern Washington and northeastern Oregon. Generalized outlines of pericratonic and exotic terranes accreted to the western edge of North America in the Phanerozoic bound the area to the west. These terrains are distinguished by bounding faults, distinctive stratigraphic record, and differing igneous and metamorphic histories (Siberling et al., 1987, Monger and Berg 1987), but identified outcrops are scattered and boundaries in many places are inferred. Mo=Monashee metamorphic terrane, Ko=Kootenay terrane, Qn=Quesnellia terrane, Wa=Wallowa terrane, Ba=Baker terrane, Of=Olds Ferry terrane, Rm=Roberts Mountains allochthon. (Numerical values of the gravity and magnetic anomalies are given in figures 2 and 3.)
farthest to the west. Quesnellia and the Olds Ferry terrane (northwest and southwest corners of the study area, respectively) belong to the Eastern Arc (Oldow et al., 1989), a volcanic arc/marine basement complex which developed at the edge of the North American plate and was subsequently thrust onto the distal edge of the continental margin in Early Jurassic time (Murphy et al., 1995). The Baker terrane is thought to be part of the accretionary complex which developed between the Eastern Arc and another volcanic arc assemblage to the west which includes the Wallowa terrane currently exposed in parts of northeastern Oregon, southeastern Washington, and western Idaho (Oldow et al., 1989).

To the east of Quesnellia, rocks of the Kootenay and Monashee terranes (figure 4C) are probably displaced rocks of North American affinity (Oldow et al., 1989). The Kootenay terrane represents two rift assemblages thought to have been deposited on transitional crust continuous with, but outboard of, the North American miogeocline (Murphy et al., 1995). The Kootenay structurally overlies the Monashee terrane, an area of high grade metamorphic rocks exposed as part of an Eocene metamorphic core complex (Parrish et al., 1988). The Monashee is considered to be a separate terrane based on a combination of its age (2.0 Ga and older) and a lower Proterozoic sedimentary sequence which has not been found elsewhere in the Canadian cordillera (Monger and Berg, 1987). It most likely represents crust from a distal location within the former continental passive margin (Oldow et al., 1989).

The $^{87}\text{Sr} / ^{86}\text{Sr} = .706$ isopleth (figure 4C) is thought to roughly delineate the mid-Paleozoic continental margin (Oldow et al., 1989) and in fact, is nearly coincident with the extensively mapped Western Idaho Suture Zone which directly juxtaposes allochthonous island-arc rocks with continental rocks to the east (Lund, 1988). Further to the north, the strontium .706 line is unfortunately obscured by the voluminous Columbia River flood basalts. However, the eastern boundary of the accreted terranes is approximately outlined by a change in the magnetic fabric of the crust (figure 3). Crust accreted to the edge, or deposited on top of the late Proterozoic/early Paleozoic passive margin is characterized by detailed, short wavelength magnetic features. The magnetic signature of this crust is very different from the relatively smooth and featureless magnetic fabric of the Proterozoic crust east of the accreted terranes, and from the beautifully defined, long wavelength and linear magnetic trends of the Archean crust even further to the east (figure 3). The lack of magnetic fabric within the Proterozoic crust may be partly due to the deep burial of crystalline basement by the thrust-thickened sedimentary rocks of the western Cordillera as well as to a fundamental difference in the nature of Proterozoic crust.

Within the Proterozoic crust along the southern boundary of the study area, the Roberts
Mountains allochthon disappears beneath rocks of the Idaho batholith to the north (figure 4C). The allochthon, which extends through Nevada to the south, was emplaced during the late Devonian/Early Mississippian Antler orogeny, when deep marine, early Paleozoic strata were thrust eastward over the continental shelf (Frazier and Schwimmer, 1987). Although the configuration of any outlying volcanic arcs or western margin subduction zones during this time is unclear, the Antler orogeny is thought to have been initiated by the collision of a volcanic arc with the western edge of North America and deformation from southern Nevada to Alaska has been attributed to this event (Oldow et al., 1989).

Two major Archean provinces, the Wyoming and the Hearne, comprise the rest of the study area (figure 4C, eastern half). The Hearne province is subdivided to the north by a prominent magnetic discontinuity known as the Vulcan low (figure 4B and C). This magnetic low has been interpreted as a north-dipping Archean collisional suture (Hoffman, 1989) based on its prominent truncation of magnetic fabric, a parallel magnetic high to the north (the magmatic belt), and a coincident paired gravity anomaly (Thomas et al., 1987). To the north of the Vulcan Low, the Lovernia block of the Hearne province is bounded on the northwest by the magnetic high of the Rimbeey magmatic arc (figure 4C). Ross et al. (1991) suggest that the Rimbeey granites may have either formed from subduction related magmatism along a region of oblique convergence between the Hearne and Rae provinces, or from crustal thickening without subduction during the collision. To the east of the Lovernia block lies an area of distinctive, but complex, magnetic fabric (figures 3 and 4) that may represent a fold and thrust belt of early Proterozoic sediments (Hoffman, 1989). This is bounded to the east by a magnetic low (figure 3) which defines the western edge of the 1.8 Ga Trans-Hudson orogen (figure 4C), the collision zone between the Superior province and the Wyoming and Hearne provinces to the west.

South of the Vulcan Low, the Medicine Hat block of the Hearne province shows Nd and U-Pb crust formation ages of 2.8 Ga (Hoffman, 1989) to 3.27 Ga (Villeneuve et al., 1993), similar to other parts of the Hearne province (Collerson et al., 1988), but older than the 2.6-2.7 Ga ages determined from areas within and immediately around the Vulcan low (Hoffman, 1989; Ross et al., 1991). The distinctive northwest-southeast trending magnetic fabric of the Medicine Hat block (figure 4B and C) terminates to the south against the northeast-southwest trending magnetic fabric of the approximately 150-250 km wide (O’Neill and Lopez, 1985) Great Falls Tectonic Zone (GFTZ). Basement ages within this zone reflect a 1.8 Ga metamorphic event whose signature dies out to the south in rocks along the northern boundary of the 3.1 Ga Wyoming province (O’Neill and Lopez, 1985; Hoffman, 1989). The GFTZ is assumed to predate
the Trans-Hudson orogen since magnetic trends within the GFTZ are truncated by the magnetic trends of the Trans-Hudson orogen (Thomas et al., 1987). O’Neil and Lopez (1985) interpret the GFTZ as an early Proterozoic shear zone which has remained a zone of crustal weakness and continued to influence the structural evolution of this area into Holocene time. Recent studies of 1.8 Ga diorite from Montana’s Little Belt Mountains (within the GFTZ) show whole-rock elemental abundances and isotopic ratios indicative of formation within a collisional tectonic environment containing both Archean and Proterozoic lithospheric components (Mueller et al., 1999). In contrast, Boerner et al. (1998) note that the GFTZ lacks a colinear magmatic arc as well as any electromagnetic evidence of a plate-edge foreland basin, both typical features of other Proterozoic accretionary boundaries or collision zones. They postulate an Archean, intracontinental shear zone origin for the GFTZ.

**Horizontal gradients in gravity and pseudogravity**

Because potential field data is so useful for delineating large-scale structural trends, it is often used to help define inherited crustal structures (e.g. Gibson, 1987; Sharpton et al., 1987; Thomas et al., 1987; Thomas et al., 1988; Ross et al., 1991). Both magnetic anomaly data (as utilized in the analysis above) and the horizontal gradient of gravity (usually the Bouguer anomaly) are commonly used. Given the incredible detail imaged in the magnetic anomaly data (figure 3), especially within the Archean terranes, it would seem that a horizontal gradient analysis of the magnetic data might also be instructive. To avoid the interpretational complexity associated with the dipolar nature of the magnetic field and facilitate comparison between the gravity and magnetic horizontal gradient data, the magnetic anomaly data can first be transformed to pseudogravity.

Pseudogravity is an artificial quantity which is derived from magnetic anomaly data and illustrates something akin to magnetic potential. The pseudogravity transformation is a consequence of Poisson’s relation, which states that the magnetic scalar potential is directly proportional to the component of gravity in the direction of magnetization. This implies that magnetic anomalies can be transformed into the gravity anomalies that would result if the distribution of magnetization equals the distribution of density (Blakely, 1995). For this study, pseudogravity anomalies were computed from the magnetic anomaly data (figure 3) assuming a local magnetic declination of 16 degrees and an inclination of 68 degrees. These values were derived from the IGRF 1990 for a location near the center of the study area and represent only a crude approximation to the correct field since the gridded data set covers such a large area. The
direction of remanent magnetization was set parallel to the induced field, given above, since most of the study area covers continental shield areas where remanent magnetization is usually weak (Lowrie, 1997).

Horizontal gradients of the Bouguer anomaly and pseudogravity data are shown in figures 5 and 6, respectively. Large horizontal gradients (in either gravity or magnetic data) may reflect the presence of a negative-positive anomaly pair which marks the position of an ancient collisional suture, or thick deposits of supracrustal rocks and intrusions contained in a fold belt along the terrane margin. Horizontal gradients highlight discontinuities in upper crustal rocks and tend to suppress the longer wavelength features related to regional, isostatic compensation (Thomas et al., 1988). Sharpton et al. (1987) have shown good correspondence between horizontal gravity gradient maxima and significant structural trends in exposed areas where detailed geologic data is available. Horizontal gradient maxima in gravity and pseudogravity represent abrupt lateral changes in density and magnetization, respectively (Blakely, 1995).

To emphasize differences in structural trend between terranes in the northern Rocky Mountains area, local maxima in the horizontal gradient data are plotted in figure 7. Maxima in the horizontal gradients of both gravity and pseudogravity were determined by searching for data points that were a local maximum in the horizontal gradient along a north-south direction, or either diagonal direction. Points plotted in figure 7 correspond to maxima greater than or equal to the average horizontal gradient. Because upward continuation of potential field data accentuates anomalies caused by deeper sources (Blakely, 1995) the pseudogravity data was upward continued 10, 20, and 40 km and horizontal gradients were computed for these transformations as well. Deeper sources associated with fundamental boundaries between tectonic provinces show up well in the horizontal gradient maxima of the upward continued data (figures 4C and 7B).

Over the very large area examined in this study, horizontal gradient maxima should depict the location of density or magnetization boundaries reasonably accurately. For simple horizontal slabs with steep bounding edges, the maximum of the horizontal gradient will directly overlie the boundary. More complicated geometries and overlapping, or closely spaced, sources can offset the gradient maxima from the true edge. Grauch and Cordell (1987) modeled the expected offset for isolated slabs and dikes as a function of their thickness, dip angle, and burial depth. They found that the offset of the gradient maximum depends on the depth to the top of the density, or magnetization, contrast and scales as the cotangent of the boundary dip angle. Only very deep sources and/or extremely low angle contacts (<< 10°) will produce significant offsets. Additionally, as the size of the study area grows (Grauch and Cordell, 1987), other factors which
**Figure 5.** Horizontal gradient of the Bouguer anomaly. Stippled pattern above the United States / Canadian border is an artifact resulting from the merging of overlapping data sets.

**Figure 6.** Horizontal gradient of pseudogravity.
Figure 7. A. Density boundaries delineated by maxima in the horizontal gradient of Bouguer anomalies over the study area. Small blue dots correspond to maxima greater than or equal to the average horizontal gradient. Large blue asterisks correspond to maxima greater than or equal to 1.5 times the average horizontal gradient. (Large gradients along the United States border are artifacts resulting from the merging of Canadian and United States data sets.)

B. Magnetization boundaries delineated by maxima in the horizontal gradient of pseudogravity. Pseudogravity maxima were calculated for horizontal gradients of the standard pseudogravity transformation (smallest red dots) and the upward continuation of that transformation to 10 km (yellow), 20 km (green), and 40 km (blue). Horizontal gradient maxima preferentially overlie boundaries with steep structural or lithologic contacts (Cordell and McCafferty, 1989). Deeper contacts are more likely to have maxima which survive the upward continuation process. Within this study area, pseudogravity maxima are especially useful for delineating structural trends corresponding to terrane boundaries (see figure 4).
lead to misplaced horizontal gradient maxima become much less significant. With a 4 km spacing between gridpoints in this study, most boundary configurations will yield offsets which are negligible compared to the horizontal distances being considered.

Regional scale horizontal gradient data provide no help in defining the boundaries of accreted terranes along the western edge of the study area, but do help refine the extent and boundaries of the Precambrian terranes. While both gravity and pseudogravity horizontal gradient maxima depict structural trends useful for identifying province boundaries (compare figures 7A, 7B, and 4C), the more continuous, large-scale magnetization boundaries depicted by the pseudogravity data are the most helpful. Even the strontium .706 line thought to define the western, rifted edge of North American Proterozoic crust is reasonably well outlined by maxima in the pseudogravity horizontal gradient data (figures 4C and 7B). Also, strong horizontal gradient maxima from late-Phanerozoic features (e.g. deep, sediment filled graben resulting from Tertiary extension in figure 7A, magnetization boundaries along Laramide uplifts in figure 7B) obscure the larger scale structural trends of the inherited terranes less in the pseudogravity data (figures 5-7).

Horizontal gradients of pseudogravity and the Bouguer anomaly provide the most complete, large-scale estimate of the location and extent of inherited terrane boundaries over a significant portion of the study area. Localized, high-resolution geologic and geophysical data provide control within the western accreted terranes, verify boundaries determined by potential field data over the eastern portion of the study area, and quantify the timeline of both continental amalgamation and subsequent, episodic reactivations of terrane boundaries. While the exact origin and history of these boundaries may never be completely known, current data sets and modeling techniques are increasingly capable of addressing questions concerning the importance of inherited crustal terranes and terrane boundaries on subsequent tectonic events. To explore one aspect of this question in the northern Rocky Mountains, the inherited crustal domains which underlie this region are compared to current crustal strengths (as determined by estimates of the flexural rigidity of the crust) and the observed distribution of deformational styles.

**Flexural rigidity of the Northern Rocky Mountains**

*Introduction*

The apparent strength of the lithosphere, as characterized by the parameters of flexural rigidity ($D$) and effective elastic thickness ($T_e$), is an extremely useful quantity for comparing the integrated strength of the lithosphere between, and among, large geographic regions. Continent
scale studies have shown that high values of $T_e$ generally correlate with Archean shield areas while low values of $T_e$ are most often associated with younger rifts and fold belts (e.g. Zuber et al., 1989, Bechtel et al., 1990, Hartley et al., 1996). High resolution studies along mountain belts around the world indicate that $T_e$ commonly varies along strike and may reflect long-lived crustal features inherited from one tectonic event to the next (Watts et al., 1990; Stewart and Watts, 1997). Even without detailed knowledge of all the variables that affect $T_e$, differences in flexural rigidity throughout a region can elucidate changes in thermal and mechanical properties of the lithosphere and the geographic distribution of tectonic features which depend on these properties. Tectonic style, fold and thrust belt geometry, and foreland basin patterns of sedimentation may all be influenced by, and reflect, these lateral changes in crustal properties evidenced by changes in $T_e$ (Stewart and Watts, 1997).

It should be kept in mind, however, that flexural rigidity is a highly complex variable whose exact physical meaning, in the case of continental lithosphere, can be difficult to quantify (Burov and Diament 1995; Stewart and Watts, 1997). Flexural rigidity depends on the thermal state of the lithosphere, its composition, and its state of stress (Burov and Diament, 1995; Lowry and Smith, 1995). Low flexural rigidities correspond with high heat flow since flexural rigidity is a strong function of power-law creep (Lowry and Smith, 1994), which in turn, depends on temperature and composition. Flexural rigidities also decrease rapidly with increasing tectonic stress, regardless of whether the stress is compressional or extensional in nature (Lowry and Smith, 1995). Unfortunately, some or all of these quantities are often poorly resolved. Additionally, determinations of $T_e$ are model dependent and may suffer from limitations imposed by such factors as the amount of correlation between surface and subsurface loads (Forsyth, 1985; Macario et al., 1995), the resolution of crustal heterogeneities (Burov et al., 1998), or an appropriate choice of elastic plate and isostatic compensation models (Stewart and Watts, 1997).

Determinations of $T_e$ ultimately arise from the theory of isostasy. In its most general sense, the theory of isostasy assumes that there exists some compensation depth below which the earth is in hydrostatic equilibrium. To maintain this equilibrium over geological time scales, loads placed upon, or within the rigid lithospheric plates must have some form of long-term support. Local support, or isostatic compensation, can be achieved by either thickening a constant density crust to produce a buoyant root (Airy isostasy), or creating lateral variations in crustal density to compensate for the load (Pratt isostasy) (Airy, 1855; Pratt, 1855). The Vening Meinesz elastic plate model provides an additional, regional support mechanism (Lowry, 1997). In this model, the earth's upper layer behaves like a strong, elastic plate overlying a weak fluid.
Loads are supported by stresses within the plate and by flexure of the plate which displaces the fluid below, providing buoyant support over a broad area. The amount of flexure is determined by the elastic properties of the plate, and as such, can be parameterized by $D$ or $T_e$.

Specifically, the flexure of elastic plates is determined by the fourth-order partial differential equation

$$\frac{D}{d^4} \frac{w}{dx^4} + P \frac{d^2}{dx^2} + \Delta \rho g w = q_a(x)$$

(1)

in which $w$ is the vertical deflection of the plate, $P$ is the horizontal stress applied to the plate, $\Delta \rho$ is the density contrast between the material within and below the plate, $g$ is the gravitational acceleration, $q_a$ is the vertical stress, or load, applied to the plate, and $D$ is the flexural rigidity. $T_e$ is related to $D$ through the equation:

$$D = \frac{E T_e^3}{12(1 - v^2)}$$

(2)

where Young’s modulus, $E$, and Poisson’s ratio $v$, are material constants which characterize the elastic behavior of the plate (Turcotte and Schubert, 1982). For a given load, either surface or subsurface, the elastic plate equation can be solved for the resulting vertical deflection of the plate for a range of assumed flexural rigidities. This can be compared to the “observed” flexure determined by modeling the load response of individual tectonic features, by the forward modeling of gravity data, or by various spectral methods such as admittance and coherence.

The “observed” $T_e$ represents the thickness an imaginary, elastic plate would have to have in order to match the deflection of the real, inelastic lithosphere under identical, applied tectonic loads (Burov et al., 1998). Flexurally rigid plates have a large $T_e$ and deform significantly under only the largest (i.e. longest wavelength) loads. As $T_e$ decreases, the amount of plate flexure increases, the plate thickens, and Airy compensation becomes the dominant mechanism of support. For small $T_e$, only the shortest wavelength loads can be supported by stresses within the plate. The magnitude of $T_e$ determines the wavelength at which the dominant load support transitions from elastic stresses within the plate to increased plate flexure and compensation by local isostatic mechanisms. Plate flexure becomes more pronounced at progressively shorter wavelengths with decreasing $T_e$.

Since the amount of flexure, or deflection, of the continental lithosphere cannot be observed directly, it must be determined from gravity modeling and seismic data. The relatively
few, deep crustal seismic profiles that exist offer limited coverage, but provide essential constraints and boundary conditions for the flexural models. Of the two general modeling methods most applicable to the Northern Rocky Mountain region, coherence and forward modeling, the Maximum Entropy Spectral Estimation (MESE) coherence method of Lowry and Smith (1994) and the forward gravity modeling techniques of Stewart and Watts (1997) provide the highest resolution. In order to facilitate comparisons with the region just south of this study area, previously modeled by MESE coherence (Lowry and Smith, 1995), and because of its inherent ability to resolve $T_g$ equally along any azimuthal direction, the MESE coherence method of Lowry and Smith (1994) was chosen for this study.

**Coherence: Theory**

Spectral methods for modeling flexural rigidity exploit the relationship (i.e. correlation, or lack of it) between gravity and topography as a function of wavelength. In essence, spectral methods seek to predict at what wavelengths gravity can be explained by topography. Since the Bouguer gravity anomaly reflects subsurface mass distributions, it should correlate well with topography if crustal loads are isostatically compensated locally, but show little correlation with topography if crustal loads are partially supported regionally by stresses within a flexurally rigid elastic plate. Each spectral method chooses a way to define this statistical relationship between gravity and topography, assumes a specific isostatic model, and then uses these to solve for the loads and load response. Spectral models differ from each other by their ability to accommodate different loading depths, their use of either the Bouguer or free air gravity anomaly, and the methods they use to compute power spectra in the Fourier domain.

Since sinusoids govern the general solution to the elastic plate flexure equation (1), spectral models assume periodic loading, and the relationship between gravity and topography can be described in the spectral domain. Early spectral methods modeled the admittance, the Fourier transform of a linear filter operating on the topography that best predicts the gravity field in a least squares sense (Dorman and Lewis, 1970):

$$B(k) = Q(k) \cdot H(k)$$

(3)

where $Q$ is the linear isostatic response function (admittance), $B$ and $H$ are the Fourier amplitudes of the Bouguer gravity anomaly and topography, respectively, and $k = (k_x^2 + k_y^2)^{1/2}$ is the two dimensional wavenumber:

$$k = \left( \frac{2\pi}{\lambda_x}, \frac{2\pi}{\lambda_y} \right).$$

(4)
In order to avoid bias by noise, the observed admittance is traditionally estimated from

$$Q(k) = \frac{\langle B(k) \cdot H(k)^* \rangle}{\langle H(k) \cdot H(k)^* \rangle}$$

(5)

where the angle brackets indicate averaging over discrete wavenumber bands, \( \bar{\kappa} \) is the average wavenumber for the waveband, and the asterisk indicates the complex conjugate (Forsyth, 1985).

The flexural rigidity is determined by comparing the observed admittance to theoretical values of \( Q(k) \) calculated from the particular isostatic response model assumed. Most admittance studies have assumed a laterally homogeneous density distribution and modeled deformation as being due to topographic loads alone (e.g. McKenzie and Bowen, 1976; McNutt and Parker, 1978; McNutt, 1980). As Forsyth (1985) has shown, the admittance method has difficulty discriminating between areas of uniform and non-uniform flexural rigidity. Also, the presence of large residual isostatic anomalies from subsurface loads can lead to significantly underestimated values of \( T_e \). Perhaps more importantly, by its very definition, the observed admittance (equation 5) weights the average isostatic response from different regions by factors proportional to the square of the amplitude of the topography. This gives undue emphasis to areas of high topographic relief which often have lower values of \( T_e \) than adjoining, undeformed shield areas (Forsyth, 1985).

Forsyth (1985) developed the coherence method to address these biases. Coherence is defined as the square of the correlation between gravity and topography. The correlation of two functions is described mathematically by

$$\text{Corr}(b, h) = \int_{-\infty}^{\infty} b(X + x) h(X) \, dX$$

(6)

In the Fourier domain, the correlation is simply given by \( B(k) \cdot H^*(k) \), where \( B \) and \( H \) are the Fourier amplitudes of \( b \) and \( h \), respectively. The coherence is a measure of the fraction of the gravity field that can be predicted from the topography using the admittance (Lowry and Smith, 1994). Similar to admittance, the flexural response is assumed to be isotropic (although in this model, the results reflect when this is a bad assumption), uncorrelated noise processes are reduced by averaging spectra within annular wavenumber bins (denoted by the angular brackets), and the coherence is reduced to one-dimension:

$$\gamma^2_o(\bar{\kappa}) = \frac{\langle B(k) \cdot H(k)^* \rangle^2}{\langle H(k) \cdot H(k)^* \rangle \langle B(k) \cdot B(k)^* \rangle}$$

(7)
The coherence ($\gamma^2$) varies between zero, when stresses within the elastic plate completely support applied loads and there is no discernible correlation between gravity and topography, and one, when elastic stresses within the plate can support none of the load, the load must be completely compensated locally (e.g. Airy isostasy), and gravity and topography are completely correlated. The wavelength at which the transition between correlated and uncorrelated occurs is strongly dependent on $T_e$. Since flexurally stronger plates can internally support longer wavelength loads, the transition from uncorrelated to correlated will occur at longer wavelengths with increasing $T_e$ (figure 8).

![Figure 8](image_url)

**Figure 8.** (a) Dependence of coherence on $T_e$. As the flexural rigidity, or effective elastic thickness of the plate increases, larger wavelength loads can be supported by stresses within the plate and the transition from incoherent (coherence close to zero) to coherent (coherence close to one) occurs at larger wavelengths. Coherence values computed using the MESE coherence method of Lowry and Smith (Lowry and Smith, 1994) for the window centered on the position marked with an “M” in figure 8. (b) Residual error plot for the 400 x 400 km window for which $T_e$ was calculated in (a). The predicted coherence is calculated for an initial set of seven $T_e$ values from 4.0 to 128.0. The best-fit log-hyperbolic error function is then calculated from the resulting L1 error norm of the observed and predicted coherence functions as a function of $T_e$. The error function minimum, subsequently recomputed after each iteration, adaptively guides the grid search for the best-fitting $T_e$. Note: The best-fit coherence for this window (19 km) is not illustrated in (a).
Theoretically, coherence is capable of modeling applied loads at the surface and any number of subsurface horizons. In practice, multiple loading depths are difficult to determine, are often physically unrealistic (Bechtel, 1989), and have a negligible effect on the modeled $T_e$ (Forsyth, 1985). Coherence studies usually limit the number of subsurface loading horizons to one (e.g. Forsyth, 1985; Lowry, 1994, Hartley et al., 1996) or two (Bechtel, 1989, Zuber et al., 1989). In this study, a single subsurface loading horizon was assumed.

For the correlation between gravity and topography to accurately predict plate strength, it is also important to determine the distribution and magnitude of the subsurface load as well as the distribution and magnitude of the topographic load. For computational convenience, both topographic and subsurface loads are modeled as harmonic relief on the loading horizon. Each load results in a deflection of both the top of the plate and the subsurface loading horizon (figure 9), which can be predicted from the thin plate equation (1), dependent on an assumed flexural

\[ A. \text{ Initial loading} \quad B. \text{ Equilibrium state} \quad C. \text{ Observed deflections} \]

\[ \begin{align*}
H_i & \quad \rightarrow \quad H_T \\
& \quad \leftarrow \quad W_T \\
W_i & \quad \rightarrow \quad H_B \\
& \quad \leftarrow \quad W_B \\
\end{align*} \]

**Figure 9.** Plate flexure with both surface and subsurface loading. Black arrows show how initial loads (a) cause equilibrium deflections (b) which sum to give the observed surface and subsurface deflections. Specifically, the weight of an initial topographic load with amplitude $H_i$ will deform the plate, leading to surface relief with amplitude $H_T$ and subsurface relief with amplitude $W_T$. The weight of an initial subsurface load with amplitude $W_i$ will also deform the plate and result in surface relief with amplitude $H_T$ and subsurface relief with amplitude $W_B$. When both surface and subsurface loads are present, the amplitude of the observed topography is the sum of $H_T$ and $H_B$. Similarly, the amplitude of the subsurface "topography" is the sum of $W_T$ and $W_B$. The subsurface "topography" is a convenient way to represent subsurface density variations which can be modeled by the Bouguer anomaly. Pink arrows indicate how these relationships can be modeled in the reverse direction. If the plate rigidity is known (or assumed), the thin plate equation, coupled with boundary conditions from the topography and Bouguer anomaly, allows separation of the components of relief on each interface and a determination of the original loads.
rigidity and the wavelength of the load. The total deflection of the plate at both the surface (H) and subsurface (W) loading horizons is just the sum of the contribution from the top (T) and bottom (B) loads:

\[ H = H_T + H_B \]
\[ W = W_T + W_B \]

(8)
(9)

H can be determined directly from the measured surface topography, and W can be modeled from the downward continuation of the Bouguer gravity anomaly, which arises solely from subsurface density variations:

\[ W(k) = B(k) \exp(kz) / 2\pi\Delta \rho G \]

(10)

where G is the gravitational constant, z is the subsurface load depth, and \( \Delta \rho = \rho_1 - \rho_0 \) is the difference in density above (\( \rho_1 \)) and below (\( \rho_0 \)) depth z. The components of relief at the top and subsurface interfaces must be modeled from the thin plate equation (1). If the initial relief at the top (before flexure and compensation) is

\[ H_i = H_T - W_T \]

(11)

then for an initial load \( \rho_0 g(H_T - W_T) \), the amplitude of relief induced at depth z (from equation (1)) will be

\[ W_T(k) = -\rho_0 H_T(k) / \Delta \rho \zeta \]

(12)

where

\[ \zeta = 1 + D k^4 / \Delta \rho g. \]

(13)

Similarly, initial relief on the subsurface interface

\[ W_i = W_B - H_B \]

(14)

gives rise to an initial load \( \Delta \rho g(W_B - H_B) \) which will produce an additional component of relief at depth z, given by

\[ W_B(k) = -\rho_0 H_B(k) \phi / \Delta \rho \]

(15)

where

\[ \phi = 1 + D k^4 / \rho_0 g. \]

(16)

Equations (8), (9), (11), (12), (14), and (15) can be combined to yield

\[ H = H_1 \left[ \frac{\Delta \rho \zeta}{\rho_0 + \Delta \rho \zeta} \right] - W_1 \left[ \frac{\Delta \rho}{\Delta \rho + \rho_0 \phi} \right] \]

(17)

\[ W = -H_1 \left[ \frac{\rho_0}{\rho_0 + \Delta \rho \zeta} \right] + W_1 \left[ \frac{\rho_0 \phi}{\Delta \rho + \rho_0 \phi} \right] \]

(18)
which can be solved for the initial loads. If the determinant of the matrix of coefficients from equations 17 and 18 (i.e., $\xi - 1$, in reduced form) equals zero, then a unique solution does not exist and the initial loads cannot be solved for. This will occur when $D = 0$, the case of perfect Airy isostasy. For any case where the plate does possess some internal strength, equations 17 and 18 can be solved for the initial loads $H_i$ and $W_i$ in terms of the known or inferred quantities $H$, $W$, $\rho_0$, $\rho_1$, $z$, and an assumed $D$.

$$H_i = \left( H + W \frac{\Delta \rho}{\rho_0 \xi} \right) \left( \frac{\rho_0 + \Delta \rho \xi}{\Delta \rho} \right) \left( \frac{\phi}{\phi - 1} \right)$$ \hspace{1cm} (19)

$$W_i = \left( W + H \frac{\rho_0}{\Delta \rho} \right) \left( \frac{\Delta \rho + \rho_0 \phi}{\rho_0} \right) \left( \frac{\xi}{\phi - 1} \right)$$ \hspace{1cm} (20)

$H_T$ can then be determined from equations 11 and 12, $H_B$ from equations 14 and 15, and finally, $W_T$ and $W_B$ from equations 11 and 14, respectively.

The predicted coherence, $\gamma_p^2$, can now be found from equation 7, where $B$ is related to the subsurface relief by equation 10 and $H$ is now due to both surface and subsurface loading (equation 8). If the surface and subsurface loading are statistically independent, the resulting equation reduces to

$$\gamma_p^2(k) = \frac{\langle H_T(k) W_T(k) + H_B(k) W_B(k) \rangle^2}{\langle H_T(k)^2 + H_B(k)^2 \rangle \langle W_T(k)^2 + W_B(k)^2 \rangle}$$ \hspace{1cm} (21)

which can be solved for different values of $D$ and compared to the observed coherence (equation 7). The value of $D$ which minimizes the difference between predicted and observed coherence determines the best-fit $T_e$ (equation 2). If surface and subsurface loads are statistically correlated, there will still be some coherence between gravity and topography even when loads are completely supported by the strength of the plate. This moves the coherence curve transition region to shorter wavelengths and results in an underestimation of $T_e$ (Macario et al., 1995).

**MESE Coherence**

As defined by Lowry and Smith (1994), the maximum entropy spectral estimation (MESE) coherence is determined from

$$\gamma_o^2(k) = \left| \frac{(B(k) \cdot H(k)^*)^2}{(H(k) \cdot H(k)^*)(B(k) \cdot B(k)^*)} \right|$$ \hspace{1cm} (22)
where the method of averaging over discrete wavebands is slightly different than the original formulation (equation 7) of Forsyth (1985). To improve the resolution and accuracy of the traditional coherence method, Lowry and Smith (1994) made several modifications to the technique.

Lowry and Smith's (1994) most significant change was to replace the traditional fast Fourier transform periodogram method of estimating the power spectra in the coherence equations with a computationally efficient algorithm (Lim and Malik, 1981) for estimating power spectra by the maximum entropy method. Because geophysical data is not truly periodic, and computations must be performed over a finite window, Fourier methods can produce spectral artifacts which can partially mask, or interfere with the true spectral signature of the data. Convolution with a broad, finite window can lead to spectral bias (a smoothing of the spectrum which can result in the loss of fine detail, especially adjacent to major spectral peaks) and non-periodicity can lead to leakage (a widening of the base of primary spectral components not harmonically related to the window size). Coherence studies employing the periodogram method artificially create periodicity in the data by some form of mirroring. Data from the area of interest is reflected outward to create a larger grid where the data values on each edge exactly match the data values on the opposite edge. The spectral resolution of the periodogram method is fairly low and large data windows must be used to confidently resolve the transition waveband. MESE also attempts to reduce spectral artifacts by creating a reasonable extrapolation of the data beyond the window of interest, but accomplishes this in an entirely different way. The MESE algorithm of Lim and Malik (1981) uses a statistical approach to estimate the least biased power spectrum over an implicitly extrapolated area. For a given two dimensional signal, \( t(x,y) \), the power spectrum is estimated by iteratively searching for the extrapolation of the signal's known autocorrelation function which has the maximum entropy:

\[
E = \int\int_{-\pi}^{\pi} \log[\hat{P}_t(k_x,k_y)]dk_xdk_y. \tag{23}
\]

\( E \) is the Shannon entropy, and \( \hat{P}_t \) is an estimate of the power spectrum of \( t(x,y) \) over the extrapolated area. The entropy is maximized when

\[
F^{-1} \left\{ \frac{1}{\hat{P}_t(k_x,k_y)} \right\} = 0 \tag{24}
\]

for all \( x, y \) in the extrapolated area, where \( F^{-1} \) represents the inverse fourier transform. This
method achieves higher spectral resolution and allows flexural rigidities to be determined within smaller data windows.

Strictly speaking, Lowry and Smith (1994) developed an ad hoc coherence estimation technique in that the method of Lim and Malik (1981) applies to the determination of auto-power spectra, but not cross-power spectra. Lowry and Smith (1994) adapted the method to the determination of both, but the maximum entropy criteria is not strictly met in the case of cross-power spectra (Ulrych and Jensen, 1974; Lowry, 1994). Nevertheless, extensive simulations of the resultant MESE coherence function and comparisons to estimates from periodogram-based coherence methods have demonstrated the improved accuracy and resolution of the method (Lowry, 1994). The maximum entropy method is especially good at fitting sharp spectral features (Press et al., 1992), can adequately resolve the transition waveband in the coherence function with much smaller windows (Lowry, 1994), and thus can begin to provide discrimination of $T_e$ over individual tectonic regions.

**Application of MESE coherence to the Northern Rocky Mountains**

A flexural rigidity map for most of Montana and Idaho, along with portions of surrounding states and Canadian provinces (figure 10) was produced from coherence functions (e.g. figures 8 and 11) calculated using the methodology of Lowry and Smith (1994). Once program parameters were experimentally determined within a given region (as discussed in Appendix A), each coherence estimate took from one to three hours to compute on a DEC Personal Workstation 500au. Topography and Bouguer anomaly data on which the calculations were based were taken from the identically parameterized grids depicted in figures 1 and 2. Both flexural rigidity ($D$) and effective elastic thickness ($T_e$) estimates were obtained for 200 x 200, 400 x 400, or 560 x 560 km windows with centers spaced at 48 km intervals throughout the 912 km (north-south) by 1008 km (east-west) grid (area outlined in figures 1 and 2). Sample 200 x 200 and 400 x 400 km windows in figure 10 illustrate the resolution of the coherence calculations and the large overlap between adjoining windows. To create the map, rigidity estimates were interpolated to an 8 km spacing and contoured with a bicubic interpolation algorithm. Assumed values for the physical parameters relating $D$ and $T_e$ (equation 2) are given in Table 1.

The variable window sizes used in this study reflect the balance between maximizing the resolution of $T_e$ and the necessity of choosing large enough spectral windows to identify the wavelength at which the integrated strength of the crust can no longer support the applied loads. The increased spectral resolution provided by the maximum entropy algorithm enables observed
Table 1. Physical parameters used in coherence calculations.

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Definition</th>
<th>Value</th>
</tr>
</thead>
<tbody>
<tr>
<td>E</td>
<td>Young's modulus</td>
<td>$1 \times 10^{11}$ N/m²</td>
</tr>
<tr>
<td>v</td>
<td>Poisson's ratio</td>
<td>0.25</td>
</tr>
<tr>
<td>g</td>
<td>acceleration of gravity</td>
<td>9.8 m/s²</td>
</tr>
</tbody>
</table>

and predicted coherence functions to be calculated within relatively small (200, 400, and 560 km) window sizes. However, the maximum entropy method applies only to the estimation of power spectra and these are not involved in the wavenumber domain deconvolution of surface and subsurface loads (equations 8-20). To ensure adequate resolution of long wavelength relief at the loading interfaces, the deconvolutions were performed over much larger (1116 x 1116 km) data windows. $H_T$, $H_B$, $W_T$, and $W_B$ were then inverse transformed back to the spatial domain where a window identical in size and location to that used in calculating the observed coherence was then extracted for use in estimating the predicted coherence. As long as the coherence window size is large enough to image the top of the transition waveband, the inclusion of long wavelength load information in the calculation of $H_T$, $H_B$, $W_T$, and $W_B$ should lead to a more accurate estimate of the flexural rigidity based on the response of the plate to loads at all wavelengths.

Optimal coherence and deconvolution window sizes were determined experimentally. (Results from a number of sensitivity tests are discussed in the Appendix.) In general, if the flexural rigidity averaged over the larger window used for deconvolution was greater, a larger window was required for estimation of the observed and predicted coherence functions. While use of a large deconvolution window does minimize windowing effects, it also makes the predicted coherence more sensitive to long wavelength relief at the modeled interfaces than the observed coherence. If the surface to subsurface load ratio is different at long wavelengths, this will introduce a long wavelength dependence into the selection of the best fit $T_c$. To minimize the contribution of this effect in locales with highly different $T_c$ across the grid, the deconvolution window size was kept constant at 1116 x 1116 km for computations throughout the study region.

While the resolution of the coherence estimates does not necessitate detailed local geology to be modeled, regional differences in structure (which cover an area equal to or larger than the window size for a single coherence calculation) are factored into the model where possible. One of the minor modifications made to the coherence model by Lowry and Smith (1994) was the incorporation of a more sophisticated density model than the one or two layer models that had been used previously (e.g. Forsyth, 1985; Bechtel et al., 1990; Pilkington, 1991;
Figure 10. Effective elastic thickness of the Northern Rocky Mountains. Small boxes are 48 km apart and represent centers of 200 x 200 km (smallest boxes), 400 x 400 km (medium size boxes), and 560 x 560 km windows (largest boxes in SE corner) used to estimate $T_e$. The two larger boxes near the top illustrate the size of a 200 x 200 km window (in the top left) and a 400 x 400 km window (in the top right). A, B, C, and D locate the window centers for the sample coherence functions plotted in figure 11. M locates the coherence function plotted in figure 8. Physiographic provinces (modified from Zoback and Zoback, 1989) are outlined in white and labeled (large italicized letters) as defined in figure 12.
Hartley et al., 1996). Lowry (1994) showed that the load deconvolution (which is dependent on \( \rho \) as well as \( H, B, \) and \( D \)) can be accomplished using any one-dimensional density model. Where reliable seismic P-wave velocities can be determined from deep crustal seismic refraction profiles, a more detailed model of density variation with depth can be estimated.

In this study, the density-depth relationship is determined from the nonlinear velocity-density regression line parameters empirically determined by Christensen and Mooney (1995) based on their extensive high-pressure laboratory measurements of seismic velocities in a wide variety of rock types. The study area was broken into seven physiographic provinces (figure 12) and deep crustal seismic refraction profiles which crossed the region (figure 13) were used to
Figure 12. The seven physiographic provinces of the study area (adapted from Zoback and Zoback, 1989). A different crustal density model is derived for each province as input to the coherence calculations. Provinces are: CP = Columbia Plateau, NRM = Northern Rocky Mountains, NSRP = area immediately north of Snake River Plain, SRP = Snake River Plain, MRM = Middle Rocky Mountains, USGP = United States Great Plains, CGP = Canadian Great Plains.

Figure 13. Index map showing locations of deep crustal seismic refraction profiles used to constrain P-wave velocities from which the density models for the seven selected physiographic provinces were calculated. References for each named profile are given in the caption of figure 14.
Figure 14.
Seismic velocity models for the seven physiographic provinces of the Northern Rocky Mountains outlined in figure 12. Depths are in meters, velocities are in m/sec. The velocity discontinuity chosen as the load depth is represented by the bold dashed line.

Profiles compiled from the seismic refraction studies of:
1 McCamy and Meyer (1964), Meyer, Steinhart, and Bonini (1961) [NR1, NR2, GP1]
   Hales and Nation (1973) [NR3, NR4]
   Johnson and Couch (1970) [NR5, NR6]
   White and Savage (1965) [NR7]
2 Snelson, et al. (1998) [GP3]
   Richards and Walker (1959) [GP4]
   White and Savage (1965) [NR7]
3 Catchings and Mooney (1988) [CP1]
   Hill (1972) [CP2] and Dehlinger et al. (1965)
4 Snelson, et al. (1998) [GP3]
   McCamy and Meyer (1964), Meyer, Steinhart, and Bonini (1961) [GP1, GP2]
5 Sheriff and Stickney (1984) [NSR1]
6 Sparlin et al. (1982), Braile et al. (1982) [SRP1, SRP2]
7 Snelson, et al. (1998) [MRM1] and Braile et al. (1974)

Designations in [ ] identify seismic lines on figure 13.
construct a model of the distribution of P-wave velocities with depth (figure 14) within each of the seven provinces. Subsurface load depths (figure 14) were chosen based on Lowry and Smith’s (1995) analysis of the relationship between $T_e$ and heat flow for the eastern Snake River Plain, northern Rocky Mountains, and middle Rocky Mountains, and the correlation of this relationship with a range of crust and mantle lithologies. Their analysis implied primary accommodation depths in the middle to upper crust for the Snake River Plain and northern Rocky Mountains, and near the crust-mantle boundary for the middle Rocky Mountains. In fact, Lowry and Smith (1994, 1995) chose the shallowest, first-order velocity discontinuity as their subsurface loading depth in these regions. Based on their analysis, and to facilitate comparison of $T_e$ values determined from this study and the study of Lowry and Smith (1995), shallow subsurface loading depths were assumed for most of the physiographic provinces shown in figure 12. Based on Lowry and Smith’s (1995) estimation of a deeper accommodation depth for the middle Rocky Mountain region, both middle and lower crust subsurface load depths were modeled for a number of windows in this area. The mid-crust load depth produced the most robust results (see the Dependence on load depth section in Appendix A) and was therefore adopted for coherence calculations within this region. Although some care was taken to include physically realistic estimations of the subsurface load depth in this study (figure 14), detailed sensitivity analysis has in fact shown that even sizeable errors in loading depth have a small effect (a factor of no more than two in flexural rigidity and thus only a factor of $~1.3$ in $T_e$) on the determination of $T_e$ (Forsyth, 1985; Lowry, 1994).

Residual errors for the flexural rigidity estimates were calculated by subtracting the observed coherence from the predicted coherence calculated from the best-fit $T_e$. To determine the reliability of the coherence estimates, the residual error of the best-fit $T_e$ within each calculation window was interpolated and plotted (figure 15) in a manner identical to that used to produce the flexural rigidity map (figure 10). Unfortunately, the rough correlation between high residual error and high flexural rigidity (which can be seen by comparing figures 10 and 15) is an artifact of an enforced wavenumber cutoff in the calculation of the residual error (see Dependence on the wavenumber cutoff in the error function determination in Appendix A for details). The empirically determined, minimum expected standard deviation of the MESE coherence estimates is approximately 0.03 (Lowry, personal commun.). Local variations in the residual error by more than this amount probably indicate some failure of the modeling assumptions, such as correlation of surface and subsurface loads or improper choice of program parameters (see Appendix A). In general, the coherence estimates for most of the intermountain region appear reliable. However,
Figure 15. Residual error for the $T_e$ estimates of figure 7. (Predicted coherence calculated from the best-fit $T_e$ minus the observed coherence, averaged over all wavelengths above a predetermined wavelength limit (see Appendix A for details).) The interpolation method is identical to that used in figure 7.

Local variations in the residual error do exceed 0.03 within the southern Columbia Basin, in part of the fold and thrust belt in the Canadian Rockies, and in a number of small, isolated areas in the Great Plains. The $T_e$ values estimated in these places may be less accurate.

**Results**

The southern third of this study area overlaps the northern third of Lowry and Smith’s (1995) coherence analysis of the southern Rocky Mountains/Basin and Range area (figure 16) and even though there are small differences in some of the model input parameters, there is excellent agreement in the flexural rigidities determined within the overlap area. In this study, $T_e$ values vary from a low of 4.5 km in the western Snake River Plain to a high of 53.8 km over the Selkirk fan structure (an asymmetric structural fan possibly formed by the detachment and obduction of the upper crust of an accretionary terrane) located beneath the Columbia Mountains.
Figure 16. MESE coherence estimates of flexural rigidity for the middle Rocky Mountains region (from Lowry and Smith, 1995). Windows are 200 x 200 km (small boxes) or 400 x 400 km (large boxes) and spaced 50 km apart. (Note that color scale is somewhat different from figure 9.) Physiographic provinces are: CB=Columbia Basin, NRM=northern Rocky Mountains, MRM=middle Rocky Mountains, SRP=Snake River Plain, BR=Basin and Range, CP=Colorado Plateau.
of southern British Columbia. $T_e$ is low throughout the eastern and western Snake River Plain, as well as around the eastern edge of the Columbia Basin. Low values of $T_e$ also continue northward from the eastern end of the Snake River Plain (around the area of the Madison River valley) and then turn westward toward the Montana-Idaho border along a line through Butte. Relatively low values of $T_e$ finger northeastward through the Montana Plains. High values of $T_e$ predominate elsewhere in the Great Plains and extend over some of the mountainous areas of northwestern Wyoming as well. Isolated regions of high $T_e$ occur over the Columbia Basin and the inferred location of the Beaverhead meteorite impact site (McCafferty, 1995) just north of the Snake River Plain in south-central Idaho.

Crustal strengths show significant correlation to inherited crustal structure, most notably to fundamental crustal boundaries between some of the older terranes (figure 17). Not surprisingly, deviations from inherited trends show up most strongly in areas of recent crustal restructuring, such as the Columbia Basin and along the Yellowstone hotspot track, where the crust is thought to have been exposed to abnormally high asthenospheric temperatures (Catchings and Mooney, 1988; Lowry et al., 2000). Flexural rigidities (both from this study and Lowry and Smith, 1995) also show interesting correlations with mantle velocity structure (figure 18). Detailed comparisons of $T_e$ with the underlying mantle, inherited crustal boundaries, structural provinces, and important geologic features of the area are examined in the following sections.

**Interpretation and discussion**

*Correlation of flexural rigidity with lithospheric age*

Because flexural rigidity is strongly correlated with lithospheric temperature and composition (Burov and Diament, 1995; Lowry and Smith, 1995), it should show significant correlation to lithospheric age in areas which have not been thermally reset by more recent activity. Burov and Diament (1995) modeled the thermal state of the lithosphere as a function of age (using a cooling model which incorporated heating from radioactive decay in the crust and viscous friction at the crust-mantle boundary) and found a relatively rapid deepening of lithospheric isotherms between a thermal age of zero and 200 million years. By 400-500 million years, the lithosphere cooled to an equilibrium state. To quantify the relationship between strength and thermal age, Burov and Diament (1995) calculated the dependence of $T_e$ on thermal age and crustal thickness. They found that $T_e$ increases significantly slower than the rate of lithospheric cooling. Their results suggest that approximately one billion years are needed for $T_e$ to increase to an equilibrium value, regardless of crustal thickness. Not surprisingly, modeling
Figure 17. Inferred tectonic domains (from figure 4C) superimposed on effective elastic thickness estimates of the Northern Rocky Mountains (from figure 10).

studies have also shown that $T_e$ relates to lithospheric composition. Calculated values of $T_e$ are higher for drier, more mafic rocks, and lower for wetter, more felsic rocks (Burov and Diament, 1995; Lowry and Smith, 1995). Perhaps equally importantly, rock composition can also affect the geotherm, and thus indirectly influence $T_e$ (Lowry and Smith, 1994).

A number of flexural rigidity studies support these conclusions. $T_e$ shows a broad correlation with age in North America (Bechtel et al., 1990), the South American Andes (Watts et al., 1995), Africa (Hartley et al., 1996), and Australia (Zuber et al., 1989), especially within the continental shields. Given the very long time to reach equilibrium postulated for $T_e$, it is not surprising that old cratonic areas have flexural rigidities commensurate with their age. However, fundamental rheological differences between Archean and Proterozoic crust (Karlstrom and Humphreys, 1998) may also play a role. These differences may inhibit thermal reactivation and
Figure 18. Mantle velocity structure at 100 km depth. (From Karlstrom, 1999, as modified from Humphreys and Dueker, 1994, and Grand, 1994.) Yellow, orange, and brown colors denote low velocity, hot mantle with a few percent partial melt. Blue colors indicate high velocity, cold mantle. Cold mantle to the east is Proterozoic in age, hot mantle to the west is Cenozoic in age (Karlstrom, 1999). Along the Rocky Mountains, old, northeast trending shear zones between the Archean and Proterozoic terranes are providing conduits for the migration of hot, young mantle material from the west into the old, cold, mantle material to the east. These areas correlate especially well with east-northeast extensions of relatively low flexural rigidities in Utah and Idaho (figures 9 and 14).
subsequent decreases in $T_e$. In younger provinces, the distribution of $T_e$ often varies widely within the same crustal domain. While this may reflect the proximity of these areas to plate boundaries and the complex deformational processes that occur there, compositional differences within, or between these younger terranes may be important as well.

The $T_e$ results of this study also show a correlation between flexural rigidity and lithospheric age. The large region of high $T_e$ (> 20 km) which underlies much of the eastern portion of the study area lies within the Archean crust of the Wyoming and Hearne provinces (figure 17). $T_e$ decreases to a maximum of 11-16 km within a zone roughly coincident with the location of Proterozoic crust. The lowest $T_e$ values lie close to the $^{87}\text{Sr}/^{86}\text{Sr} = .706$ isopleth (figure 4C) which is thought to roughly delineate the mid-Paleozoic continental margin (Oldow et al., 1989) and presumably lies close to the late Proterozoic rifted edge of North America. Low $T_e$ values also occur along and within the Yellowstone hotspot track where the crust has been extensively heated (Morgan and Gosnold, 1989), extended, and in some places intruded with mafic material from the anomalously warm mantle below (Sparlin et al., 1982). Restructuring of the lithosphere due to passage of the Yellowstone plume has essentially reset the thermal age and flexural rigidities of this region.

**Localized areas with high flexural rigidities**

Locations within the study area which do not seem to correlate directly with crustal impact site, and the low $T_e$ values which extend northward from the southwest Montana/northwest Wyoming border.

**The Columbia Basin**

The high $T_e$ values of the Columbia Basin are especially problematic. Within the Columbia Basin, values of high $T_e$ correspond to a strong regional high in the gravity field. A seismic refraction profile across the basin indicates that this gravity high coincides with the dramatic upward thickening (from ~36-40 km depth at the basin edge to ~24-41 km depth at the basin center) of a deep, high velocity crustal layer which lies beneath a sediment filled graben underlying the basalts in the central part of the basin (Catchings and Mooney, 1988). Since much of the Columbia Basin area is characterized by factors which usually lead to a reduction in $T_e$ (i.e. relatively young underlying crust, high heat flow (Morgan and Gosnold, 1989) and recent rifting (as may be evidenced by the seismic data, Catchings and Mooney, 1988)), the high values of $T_e$ seem anomalous. Because the coherence technique assumes that surface and subsurface loads are uncorrelated, it is likely that the flexural rigidities calculated for this area are inaccurate. This is
strongly confirmed by the high residual errors which characterize these coherence estimates (figure 15). Unfortunately, Macario et al., have shown that correlation between surface and subsurface loads leads to a downward bias in the estimated $T_e$. Perhaps the laterally extensive, overlying basalt flows diminish the topographic signature of the area and lead to a lack of correlation between the topography and gravity signals at longer wavelengths. It is also possible that the 200 x 200 km windows used to estimate the coherence functions in this area were not long enough to resolve the longer wavelengths, leading to an upward bias in the estimates of $T_e$ (Macario et al., 1995). Alternatively, structural and compositional changes in this area relating to the creation of an extensive flood basalt province may actually have led to an increase in crustal strength. Some support for this hypothesis may be found in the isolated area of cold, high velocity mantle which almost exactly underlies the area of high $T_e$ (figure 18).

The Beaverhead Meteorite impact site

The $T_e$ estimates (figure 10) also reveal a few locations where there is a significant variation in $T_e$ over the size of one or two coherence windows. While it is tempting to ignore these aberrant values as unrealistic, the area of locally high $T_e$ at a longitude of 114° west, latitude 44.4° north occurs in a provocative locale. This strong, highly localized signal is directly centered over the area proposed as the site of the approximately 600 Ma Beaverhead impact structure (McCafferty, 1995). While a relationship to the impact site is highly speculative and would be difficult to prove, it does provide a unique feature of about the right size (60-75 km in diameter, as indicated by an annular gravity low) (Hargraves et al., 1990; Hargraves et al., 1994) in exactly the right place. Impact modeling studies and scaling relationships between crater size and the amount of structural uplift (Pilkington and Grieve, 1992) suggest that heavily fractured rocks may have extended almost 7 km beneath the surface. Given that shatter cones have been found on the Cabin thrust plate of the Medicine Lodge thrust system, and the Medicine Lodge thrust system exists at relatively shallow depths, with the Precambrian Yellowjacket formation directly below (Ruppel, 1978), a large fraction of the crust affected by the impact probably still exists in its original location. The density of material in this location should be lower than the surrounding area (leading to the gravity low noted above) due to fracturing and brecciation of the target rocks (Pilkington and Grieve, 1992) and would show no correlation to existing topographic features. This could result in a lack of correlation between the topographic and Bouguer anomaly signals at significantly larger wavelengths than the surrounding area, and hence, an increase in the estimated $T_e$.

While other local gravity anomalies do occur in the area, none seem to offer a convincing
alternative to the impact hypothesis. The deep, sediment filled graben of the Lemhi valley occurs a bit to the east of the high \( T_e \) feature and two caldera complexes from the Eocene Challis volcanics (Hobbs, *et al.*, 1991) occur at its northern edge. However, they seem unlikely sources for the \( T_e \) high since similar features occur elsewhere in the region without influencing the generally low values of \( T_e \) which characterize the area. Similarly, the Roberts Mountains allochthon occurs just to the west, but it continues to the north and far to the south of the area of high \( T_e \). It is interesting to note that, similar to the Columbia Basin, the small, isolated area of rigid crust which correlates with the proposed impact site is underlain by a relatively isolated area of anomalously cold mantle (figure 18).

**Correlation of flexural rigidity with Archean and Proterozoic tectonic province boundaries**

**The Great Falls Tectonic Zone**

O'Neill and Lopez (1985) hypothesize that the GFTZ is an early Proterozoic shear zone between the Hearne and Wyoming provinces which has remained a zone of crustal weakness into Holocene time. If this crustal boundary is indeed a zone of weakness, flexural rigidities in this area should be low. This study strongly supports their hypothesis.

One of the most intriguing features of the \( T_e \) map (figure 17) is a north-eastward extension of comparatively low \( T_e \) (\(~ 10-20 \) km) into the eastern Montana Plains. This feature is remarkably close to the position of the GFTZ as inferred by prominent, northeast trending maxima in the horizontal gradient of pseudogravity (figure 7) and by studies of O’Neill and Lopez (1985) and Boemer *et al.* (1998). The continuity of this feature (figure 17), and the low residual errors associated with the \( T_e \) estimates in this area (figure 15), lend confidence to its significance.

This area of weak crust seems a likely position for the boundary between the Archean Hearne and Wyoming provinces. Price and Sears (2000) infer that the northwest edge of the Archean Wyoming province is bounded by an active fault-line scarp along the southern edge of the Central Montana Trough and the east-northeast trending arm of the intracontinental Middle Proterozoic rift which led to the formation of the Belt Basin (figure 19). However, although the Central Montana Trough has apparently been periodically reactivated as an area of subsidence from the Precambrian through Mesozoic time (Peterson, 1986), it is considered to have a much more easterly trend (extending almost due east from the its labeled, western end in figure 19) than the area of crustal weakness indicated by the flexural rigidity estimates. Furthermore, the boundaries of the Central Montana Trough have virtually no magnetic signature (figure 7). It seems more likely that the northwest boundary of the Wyoming province is delineated by the
Figure 19. Proposed features of the Belt Basin superposed on estimated values of $T_e$. Dashed black lines depict the palinspastically restored isopach map of lower Belt-Purcell rocks (Price and Sears, 2000). Grey arrows show the direction and amount of net horizontal displacement between selected points within the palinspastic reconstruction and the location of those points today. White lines delineate the hypothetical Proterozoic fault trends of Winston (1986). These “lines” have been interpreted as growth faults between crustal blocks whose differential subsidence may have produced the Belt Basin (Winston, 1986).

coincidence of strong northeast trending magnetic boundaries (figure 7), weak crust, and the location of the GFTZ (figure 17). It is extremely unlikely that this northeastward extension of relatively weak crust is an artifact of the modeling procedure since two of the largest sources of bias in the coherence method can be specifically addressed. As pointed out by Macario et al. (1995), the largest inherent biases in the coherence method arise from choosing a window size too small to resolve the longest flexural wavelengths in the study area (which leads to an upward bias in the results), or correlations between surface and subsurface loads (which may be especially common along
mountain belts and leads to a downward bias). While a different choice of coherence window size might lead to a slight compression or expansion of the boundaries of the GFTZ (due to a change in the resolution of $T_e$), it is difficult to see how bias due to window size could produce the current results. 400 x 400 km coherence windows were used throughout this portion of the study area and these windows are much larger than the width of the GFTZ (see figure 10). This means that areas of high $T_e$ were included in the estimation of $T_e$ within the GFTZ and any upward bias due to inadequate window size should have affected values both within, and beyond, the GFTZ.

The possibility of correlated surface and subsurface loads is more difficult to dispel. Fortunately, a recent seismic refraction profile from the Deep Probe seismic experiment (Snelson et al., 1998, figure 20) crosses this region along a roughly north-south direction from a longitude of approximately 111° west along the northern boundary of the study area, to approximately 108° west at the southern boundary. Although this profile shows a gradual southward deepening of the Moho from the Vulcan Low through most of the Wyoming province, and a coincident thick lower crustal layer with velocities indicative of underplated gabbroic material (Snelson et al., 1998), it does not image any subsurface crustal anomalies correlative with surface loads. A detailed gravity model through the study area along the 106th meridian shows similar results (Snelson et al., 1998).

It is interesting that no sign of the GFTZ appears in the seismic record, especially considering the apparent periodic reactivation of this feature from middle Proterozoic to Holocene time (O’Neill and Lopez, 1985) and its obvious expression in the $T_e$ data. Snelson et al. (1998) infer that the thick lower crustal layer most likely originated in the Archean and note that a recently discovered Archean analog in the Baltic shield does contain a suture zone. Unfortunately, an exceptionally active tectonic interval from about 1.8-1.65 Ga (Karlstrom and Humphreys, 1998) affected the boundary between the Wyoming and Hearne provinces and rocks within the GFTZ were metamorphically reset to 1.8 Ga. Thus, it is difficult to determine whether the suture zone between these two provinces was created or merely reactivated (Boerner et al., 1998; Mueller et al., 1999) at this time.

**Reactivated crustal boundaries in the northern and central Rocky Mountains**

If the inherited, periodically reactivated crustal boundary between the Archean Wyoming and Hearne provinces is imaged so prominently in the flexural rigidity data, might other inherited crustal boundaries show up as well? An examination of Lowry and Smith’s (1995) flexural rigidity data for the middle Rocky Mountains provides evidence that they do.
Figure 20. Velocity model from the Deep Probe transect (adapted from Snelson et al., 1998). Velocities are in km/sec. Dashed lines show the northern and southern boundaries of the present study area over which \( T_e \) has been estimated. Dotted lines show approximate position of the northeastward extension of low \( T_e \) values into the Montana Plains.

Their results show a similar, though less pronounced correlation between a number of other northeast trending, periodically reactivated Archean and Proterozoic crustal boundaries and lower than average \( T_e \). Specifically, \( T_e \) values decrease dramatically (from > 35 km to ~20 km) across the east-northeast trending Cheyenne Belt, the suture zone between the Archean Wyoming province and accreted Proterozoic terranes to the south. South of the Cheyenne Belt, in central and east-central Utah, \( T_e \) values segment into sections with higher (> 30 km) \( T_e \), and lower (~15-20 km) \( T_e \). The lower \( T_e \) values are roughly coincident with the northeast trending Gneiss Canyon Shear Zone and Crystal Shear Zone which delineate the boundary between the Proterozoic Mojave Province (>1.8 Ga crust incorporated into a 1.75 Ga arc) and Yavapai Province (a 1.76-1.72 juvenile arc terrane)(Karlstrom and Humphreys, 1998).

All of the northeast trending crustal boundaries (the GFTZ, the Cheyenne Belt, Gneiss Canyon Shear Zone, and Crystal Shear Zone) which show up as areas of relatively weak crust in
the T data (from this study and Lowry and Smith (1995)) have in common substantial evidence of periodic reactivation through time (e.g. O’Neill and Lopez, 1985; Karlstrom and Humphreys, 1998; Pazzaglia and Kelley, 1998). At 1.45-1.35 Ga they were all reactivated by a heating event which occurred across the entire region (from Canada to northern Mexico), perhaps due to radiogenic heat built up in the crust during an apparent 200 million year quiescent period following terrane accretion. Northwest contraction and northeast extension (Karlstrom and Humphreys, 1998) led to extensive tectonism and granitic magmatism, with the largest intrusive bodies concentrated along these northeast trending crustal boundaries (Karlstrom and Humphreys, 1998). In Montana, a scattering of small, quartz monzonite plutons were emplaced in the GFTZ during this time (O’Neill and Lopez, 1985). In the Paleozoic and early Mesozoic, isopach maps show recurrent and differential reactivation of these northeast trending shear zones (O’Neill and Lopez, 1985; Tonnsen, 1986; Karlstrom and Humphreys, 1998) to produce local highs, troughs, or breaks in depositional patterns. Laramide-era magmatism shows some correlation as well (O’Neill and Lopez, 1985; Karlstrom and Humphreys, 1998). Within the GFTZ, the Bearpaw intrusives, Highwood intrusives, and Little Rocky Mountains intrusives date from this time. Although the history of these crustal boundaries is undoubtedly incomplete, it seems likely that they represent zones of crustal weakness inherited from the time of continental amalgamation. Structural and compositional differences between these boundaries and the surrounding provinces predispose these regions to reactivation whenever stresses become high enough and properly aligned.

The Vulcan Low

Although many of the inherited, northeast trending crustal boundaries show up in the flexural rigidity data, not all of them do. In the north-central part of the region examined in this study, the Vulcan Low (figure 4C), interpreted as a north-dipping Archean collisional suture between two separate blocks of the Archean Hearne Province (Ross et al., 1991), has no signature at all in the flexural rigidity data. This suture zone was reactivated around 1.8 Ga by deformation resulting from coeval collisional orogens which converged on the Hearne province from both the northwest and southeast (Ross et al., 1991). Additionally, Price and Sears (2000) note that palinspastic restoration of the Belt-Purcell rocks closely aligns the Mesozoic Moyie-Dibble Creek and St. Mary-Lussier River faults with the Vulcan Low. Stratigraphic relationships between Belt sediments in the vicinity of these two Mesozoic faults and erosional variations in unconformities beneath strata from a number of different time periods, lead Price and Sears (2000) to suggest that this fundamental crustal boundary has been periodically reactivated in Mesoproterozoic,
Neoproterozoic, Early Paleozoic, and Mesozoic time. Like the shear zones discussed above, the Vulcan Low is another northeast trending, periodically reactivated suture zone. Why then does it not show a similar relationship to $T_e$?

If the GFTZ does represent a Proterozoic crustal boundary, are there fundamental compositional or structural differences between it and the Archean Vulcan Low? The GFTZ does show flexural rigidities more nearly similar to the Proterozoic Cheyenne Belt, Gneiss Canyon Shear Zone, and Crystal Shear Zone. If crustal strength within the Vulcan Low was decreased by reactivation of this boundary in the Mesozoic, some signature of this event should still be present given the long relaxation times for flexural rigidity (Burov and Diament, 1995).

**Correlation of reactivated crustal boundaries with mantle velocity structure**

As discussed above, the seismic velocity data of the Deep Probe transect (figure 20) shows no sign of the GFTZ. Given the relationship between crustal strength and crustal temperature and composition, the lower than average flexural rigidities of the reactivated shear zones ought to have some signature in the seismic record. It appears that when a more regional seismic record is examined, some of them do.

Figure 18 shows that, in general, the Archean Hearne and Wyoming provinces (figure 4C) are underlain by old, cold, high velocity mantle. These provinces also correspond to areas of high $T_e$ (figure 17). Proterozoic crust to the west of the Archean terranes (figures 16 and 17) and provinces accreted to the continent south of the Wyoming province during the Proterozoic show a wide range of $T_e$ values (figure 16) and mantle velocity signatures. Within the Proterozoic provinces, hot, young, low velocity mantle material seems to be migrating into this area from the south and southwest (figure 18). Careful comparison of figures 16, 17, and 18 reveal that the northeast trending fingers of hot mantle material correspond closely with areas of comparatively low $T_e$. In eastern Idaho, hot mantle and low values of $T_e$ outline the northeast trending Yellowstone hotspot track. In Utah, two northeast trending fingers of hot mantle material underlie the areas of comparatively low $T_e$ (~16 km) in the central and eastern part of the state (figure 16) which correlate with terrane boundaries in the Proterozoic crust (the Gneiss Canyon and Crystal Shear Zones)(Karlstrom, 1999). In the northern Rocky Mountains, cold mantle material underlies both the Vulcan Low and the GFTZ, although interestingly, a region of moderately warm mantle does extend eastward beneath the smaller northeastward extension of low $T_e$ above the GFTZ. A diffuse region of moderately warm mantle also extends northward towards the GFTZ from the present position of the Yellowstone hotspot.

Apparently, lineaments which lie along the northeast trending Proterozoic shear
boundaries south of the Wyoming province are acting as conduits for the penetration of hot, young mantle material from the west (Karlstrom, 1999). Upper mantle material there is being restructured by dynamic processes resulting from buoyancy anomalies in the asthenosphere below. This anomalously warm mantle is thought to result from the presence of a mantle plume (Lowry et al., 2000) and/or the present location and influence of the East Pacific Rise (Humphreys and Deuker, 1994). To what extent is the present correlation of $T_e$ with these inherited crustal boundaries due to their current reactivation by convective flow patterns within the upper mantle?

The regions of low flexural rigidity which correspond to the Gneiss Canyon and Crystal Shear Zones in central and eastern Utah continue into Colorado to the northeast (figure 16) while the regions of low mantle velocity do not. Without continuing Lowry and Smith's (1995) flexural rigidity map further to the east, it is hard to make definitive comparisons, but in both Utah and Montana, it seems that low flexural rigidity values within the northeast trending crustal boundaries extend further to the northeast than do the low mantle velocities. However, in the area of the Yellowstone hotspot track, the low mantle velocities extend further to the northeast than do the low flexural rigidities. This may offer some evidence that the crustal boundaries are providing conduits for the hot mantle material because they were weaker than the surrounding crustal provinces to begin with. In contrast, the Yellowstone hotspot is currently cutting a path into the Archean Wyoming province and hot mantle velocities there precede the weakening of the upper crust as represented by large reductions in $T_e$ (figures 17 and 18).

**Comparison of flexural rigidity with structural deformation**

Numerous studies have demonstrated that stretched continental lithosphere can remain weak for very long periods of time (e.g. Burov and Diament, 1995; Stewart and Watts, 1997) and hence, highly susceptible to continued deformation from subsequent tectonic events. Within the area of this study, continental crust was stretched and weakened by the formation of a three-armed intracratonic rift around 1.47 Ga (Sears et al., 1998) which led to the formation of the Belt Basin (figure 19). Along the west side of the area, extensive rifting in the late-Proterozoic led to the onset of seafloor spreading and the development of an Early Paleozoic passive margin (currently defined by the $^{87}\text{Sr} / ^{86}\text{Sr} = .706$ isopleth, figure 4C) along the west coast of North America (Sears and Price, 2000). Areas along this rift zone became the locus of contraction in the Mesozoic and extension in the Eocene. Currently, the regions along these ancient, inherited rift zones are characterized by the lowest flexural rigidities in the area.
Segmentation of Tc along the fold and thrust belt and its relationship to structural deformation

Most of the contraction and magmatism which affected the study area during the active margin tectonics of the Mesozoic was concentrated within the Cordilleran miogeocline which had formed on top of crust previously thinned by the late Proterozoic (Oldow et al., 1989) rifting. During late Cretaceous and Paleocene time, Paleozoic passive margin sediments, the Late Proterozoic rocks of the Windermere Supergroup which underlay them, and the 10-20 km thick (Hoffman, 1989) intracratonic Belt-Purcell Basin sediments (Winston, 1986) were tectonically inverted and thrust east-northeastward over the North American craton (Price and Sears, 2000). Figure 19 depicts the original location and thickness of Lower Belt-Purcell rocks and the amount of their east-northeastward thrust displacement during the Cretaceous and Paleogene, as determined by the palinspastic reconstructions of Price and Sears (2000).

Stewart and Watts (1997) and Watts et al., (1995) have documented along-strike segmentation of Tc in major fold and thrust belts around the world. They concluded that these variations are the result of inherited thermal and mechanical properties of the foreland lithosphere (often associated with pre-existing crustal boundaries) which continue to influence structural deformation over an entire Wilson cycle (i.e. crustal rifting, passive margin development, and eventual continental collision). In their flexural rigidity study of the Appalachians, they not only found that Tc was strongly segmented along strike, they also found that regions of low Tc correlated to pre-existing rift zones within the basement. They suggested that these zones of crustal weakness influenced the style of rifting during the Paleozoic onset of seafloor spreading (creating a strongly segmented, transform-dominated margin) and through time, localized the deposition of sediments along the continental passive margin.

In the northern Rocky Mountains region, the location of pre-existing structural features clearly had similar effects. Crust weakened by the formation of the Middle Proterozoic intracratonic rift localized the accumulation of large amounts of sediment. These sediments probably controlled the local structure of the passive margin which formed during the Late Proterozoic and may have influenced deformation resulting from the Paleozoic Antler orogeny. When pericratonic and exotic terranes collided with the western margin of North America in the Mesozoic, the amount and location of sediments in the Proterozoic Belt Basin were a determining factor in the structure of the fold and thrust belt through the central and northern parts of the area (i.e. due to basin inversion, Price and Sears, 2000). Currently, strong variations in Tc exist along the strike of the fold and thrust belt, with lower values of Tc occurring in the area close to the original zone of weakness and higher values of Tc occurring farther away from it (within the
southern Canadian Rockies (figure 21) and western Wyoming (figure 16)). Interestingly, if the reconstructions of Price and Sears are correct, the position of the northwest arm of the rift (where sediment accumulations were especially thick), as outlined by the north shelf of the Belt Basin, roughly follows the northern extent of the current, exceptionally low (5-10 km) values of $T_e$ within the study area (figure 19).

Based on studies of flexural rigidities in the Central Andes, Watts et al. (1995) also hypothesized that high values of $T_e$ (> 25 km) correlate with thin-skinned folding and thrusting above a basal decollement and that low values of $T_e$ (< 25 km) correlate with thick-skinned deformation involving basement shortening. That hypothesis is not supported by the results of this study. In the northern Rocky Mountains, thin-skinned folding and thrusting above a shallow decollement (thrusts outlined in white, figure 21) occurs equally in areas of high and low $T_e$. Thick-skinned, basement involved structures of the Laramide uplifts (figure 21) occur within areas of $T_e$ >25 km (i.e. the Beartooth Mountains and the Bighorn Mountains) as well as areas of low $T_e$. Although the Rocky Mountains are significantly older than the Central Andes and hence, $T_e$ values may have changed in the intervening years, it is likely that they would have changed in ways that strengthen this conclusion. Where basement involved structures occur within areas of low $T_e$, the crust may have recently been weakened by the presence of the Yellowstone hotspot and thus, $T_e$ may have actually been higher during the time of Laramide deformation. Also, given the general strong correlation between Archean provinces and high crustal strength, and the long relaxation times of $T_e$, it seems highly unlikely that the crust in these locations was significantly weaker at the time the uplifts were formed.

**Extensional regimes**

Numerous studies have demonstrated that stretched continental lithosphere can remain weak for very long periods of time (e.g. Burov and Diament, 1995; Stewart and Watts, 1997) and hence, highly susceptible to continued deformation from subsequent tectonic events. Most of the extremely low $T_e$ values within the study area define two crustal trends which mark the locations of previous rifts.

In the west, the lowest $T_e$ values are part of a zone of crustal weakness which parallels the rifted, late-Proterozoic continental margin (as defined by the $^{87}$Sr/$^{86}$Sr = .706 isopleth, figure 4C). Much of the major extension which has occurred in the Phanerozoic has taken place within, or along, this area of low $T_e$. The Cordilleran miogeoclone, which formed in this area on top of crust extensively thinned by late-Proterozoic rifting, became the locus of contraction and magmatism during the active margin tectonics of the Mesozoic and earliest Cenozoic. By the end
of the Paleogene, compression ceased and the gravitationally unstable, overthickened crust began to collapse (Karlstrom and Humphreys, 1998). This led to the formation of metamorphic core complexes in the Eocene (figure 22). Most of these large, extensional features lie to the north and east of the $^{87}\text{Sr}/^{86}\text{Sr} = .706$ isopleth in a region of mostly low $T_e$ roughly paralleling the basin axis of the northwest trending arm of the Belt Basin (figure 19). In this region, the North American crust is thought to extend further to the west beneath obducted slivers of the accreted terranes. In addition to the metamorphic core complexes, two Eocene rift zones have been
Figure 22. Relationship of flexural rigidity to distribution of extensional features. Eastern limit of major Cenozoic normal faults is depicted by the dotted white line (Muehlberger, 1992; Hamilton, 1988). This line closely parallels, but lies slightly to the east of the inferred boundary between Proterozoic and Archean crust (compare with figure 4C). Dark purple lines outline the proposed Eocene-Oligocene rift zone of Janecke (1994). Metamorphic core complexes, large Eocene extensional features which are thought to have formed from the gravitational collapse of compressionally overthickened crust are outlined in black (Parrish et al., 1988; Coney, 1980). FC=Frenchman Cap area, MC=Monashee Complex, TO=Thor-Odin area, OC=Okanagan Complex, P=Pinnacles Complex, KC=Kettle-Grand Forks Complex, PRC=Priest River Complex, VC=Valhalla Complex, BR=Bitterroot Complex, PI=Pioneer Complex.

proposed within the western half of the study area. Farthest to the west, seismic refraction profiles across the Columbia Basin have indicated the presence of rift-like grabens containing large amounts of sediment and paleomagnetic studies have determined large Eocene rotations for the area (Catchings and Mooney, 1988). Janecke (1994) proposed that the locus of extension moved slightly eastward by the late Paleogene. She proposes the existence of a rift zone which extends from north to south across the area, just to the east of the core complexes. Except for a small segment of exceptionally rigid crust in the northwest corner of the region, the eastern
border of major Cenozoic extensional features lies almost precisely along the transition to high $T_e$. This relationship was noted by Lowry and Smith (1994) for areas in the middle Rocky Mountains as well. In most places, the transition to high $T_e$ corresponds closely to the inferred western boundaries of the Archean crustal provinces (figure 4C). However, in the south-central part of the area examined in this study, the Archean lithosphere has been greatly altered by the presence of the Yellowstone hotspot. There, the eastern extent of major normal faults follows the transition to high $T_e$ rather than the probable border of the Archean crust (figure 22).

**Deformation in the foreland basin**

Several studies have addressed the effect of differences in $T_e$ on structure along a direction perpendicular to the strike of orogenic belts (e.g. Waschbusch and Royden, 1992; Watts, 1992; Stewart and Watts, 1997). Recent two-dimensional, relatively high resolution studies of $T_e$ along orogenic belts (e.g. this study; Lowry and Smith, 1995; Stewart and Watts, 1997) have shown that $T_e$ also varies widely along strike. This will influence not only deformation along the fold and thrust belt, but in the adjoining foreland basin as well. Of special concern is the use of sediment backstripping techniques which model the tectonic driving mechanisms of basin subsidence and require specific assumptions to be made about the flexural rigidity of the plate.

Figure 23 compares the flexurally backstripped subsidence profiles of Pang and Nummedal (1995) with the flexural rigidities determined for these areas by this study. The western edge of profiles A and C lie within the strong, Archean crust. The western edge of B lies within the weaker crust along the apparent perimeter of the GFTZ. As expected, the profile in weaker crust shows a narrower foreland basin. However, like many other studies which use flexural subsidence profiles to compare tectonic differences over broad regions, these profiles all use the same flexural rigidity ($1 \times 10^{24} \text{Nm}$, or $T_e \sim 48 \text{ km}$). The flexural rigidity used is appropriate for profiles A and C, but clearly not for B, where the flexural rigidity, as estimated by this study, is less than $1 \times 10^{23} \text{Nm}$ (figure 10). Presumably, the difference in foreland basin width would be even more pronounced if a smaller, more appropriate flexural rigidity had been used.
Figure 23. Flexurally backstripped subsidence profiles across the northern Rocky Mountains foreland basin (from Pang and Nummedal, 1995). Light shading marks magnitude of subsidence from 97 to 90 Ma and dark shading marks magnitude of subsidence from 90 to 83 Ma. Vertical scale is tectonic subsidence in meters and highly exaggerated compared to horizontal scale. The narrowest foreland basin (west side of profile B) lies closest to the GFTZ. The western portions of profiles A and C lie within the rigid crust of the Archean provinces.
Conclusions

The MESE (Maximum Entropy Spectral Estimation) coherence method of Lowry and Smith (1994) was used to determine crustal strengths (as parameterized by flexural rigidity, D, and effective elastic thickness, T_e) for the Rocky Mountains of Montana and portions of adjacent states. Flexural rigidities proved useful for identifying inherited crustal structures and gaining a deeper understanding of the regions structural evolution. Comparisons with other, two dimensional data sets (such as magnetic field data and upper mantle velocity structure) illustrate the potential of high resolution flexural rigidity mapping. Furthermore, high resolution flexural rigidity estimates can provide important constraints on quantitative models of structural evolution along the fold and thrust belt and throughout the foreland basin. Structural studies along mountain belts often assume a constant crustal strength along strike (e.g. Pang and Nummedal, 1995), clearly an erroneous assumption for the Rocky Mountain region (this study and Lowry and Smith, 1995) and many other orogenic belts around the world (Stewart and Watts, 1997).

Within this study area, T_e showed a strong correlation with: 1) crustal age (specifically between Archean crust and high values of T_e, 2) inherited crustal boundaries and former rift zones (both the GFTZ and the rift zones within the area are currently composed of relatively weak crust), and 3) upper mantle velocities (low upper mantle velocities underlie areas of low T_e). Suggestive correlations were found between T_e and structural deformation along the fold and thrust belt and within the foreland basin. Exceptions to these correlations underscore the importance of using flexural rigidity results in concert with other geophysical or geological data.

Further work

The lack of a flexural rigidity signature within the Vulcan Low is one of the most perplexing findings of this study. Evidence exists that at least the western end of the Vulcan Low was active as recently as Mesozoic time, yet compared to other reactivated shear zones between Archean and/or Proterozoic crust, the crust of the Vulcan Low is remarkably strong. Is this simply a function of the timing and extent of its reactivation, or is it due to some fundamental difference in composition or structure? More generally, do the crustal strengths within these inherited, often reactivated shear zones fluctuate with time (annealing, strengthening, and occasionally weakening when reactivated) and if so, do these fluctuations show up in the stratigraphic record?

Another interesting question concerns the high flexural rigidities of the Columbia Basin. Is the same process (perhaps melt depletion in the upper mantle?) responsible for both the high
flexural rigidities of the basin and the high upper mantle velocities beneath it? Given the low flexural rigidities and low mantle velocities of the surrounding areas, how stable is this situation over time? Perhaps more importantly, what controls the relationship between $T_e$ and the thermal age of the crust in this area?

Additional questions and suggestions for future work are briefly outlined in Appendix B. Clearly, flexural rigidities of the crust can reflect the inheritance of crustal properties and ancient terrane boundaries through the passage of time, and countless episodes of complex deformation. Hopefully, this study will provide the groundwork for future investigations.
APPENDIX A

Input parameters of the MESE coherence model

The coherence model estimates the flexural rigidity of the crust by assuming that the earth’s upper layers can be modeled as a strong elastic plate which can support applied loads over geologic time scales. While the strength of this plate is dependent on its physical composition, thermal profile, and state of stress, the coherence model assumes that the combined effects of these parameters will control the degree of correlation between topography and Bouguer gravity anomaly as a function of wavelength in a manner dependent on the flexural rigidity of the plate.

The coherence functions in this study were calculated using computer code developed by Anthony Lowry for computations of the flexural rigidity based on the MESE coherence method. The flow chart in figure A1 provides a brief outline of the calculations, including required inputs. In addition to the physical data required, the programs must be “tuned” for the area of study by appropriate selection of a number of hard-wired parameters (table A1). Examples of the sensitivity of $T_g$ to some of these important “hidden” variables and model inputs is described in the following sections.

Where to start

Much of the coherence program input consists of real and modeled data, but some of the hard-wired parameters control limitations of the computational methods and algorithms used (e.g. adet, wvno, and window size) and must be iteratively “tuned” by trial and error. Once an initial determination of the program parameters has been made, tests should be run for different window sizes in a number of different areas. The behavior of the coherence functions and residual error curve then guide iterations in the program parameters. An appropriate choice of parameters will lead to a smooth log-hyperbolic error curve with reasonable curvature (not too flat) and no local peaks, plus a family of coherence functions whose transition wavebands are for the most part separate and move to higher wavelengths with increasing $T_e$ (figure 8). The programs begin the search for $T_e$ by automatically computing coherence functions for $T_e = 2, 4, 8, 16, 32, 64,$ and $128$ km. Plotting these seven coherence functions on the same graph can give a quick “first cut” at determining whether or not the input parameters have been well chosen. While individual estimates of $T_e$ can certainly exhibit sizeable error, a careful selection of program parameters should not contribute significantly to errors in the general nature or trend of $T_e$ over extended areas. Systematic bias in regional results will most likely be due to a failure of one of the model’s
Calculate the observed coherence.

Extract large data window for load deconvolution.

Extract sub window for which coherence is to be computed.

Estimate the auto and cross spectral densities of gravity and topography data using MESE.

Calculate the complex Fourier transform of gridded topography and gravity data over the large data window.

Iterate for Te values of 2, 4, 8, 16, 32, 64, and 128.

Deconvolve surface and subsurface loads.

Inverse Fourier transform deconvolved loads back to spatial domain.

Calculate the residual error of predicted minus observed coherence.

Calculate a best-fit hyperbola relating the residual error of the coherence functions to the log of Te.

Assumed Te equals the minimum of the hyperbolic error function.

Determine error bounds on Te estimate. Map out range of Te for which the residual error equals the residual error of the best-fit Te plus 0.3.

Entire range mapped out, OR, all Te values between 2 and 128 have been mapped.

STOP

Figure A1. Program flowchart for MESE coherence code. Input data and program parameters described in Table A1 are shown (without boxes) where utilized in the analysis.
### Table A1. MESE coherence code program variables

<table>
<thead>
<tr>
<th>Program variable:</th>
<th>Description:</th>
<th>Value: (this study)</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Physical data: (observed or modeled)</strong></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Topography</td>
<td>Gridded, in km, parameterized identically to Bouguer anomaly data.</td>
<td>see figure 1</td>
</tr>
<tr>
<td>Bouguer anomaly</td>
<td>Gridded, in km, parameterized identically to topography data.</td>
<td>see figure 2</td>
</tr>
<tr>
<td>Seismic velocity-depth relationship</td>
<td>Determined from seismic refraction studies velocity contrast depths in meters, velocities in m/sec.</td>
<td>see figure 12</td>
</tr>
<tr>
<td>Subsurface load depth</td>
<td>Depth at which subsurface compensation is assumed to occur.</td>
<td>see figure 12</td>
</tr>
<tr>
<td><strong>Adjustable program parameters:</strong></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Data window size</td>
<td>Size of area for which the observed and predicted coherence is computed.</td>
<td>200 x 200 km, 400 x 400 km or 560 x 560 1116 x 1116 km</td>
</tr>
<tr>
<td>Deconvolution window size</td>
<td>Size of area which is used for deconvolution of surface and subsurface loads. Chosen to be much larger than the data window size to minimize any windowing effects of the Fourier transform.</td>
<td>df = -1.0 i.e. equal loads of opposite sign</td>
</tr>
<tr>
<td>Surface to subsurface load ratio</td>
<td>Hard-wired load ratio for wavenumbers less than the wavenumber cutoff used for deconvolution. Below the cutoff, all loads are essentially locally compensated and surface and subsurface loads cannot be distinguished.</td>
<td>adet = 45 (scaled limit of determinant in deconvolution equations)</td>
</tr>
<tr>
<td>Wavenumber cutoff for deconvolution</td>
<td>Computational limit below which the load ratio loses wavenumber dependence. Below this cutoff, the load ratio cannot be computationally determined.</td>
<td>wvno = .1 - .15</td>
</tr>
<tr>
<td>Wavenumber cutoff for error function determination</td>
<td>Wavenumbers above the wavenumber cutoff are not used in the determination of the residual error.</td>
<td></td>
</tr>
</tbody>
</table>

basic assumptions, such as its dependence on the lack of correlation between surface and subsurface loads (Macario, 1995).

**Dependence on window size used for MESE coherence calculations**

Window sizes for the coherence calculations must be chosen as small as possible to
maximize the resolution of \( T_e \), yet large enough that the longer wavelengths in the transition waveband of the coherence functions are adequately resolved. This implies that larger window sizes will be required in areas of higher \( T_e \). The best-fit \( T_e \) computed within a window is an average over the entire window. \( T_e \) estimates from a 200 x 200 km window and a 400 x 400 km window will be similar if both windows resolve the coherence function transition region and the range of flexural rigidities encompassed by the larger window is similar to that of the smaller window. Figure A2 illustrates the difference in best-fit coherence functions and their residual error curves for two different window sizes centered near the middle of the study area. Based on the flexural rigidities depicted in figure 5, the 400 x 400 km window encompasses a wider range of flexural rigidities than the smaller window, with a slight bias towards higher values of \( T_e \) in the Great Plains region. As figure A2 shows, the larger window size yields a slightly higher value of \( T_e \). The larger residual errors for the 400 x 400 km window are most likely a result of the hard-wired wavenumber cutoff which limits the range of wavenumbers used to calculate the residual error (see discussion below).

If the coherence window dimension is smaller than the wavelengths needed to resolve the coherence transition, the estimated \( T_e \) may not be very reliable. An extreme example of this is

**Figure A2.** Effect of window size on determination of \( T_e \). Window center is near the middle of the study area (up ten rows from the south, in 13 columns from the west) at the boundary of the NRM and USGP provinces. Both windows yield well behaved coherence functions and residual error curves.
shown in Figure A3, which shows the best-fit coherence curves, and associated error functions, for two different window sizes centered on a point in the southeast corner of the study area. The smaller window results in a $T_e$ estimate over twice as large as the larger window! At first glance, the fit of the observed and predicted coherence curves look fairly similar for the two window sizes. This is confirmed by the residual error curves which show nearly identical best-fit residual errors. However, a closer examination of the best fit coherence from the smaller window reveals that the predicted coherence begins to decrease at noticeably higher wavelengths than the observed. Also, the residual error function shows a pronounced local peak at low values of $T_e$. This can indicate that the window size is too small to adequately resolve the longer flexural wavelengths in the transition waveband and result in physically unrealistic behavior of the coherence functions. This can be clearly seen in figure A4, where a wide range of $T_e$ values are plotted on the same graph. The 400 x 400 km window shows a pronounced inversion of the smaller $T_e$ values in the transition waveband. With the longer flexural wavelengths inadequately resolved, there is not enough information to accurately determine the dependence of coherence on wavelength as a function of $T_e$ in the upper transition region. The coherence functions from the larger (560 x 560 km) window correctly show the transition from coherence to incoherence occurring at longer wavelengths for larger values of $T_e$. The 560 x 560 km window size yields the more reliable estimate of $T_e$.

A larger window size may also be indicated when the residual error curve becomes flat for a very wide range of $T_e$. Once the program finds the best-fit $T_e$, it attempts to map out the range of $T_e$ for which the residual error equals the minimum (best-fit) residual error plus 0.03 (an empirically determined estimate of the minimum expected standard deviation of the MESE coherence estimates). If there is not enough information at the longer transition wavelengths, the residual error will be based primarily on the hard-wired program variables and will remain largely the same as $T_e$ increases. The residual error curve will remain low and flat, and the program will have difficulty finding an upper error bound. The relationship between the slope of the error curve above the best-fit $T_e$ and the adequacy of the window size should be empirically determined.

The deconvolution window size can also have a large effect on the behavior of the coherence functions. If the deconvolution window is too small, the low $T_e$ coherence functions may fall below the higher $T_e$ functions with increasing wavelength towards the top of the transition waveband (much like the inversion seen in figure A4). Tests of deconvolution window size indicated a 1116 x 1116 km window was the minimum size necessary to achieve reasonable
Figure A3. Best-fit coherence functions and residual error curves for 400 x 400 (top row) and 560 x 560 km windows (bottom row) centered on the marked grid center just inside the southeast corner of the study area. (i.e. the second row from the bottom, second column in from the right on figure 5). Note the ill fit of the predicted coherence function at the top of the transition waveband in the 400 x 400 km window and the peak in the residual error function at low $T_e$.

Figure A4. Coherence function behavior for the windows depicted in figure A3.
results throughout the study area. This is quite large compared to the coherence window sizes, but small enough that a number of different, overlapping windows must be used to cover the area. Although each deconvolution window spans multiple physiographic provinces, the deconvolution is performed using the load depth and density model of the province in which the center of the smaller coherence window within lies. While it would seem that the appropriateness of the crustal model to the deconvolution window as a whole, or the variation in physical properties across the window might introduce some bias into the results, tests showed little to no difference in best-fit $T_c$ estimates obtained from different deconvolution windows over the same coherence window.

**Dependence on the wavenumber cutoff in the error function determination**

The wavenumber cutoff is a hard-wired limit which controls the calculation of the residual error between the predicted and observed coherence functions for each value of $T_c$. Maximum entropy spectra represent noise processes poorly (Lowry, 1994) and the MESE coherence functions tend to have a strong positive bias at short wavelengths where the true coherence goes to zero (figure 10). To eliminate this bias, the residual error calculation is set to exclude wavenumbers above a certain predetermined limit. Since this is a hard-wired limit, a greater percentage of the residual error will be calculated from values below coherence = 1.0 (where the predicted and observed coherence functions are most likely to differ) as the top of the transition waveband moves to lower wavenumbers for higher values of $T_c$. This can bias the residual error towards higher values in regions with larger $T_c$. It would perhaps be more meaningful to determine the residual error for the transition waveband only, but the many local minima in the MESE coherence curves would make it difficult to perform an automated search for the bottom of the transition waveband in each individual coherence function.

If there is significant inversion in the $T_c$ values in the transition waveband (as in figure A4, 400 x 400 km window), this can have a large effect on the determination of $T_c$ since the residual errors determine the log hyperbolic error function (calculated after each $T_c$ iteration) which controls the search for $T_c$. This can lead to a pronounced peak in the residual error curve at low values of $T_c$ (similar to that seen in figure A3). This occurs because the low $T_c$ coherence curves invert low in the transition waveband and then lie close to the predicted coherence for the higher $T_c$. If the true $T_c$ is large, then the inverted low $T_c$ curves will provide as good (or sometimes better!) match to the observed coherence function as will one of the high $T_c$ curves since the wavenumber cutoff may effectively eliminate comparison of the predicted and observed
coherence function curves at wavenumbers above the inversion (where they will differ markedly, as they should). Furthermore, the residual errors will probably not be able to produce a meaningful error function curve at each iteration and the search may not progress towards the true best-fit $T_e$. Fortunately, minor inversion affects only the lowest $T_e$ coherence functions and can leave the higher $T_e$ coherence functions unchanged and still capable of providing an appropriate, best-fit $T_e$ if the true $T_e$ falls within the range of the unaffected curves. Inversion of the coherence functions can result from an increase in the load depth, or a change in the hard-wired load ratio in addition to an inappropriate deconvolution, or coherence window size.

For well behaved coherence functions (i.e. no inversion), reasonable differences in the wavenumber cutoff have a negligible effect on the estimated $T_e$ (figure A5). Here, the small difference in $T_e$ estimates is due to the use of a different number of points in the residual error calculation. Note the slightly higher residual error from the calculation with the larger wavenumber cutoff (because of its calculation over a larger wavelength range which includes more of the transition waveband). The slightly different residual errors lead to minor variations in the computed error functions, and hence, slight differences in the iterative, directed search for the best-fit $T_e$. Since the MESE coherence functions for different values of $T_e$ show the greatest difference near the top of the transition waveband (figure 8), the wavenumber cutoff can essentially take on any value high enough to include the top of the transition waveband, yet low enough to exclude wavenumbers above the bottom of the transition waveband where the MESE coherence functions begin to be less reliable. The cutoff value chosen should fit these criteria for all possible values of $T_e$ (figure A5). For this study, tests indicated a wavenumber cutoff in the range of 0.1 to 0.15 was most appropriate.

**Dependence on load depth and density model**

Forsyth (1985) showed that the predicted coherence is relatively insensitive to errors in the depth assumed for subsurface loads. Computational difficulties pose greater constraints on depth since the load deconvolution requires downward continuation of the gravity signal and that process is most stable for shallow depths. Lowry and Smith (1994, 1995) used the shallowest first-order density discontinuity for all calculations in their study area for this reason. A number of base-of-crust load depth tests run for windows in this study area led to unrealistic looking coherence functions which were difficult to correct by changes in the other program parameters. This result, along with the middle to shallow crust accommodation depths determined by Lowry and Smith (1995) for areas in the southern third of this study area, led to the adoption of the load depths used in this study (figure 12).
Figure A5. Choice of wavenumber cutoff in residual error calculations. A and B show dependence of the best-fit coherence functions on the wavenumber cutoff used in the determination of the residual error. Values to the left of the dashed lines were used to calculate the residual error and guide the search for the best-fit $T_e$. The cutoff has a negligible effect on the estimate of $T_e$ as long as it includes the top of the transition waveband and excludes coherence values at the larger wavenumbers beyond the bottom of the transition waveband. For choices within this range, the cutoff has a negligible effect on the estimate of $T_e$. C illustrates the position of the .05 and .10 cutoffs in the transition waveband for different values of $T_e$. If the true $T_e$ is low, a larger wavenumber cutoff is required to include the top of the transition waveband. Values of 0.1 to 0.15 were used in this study to ensure adequate inclusion of the coherence transition for all values of $T_e$. Calculations shown here are for the 200 x 200 km window of figure A2 with all program variables other than the wavenumber cutoff held constant.
Figure A6 B and C show the effect on the estimated $T_e$ of increasing the load depth from shallow to mid-crustal levels. The predicted coherence functions in the two cases are very similar and yield $T_e$ estimates which differ by about 12 percent. Changing the seismic velocity/density model has a similarly small effect. Figure A6 A shows the best-fit coherence function for the same 400 x 400 km window calculated with the completely different velocity/density model used for the area to the north of this window. The relative insensitivity of the coherence functions to density model and loading depth can be seen in the derived flexural rigidity map (figure 9). While the modeled density structure and load depths change abruptly at the seven region boundaries, the map shows absolutely no correlation between those region boundaries and the estimated $T_e$.

Figure A6. Sensitivity of $T_e$ to density model and choice of subsurface load depth. Best-fit coherence functions are calculated for the 400 x 400 km window in the southwest corner of the USGP province. Seismic velocity models which constrain the density are from figure 8. A uses the USGP model, B and C use the MRM model.
Dependence of the hard-wired load ratio

The response of an elastic plate to surface and subsurface loads depends on both the rigidity of the plate and the wavelength of the load. If the flexural rigidity of the plate is zero, all loads will cause a deflection of the plate such that topography and gravity will be perfectly correlated. For all plates with non-zero rigidity, there will exist a load wavelength below which the elastic stresses within the plate can partially, or totally, support the load. Within this wavelength range the initial applied loads \((H_i, W_i)\) can be derived from the coherence equations and the surface and subsurface components of relief can be deconvolved. Any load whose wavelength lies above this range exceeds the capacity of the plate to support it and once again, must be compensated by local deflection of the plate. For these wavelengths, the gravity and topography signals will be completely correlated (coherence = 1.0) and surface and subsurface loads can no longer be computationally distinguished (the determinant of the matrix of coefficients of the equations used to solve for the initial loads goes to zero). However, the initial applied loads can be determined from the load ratio defined by the coherence equations. The ratio of bottom to top loading \((f)\) is:

\[
f(k) = \frac{\langle \Delta \rho W_i \rangle}{\langle \rho_0 H_i \rangle}
\]

(Forsyth, 1985). While the predicted coherence makes no assumptions about the load ratio at wavelengths where the surface and subsurface loads can be deconvolved (which ideally, should be through the entire transition waveband), a priori values of the load ratio can be used to continue the calculations into wavelengths where gravity and topography become completely coherent.

Computationally, an arbitrary lower limit (the program parameter adet) must be imposed on the value of the matrix of coefficients determinant in the initial load equations. At all wavelengths where the determinant falls below this limit, a surface to subsurface loading ratio (program parameter df) is assumed. Tests showed that the values chosen for adet, and especially, df, can have a substantial effect on the estimated \(T_e\). The use of a much larger window size for load deconvolution than coherence estimation makes the coherence functions particularly sensitive to the long wavelength load ratio (Lowry, personal commun.). If df is not carefully chosen, severe inversion of the coherence functions can result and the search algorithm, which depends on the calculated residual errors, will lead to an erroneous value of \(T_e\).

A hard wired load ratio of \(-1.0\) was chosen for this study based on several factors. First, an examination of coherence functions calculated with different load ratios for a number of...
windows within the study area indicated that this value produced the most physically realistic behavior of coherence as a function of wavelength and $T_e$. Second, Bechtel (1989) explicitly calculated the subsurface to surface load ratio as a function of wavelength for best-fit estimates of $T_e$ obtained by the traditional (periodogram-based) coherence method for large regions which overlap the area of this study. Although necessarily dependent on the value of other parameters in his model, his results indicate values close to $-1.0$ for the Rocky Mountain fold and thrust belt, the Canadian Rockies, and the Columbia Plateau. Third, Lowry and Smith (1995) used a value of $-1.0$ (Lowry, personal commun.) in their MESE coherence analysis of a large region extending southward from the southern third of the area investigated here. Since the load ratio determined was identical to that used by Lowry and Smith (1995), their value of $a_{det}$ was retained as well. This should facilitate comparisons between the results of Lowry and Smith (1995) and those of this study.
APPENDIX B

Future Work

While only the large, integrated, regional studies will be able to address many of the issues relating to hypotheses concerning the nature and mechanisms of crustal inheritance, a number of smaller scale, local studies have suggested themselves during the course of this work.

1) Expansion of the flexural rigidity map further to the north.
The Tc data hints at another area of low Tc extending into the Canadian Plains to the north. The location of these low values of Tc is north of the Vulcan Low, and seems to show no simple correlation to the inferred position of Proterozoic magmatic arcs or terrane boundaries, although this is impossible to tell without knowledge of the full extent of this feature. Also, it would be interesting to see if the magmatic arcs to the north (such as the ~1.8 Ga Rimbey magmatic arc) showed distinctive changes in Tc.

2. Forward gravity profiles across the GFTZ and Vulcan Low.
It would be illuminating to compare Tc values estimated from forward gravity profiles across the GFTZ and Vulcan Low with those estimated from the MESE coherence method. The two methods are subject to different assumptions and biases, and results from the two have never been compared within the same area. Similar results from the two methods would strengthen the interpretation of the GFTZ as an area of decreased crustal strength and perhaps more importantly, would lend credence to the idea of using the relatively high resolution MESE coherence technique to prospect for ancient crustal boundaries.

3. Additional flexurally backstripped subsidence profiles and an examination of foreland basin stratigraphy.
Several papers have addressed the relationship of Tc to deposition and deformation within the foreland basin (e.g. Stewart and Watts, 1997; Waschbusch and Royden, 1992). A number of interesting comparisons could be made between further profiles and the Tc results of this study. First, none of the profiles from Pang and Nummedal (1995) lie completely within the low Tc values along the GFTZ. Second, Waschbusch and Royden (1992) predict that migration of the foreland basin over areas with substantially different Tc will lead to episodic migration of the
foreland bulge which should be visible in the stratigraphic record. Specifically, the foreland bulge is predicted to migrate slowly over weak crust and more rapidly over more rigid crust. It might be interesting to look at rates of foreland bulge migration as a function of estimated $T_e$.

4. **Create a map of the ratio of subsurface/surface loading within selected wavebands for the best-fit $T_e$ values estimated in this study.**

The ratio of subsurface/surface loading can be directly calculated from $W_1$ and $H_1$, the initial subsurface and surface loads. Specifically, $f = \Delta \rho W_1 / \rho_s H_1$. These two quantities fall directly out of the load deconvolution equations and only a minor rewriting of the code would be necessary to solve specifically for the load ratio. Although these parameters are dependent on the coherence model assumptions and inputs, investigating changes in these parameters as a function of location and wavelength might prove illuminating.

5. **Investigate behavioral differences of the GFTZ and the Vulcan Low over time.**

The Vulcan Low has been interpreted as a north dipping Archean suture zone (Ross et al., 1991) while the GFTZ has been interpreted as an Archean (?) intracontinental shear zone (Boerner et al., 1998). Seismic refraction data shows the presence of a deep, high velocity crustal layer extending beneath the Wyoming province, GFTZ, and most of the Medicine Hat Block of the Hearne province to near the Vulcan Low. Various authors have inferred that each of these zones has shown periodic reactivation, yet the crust within the Vulcan Low is now apparently quite strong. How have the forces acting on these two zones differed over time and are their differences primarily a product of the mechanics and configuration of the original boundary, or a product of tectonic forces which have shaped the area since?

6. **Investigate the effect of density decreases due to large impacts on estimates of $T_e$.**

Relationships determined from crater modeling studies allow reasonable predictions of the amount and depth of deformation as a function of crater size. Density reductions due to fracturing and brecciation of the impacted rocks could be estimated and local structural relations used to roughly estimate the remaining volume of affected rocks after erosion and tectonic dissection. The resultant Bouguer anomaly could then be calculated and compared to the anomaly observed. Any contribution to the Bouguer anomalies from the impact could be eliminated and the resultant Bouguer anomalies could then be used to re-estimate $T_e$. 
References


Lester, A and Farmer, G.L (1998), Lower crustal and upper mantle xenoliths along the Cheyenne belt and vicinity, Rocky Mountain Geology, 33, 293-304.


Pratt, J.H. (1855), On the attraction of the Himalaya Mountains and of the elevated regions beyond them, upon the plumb line in India, Philosophical Transactions of the Royal Society, 145, p. 53-100.


