Lower Crustal Flow and Detachment in the North American Cordillera: a Consequence of Cordillera-Wide High Temperatures

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Key Points:

1. Demonstrate most of the North American Cordillera is remarkably uniformly hot in common with other backarcs, 800-850C at the Moho
2. Demonstrate uniformly thin crust and flat Moho in most of Cordillera; interpreted to be a consequence of lower crust flow and detachment associated with the high temperatures
3. Over 10’s m.y., Moho viewed as a boundary between almost ‘liquid’ lower crust over a low-viscosity upper mantle

For Journal of Geophysical Research, Solid Earth

File: LowerCrustFlow

January 6, 2016

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Abstract.

In this article I make the case for widespread lower crustal flow and detachment in the North American Cordillera. Accumulating seismic structure data show the surprising result that the crust is thin and the Moho flat in most of the Cordillera in spite of extensive normal faulting and shortening deformation. It has been previously concluded that the high elevations are due to thermal expansion from Cordillera-wide high temperatures, not a crustal root. I now argue that the constant crustal thickness and flat Moho are a consequence of lower crust flow associated with the high temperatures. Lower crust flow is inferred for Tibet and high Andes where the crust is thick. More surprising is the similar conclusion for the Basin and Range of western US where the crust is thin, but high temperatures are inferred to result from current extension. However, there is now adequate data to conclude the Basin and Range is not unique. Seismic structure studies show the crust in most of the Cordillera from Mexico to Alaska is uniformly thin, 33±3 km, with a remarkably flat Moho. Not just the Basin and Range, but most of the Cordillera is uniformly hot in common with other backarcs, 800-850°C at the Moho. The

uniform crustal thickness results from lower crust flow in a weak lower crust. The backarc Moho can be viewed as a boundary between almost 'liquid' lower crust over a low-viscosity upper mantle. The Moho boundary relaxes to a nearly-horizontal gravitational equipotential over a few 10s of m.y.

1. Introduction

In this article I summarize the evidence that there has been both lower crustal channel flow and detachment in much of the North American Cordillera, and the conditions for such flow and detachment are a consequence of wide-spread high backarc temperatures. The flow provides an explanation for why the crust of most of the Cordillera from Mexico to Alaska has remarkably uniform thickness as well as being thin, 33±3 km, with a very flat Moho. The thin crust and its uniformity are demonstrated by a wide range of seismic structure data, from seismic tomography (notably noise tomography using the high density stations of USArray), multichannel seismic reflection, wide angle seismic refraction, and receiver function studies. This surprising uniformity is in spite of the varied current and past tectonics, with regions of major crustal extension and shortening. Most normal and thrust faults that are well-defined in the upper crust do not offset the Moho which remains flat (examples by Cook et al., 1992; Klemperer et al., 1986). Also, the high grade metamorphic rocks that are interpreted to have been brought to the surface from the lower crust by tectonic processes, have no expression in Moho displacements. I argue that these observations may be explained by the increasingly convincing evidence that I summarize, that the crust of most of the North America Cordillera is very hot, 800-850^{o}C at the Moho. The high temperatures in the Cordillera compared to the cold adjacent stable areas (~450^{o}C at the Moho) explain the high elevations of the Cordillera through thermal expansion density reduction, even though its crust is thin (e.g., Hyndman and Currie, 2011). Although there are important inferred lateral temperature variations within the Cordillera, I show that the first-order pattern of high temperatures is an adequate approximation for important conclusions on lower crust flow. I argue that such high temperatures result in very low viscosity in the lower crust and allow lower crust flow that has flattened the Moho in most of the Cordillera.

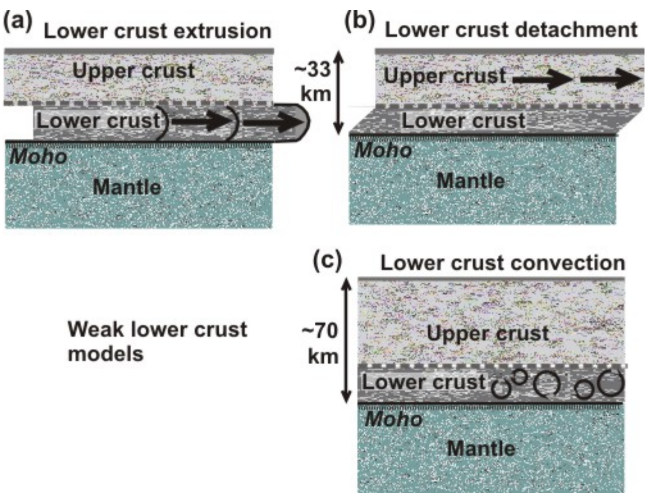

Figure 1. Hot weak lower crust: (a) Local lower crustal flow that may be elevation driven, that flattens Moho offsets, (b) Lower crust shear detachment, (c) Lower crust convection.

Whenever Moho topography develops by faulting or other tectonic deformation, it flattens over geologically short times.

I do not deal with the details of lower crust flow; for a discussion of the various forms, see Klemperer (2006). There are two related processes, lower crust detachment and lower crust convection (**Figure 1**). Lower crust detachment, where the strong upper crust moves horizontally relative to the mantle, has less severe constraints than flow on the low effective viscosity and thickness of the weak lower crust. If flow can occur, it is likely that detachment can occur. Oldow et al. (1990) showed how lower crustal detachment and upper crust orogenic float is required by tectonic continuity in foreland thrust systems. Mazzotti and Hyndman (2002) provide a current example where the upper crust is being driven 800 km westward to the eastern Cordillera mountain front by the collision of the Yakatat terrane in the Gulf of Alaska. A large scale decollement in a weak lower crust seems required. Fuis et al. (2008) concluded a similar orogenic float model for Alaska. The other extreme is lower crust convection, as concluded for the high Andes by Babeyko et al. (2002; 2006). Convection has greater constraints on effective viscosity and on the thickness of the weak layer. It is likely that convection requires very thick and hot crust such as in the high Andes and Tibet. Several pioneers in recognizing lower crustal detachment and flow in at least some areas were German geophysicist Rolf Meissner (e.g., Meissner and Mooney, 1998; Meissner and Kusznir, 1987; Meissner et al., 2006) and Canadian geophysicist Giorgio Ranalli (Ranalli, 2003; Ranalli, 2000; Ranalli and Murphy, 1987; Fernandez and Ranalli, 1997; Afonso and Ranalli, 2004). Their conclusions should be more widely appreciated. The enormous amount of new relevant data that have recently become available makes conclusions of lower crust flow and detachment much more secure. Of special importance are the thermal constraints that allow us to recognize that there is Cordillera-wide high heat flow and that lower crustal flow and detachment can occur over most of the North American Cordillera and other continental backarcs.

1.1 Thermal Definition of Backarc

In this article I define "backarc" thermally, as the generally well-defined region of crust and upper-mantle high temperatures that are now recognized landward of most continental arc volcanic chains (e.g., Hyndman et al., 2005; Currie and Hyndman, 2006). Flat slab areas are the infrequent exceptions. Hot backarcs generally correspond to tectonic "mobile belts" and to high-elevation "mountain belts". The high temperatures result in common high elevations, low effective elastic thickness Te, low seismic velocities in the upper mantle, and thin weak lithospheres that are readily deformed. In the geological

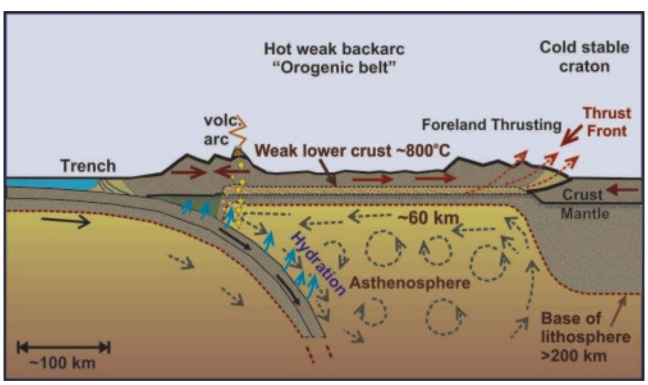

Figure 2. Cross-section of a subduction zone and hot backarc with thin lithosphere maintained by small-scale convection, in contrast to the thick lithosphere of the adjacent cold stable craton. The upper crust may move independently of the mantle.

record we interpret the main part of most orogenic belts to be ancient former hot backarcs. For the Cascadia part of the Cordillera the backarc hinterland defined this way extends from the Cascadia volcanic arc to the western side of the foreland belt (e.g., Hyndman, 2010) (**Figure 2**).

I include in the Cordillera hot backarc the former backarcs in northern British Columbia and in southwestern United States because they still have the characteristic backarc high temperatures. These former subduction zones were cut off by the Queen Charlotte and San Andreas transform faults in the Cenozoic. Their high temperatures have not significantly declined because the thermal decay time after subduction stops is 300-500 m.y. (e.g., Currie and Hyndman, 2006; Sleep, 2005). In Alaska the hot mobile backarc extends from the volcanic arc northward some 400 km to the Brooks Range which lies several 100 km from the arctic coast. North of the Brooks Range there is stable thick crust and cold thick lithosphere (e.g., Fuis et al., 2008; Veenstra et al., 2006; O'Driscoll, et al., 2015). I note that much of the eastern Cordillera foreland belt lies over the cold and thick cratonic lithosphere where backarc upper crust and advancing sedimentary thrust sheets have been thrust over the stable craton. Many of my examples come from the Canadian Cordillera where there has been no recent extension or shortening; in southern British Columbia the backarc extends to the Rocky Mountain Trench which is over the backarc-craton thermal and lithosphere thickness boundary (e.g., van der Velden and Cook, 1996; Hyndman and Lewis, 1999; Bao et al., 2014). In parts of the western U.S.A. there is considerable tectonic complexity in the Colorado Plateau and adjacent areas where the lithosphere has been thinned recently, in the currently extending Basin and Range area, and in the Yellowstone hot spot. However, all of these areas have the high temperatures characteristic of backarcs.

1.2 Previously Concluded Areas of Lower Crust Flow; Tibet, High Andes, and Basin and Range

Lower crust flow has previously been concluded for several areas of tectonically thickened orogenic crust. The most studied currently active case is the Tibet high plateau and its flanks (e.g., Clark and Royden, 2000; Beaumont et al., 2001; 2004; 2006; Godin et al., 2006; Klemperer, 2006; Harris, 2007). The thick crust of Tibet due to India-Asia collision results in hot lower crust flow into adjacent areas, from high elevation gravitational potential to lower elevations. There is little deformation in the strong upper crust or the upper crust moves independently. Tibet has a 70 km thick crust such that there should be high temperatures in the lower crust, even for a normal geothermal gradient. The temperature gradient, however, is especially high because much of Tibet is in the thermally recent hot backarc for the Tethys ocean subduction between India and Asia before collision. Very low viscosities are therefore expected in the deep crust. Similarly, the high Andes are concluded to exhibit lower crust flow associated with high temperatures because of both the thick crust (e.g., McGlashan et al., 2008) and because the area is the current subduction zone hot backarc (e.g., Springer and Forster, 1998; Currie and Hyndman, 2006). The hot weak crust has been tectonically thickened even though there is no recent collision. Lower crust flow from higher to lower elevation is expected and there is good evidence that it is occurring (e.g., Gerbault et al., 2005; Kay and Coira, 2009). Current lower crustal flow has also been suggested for the up to 60 km thick crust beneath the Yakutat terrane collision zone of the St Elias range in the corner of the Gulf of Alaska (e.g., Bauer et al., 2014). Such lateral lower crust flow has been argued for several ancient orogenic belts that had thick crust, for example in the southeastern Canadian Cordillera during the Laramide time deformation (e.g., Teyssier et al., 2005; Carr and Simony, 2006; Brown and Gibson, 2006; Simony and Carr, 2011; Gervais and Brown, 2011), and in the European Variscan belt (e.g., Schulmann et al., 2008; Maierová et al., 2014). Although this thick-crust process is occurring only in a few places at present, over geological times it may have occurred in many orogenic belts where the crust was tectonically thickened.

More surprising is that lower crust flow has been concluded for the Basin and Range province of western United States which has a thin crust. In this area there are high crustal temperatures and a thin crust, averaging about 31 km (e.g., Klemperer et al., 1986; Levandowski et al., 2014). Both the thin crust and high temperatures have often been interpreted to result from the ongoing extension at about 1 cm/yr (e.g., Bennett et al., 2003) with a total extension of about a factor of two. Important to the conclusion of lower crust flow in this region are, first, that there is remarkably little variation in crustal thickness over a lateral distance of 800 kilometers in spite of spatially variable extension over the past 17 Ma that reaches up to a factor of two (**Figure 3c**). There is some uncertainty as to whether the crust was thicker before extension. Lechler et al. (2013) concluded that the elevation at the start of extension was little different from the current elevation and therefore the crustal thickness was little changed by the extension. However, part of the Nevada area may have had Mesozoic thickened crust based on paleo-elevation data by Snell et al. (2014). It is also possible that there was addition of crustal material by mafic underplating during extension that made up the amount of thinning. However, it would be surprising if the previous thick crust and the mafic additions exactly matched the extensional thinning to give the current laterally uniform thickness. Second, there are well-developed core complexes where the extension of normal faults exhumes deep crustal rocks in the fault foot walls. In spite of large inferred upper crust normal fault displacements there is no displacement of the Moho in seismic structure data (**Figure 3**) (e.g., Klemperer et al., 1986; Chulick and Mooney, 2002, and references therein). The remarkably flat Moho is interpreted to result from lower crust detachment and flow. If the normal faults did cut the Moho, the Moho offset must be annealed and flattened very rapidly (**Figure 3a**). More likely, the normal faults flatten and sole out in the ductile lower crust (**Figure 3b**). The normal fault extension actually is accommodated by more complex interpreted extensional core complexes (see discussion by Tirel et al., 2008) with detachment between adjacent fault blocks and substantial vertical crustal motion. Tirel et al. concluded a very low initial effective viscosity is required in the lower crust for core complexes to develop. For the Basin and Range, quite fast flow is inferred to both smooth the normal fault displacement in the lower crust over 10's of kilometers and to maintain the constant crustal thickness over distances of 100's of kilometers in less than 10 Ma (e.g., Block and Royden, 1990; Bird, 1991; Wdowinski and Axen, 1992; Kaufman and Royden, 1994; McKenzie et al., 2000; Tirel et al., 2008; Wernicke et al., 2008).

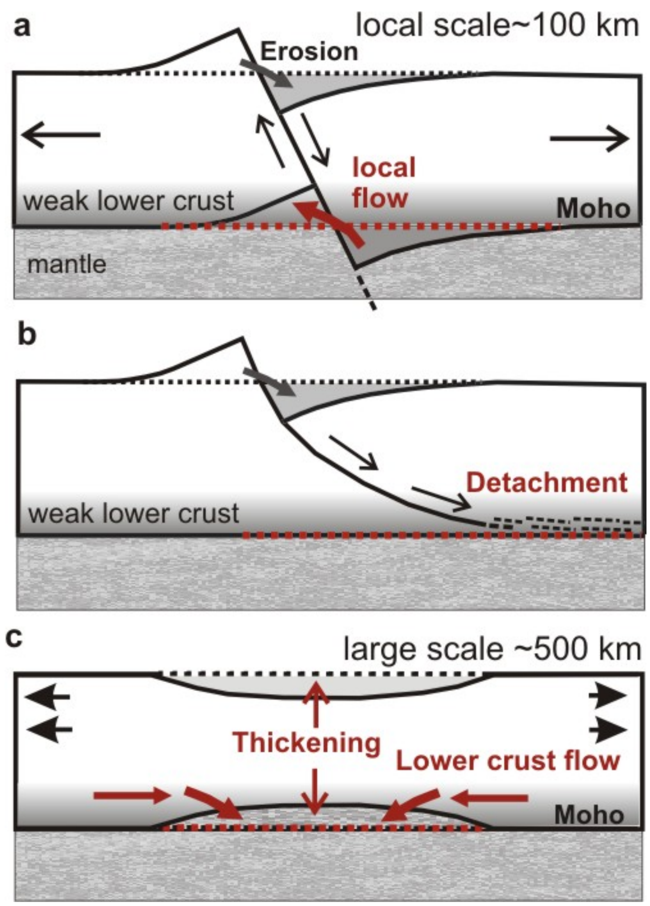

Figure 3. (a) Extensional normal fault that cuts Moho, and is subsequently flattened by lower crust flow, (b) Extensional fault with detachment in the ductile lower crust with no displacement of the Moho, (c) Large scale re-adjustment of extended thinned crust to constant thickness by lower crust flow.

A similar observation of flat Mohos with no offsets associated with past normal faulting has been reported in many areas that are no longer active. Kusznir and Matthews (1988) described deep reflection profiles showing a flat Moho over the area of Cretaceous extension off northwestern Great Britain. Similarly, crustal-scale Eocene normal fault systems in southern British Columbia appear to be listric into the middle or lower crust (e.g., Cook et al., 1992; Parrish et al., 1998). When these normal faults were active, some of the regions may have had thick crust like Tibet and the high Andes, but some may have had thin crust in extensional environments like the current Basin and Range.

Other indications of lower crust flow that I discuss briefly below are that lower crust flow and horizontal shearing provide an explanation for, (1) the wide-spread occurrence of lower crust horizontal reflectors (e.g., Meissner et al., 2006), (2) large areas of horizontal shear deformation that occurred at high temperatures evident in exhumed lower crust rocks (e.g., Dumond et al., 2010).

1.3 Cordillera High Temperatures and Thin Crust; Basin and Range is Not Unique

It has often been assumed that the Basin and Range area is unique in its thin crust and high temperature thermal regime, both resulting from the special condition of ongoing extension. However, recent data that I summarize below indicates that neither is true, the Basin and Range area is not significantly unique in the thin crust or in the high crustal temperatures. Although there are important second order variations, most of the Cordillera from Mexico to Alaska has both high temperatures and thin crust, 31-34 km, in contrast to the stable North America (and global) average of about 40 km (e.g., Hasterok and Chapman, 2007; Chulick and Mooney, 2002). As an example, the area of south-central British Columbia which has not had recent extension since the Eocene and northern British Columbia where no significant extension has been identified, have similar crustal thickness to the Basin and Range (e.g., Kim et al., 2014; Cook et al., 2010; Clowes et al., 2005). The first order approximation of a regionally thin crust and flat Moho in most of the Cordillera has been little appreciated. The uniform crustal thickness is in spite of a complex tectonic deformation history with highly varied current and past tectonics. There has been extension in normal faulting core complexes and major crustal shortening and exhumed high grade metamorphic rocks interpreted to have been brought from the lower crust by large-scale tectonic processes. I conclude from the uniform crustal thickness and flat Moho, that the deep crust of most of the Cordillera has at some time been subject to lower crust flow. In hot backarcs, only a small gravitational potential from thicker to thinner crust apparently is required to level the crustal thickness to within a few kilometers by lower crust flow (see also Jones et al., 1996, and Kaban et al., 2014, for discussions of gravitational potential).

The second important related observation is that, relative to stable eastern North America, most of the Cordillera backarc is surprisingly uniformly hot. Although, like crustal thickness, there are important local variations in temperature, they are small compared to the contrast with the adjacent cold craton. As for most backarcs, the temperatures at the Cordillera Moho are commonly 800-850^{o}C (e.g., Currie and Hyndman, 2006). These high temperatures relative to the stable continent to the east, result in a substantial contribution to elevation. Goes and van der Lee (2002) estimated 1,500 m thermal elevation for the Basin and Range and Hyndman and Currie (2011) reached a similar conclusion for the whole Cordillera. They demonstrated that the thermal elevation effect applies to most of the Cordillera, a remarkably constant 1,600 m relative to stable North America. This thermal elevation explains the high elevations of the Cordillera with a thin crust, average about 1,500 m compared to the east (see Lachenbruch and Morgan, 1990 for discussion of thermal isostasy). With the Cordillera average thin crust, the average elevation would be about 500 m below sealevel if the Cordillera had the cold thermal regime of the craton (Hyndman and Currie, 2011). The high temperatures also allow us to understand the crustal thickness uniformity. The high temperatures result in very low strength in the lower crust and quite low strength in the uppermost mantle.

2. Crustal Thickness of the North American Cordillera

The remarkably constant thickness of the crust in the Cordillera provides important support for the conclusion of lower crust flow. Four principal seismic structure techniques define crustal thickness:

(1) multichannel seismic reflection,
(2) wide angle seismic refraction,
(3) earthquake and noise surface wave tomography
(4) receiver function studies.

They provide complementary data such that the Cordillera crustal thickness is very well constrained. In a number of areas seismic reflection gives good spatial resolution of a few kilometers that shows the Moho is flat locally, but this method has very incomplete coverage. In contrast, noise tomography provides coverage over the whole Cordillera and shows the uniformity of crustal thickness, but this method has low spatial resolution. Crustal thicknesses from all of the methods are generally in good agreement, and show that in most of the Cordillera from Mexico to Alaska the crust is surprisingly thin, 31-35 km, with the Moho being locally flat. A few significant local variations in crustal thickness have been reported, for example a few kilometers in the currently extending northwestern US Basin and Range Province (e.g., Holbrook, 1990) that may indicate transients in crustal thickness that have not yet been smoothed. Such areas are important for understanding the rate at which crustal thickness variations are smoothed by lower crust flow.

However, the data described below shows the crustal thickness over most of the Cordillera is very constant. The Cordillera average crustal thickness is 33±5 km from the compilation of Hasterok and Chapman (2007) compared to 40±4 km for the adjacent stable areas to the east. The Cordillera variability is even less, about 33±2 km if a few special areas are excluded, especially thicker crust areas where formerly stable thick cold lithosphere has been thinned recently such as the Colorado Plateau and adjacent areas. These areas appear to have been cold and stable with about 40 km crusts until recent lithosphere thinning and uplift. I also exclude the recently thinned crust of coastal areas of western California and northwestern Mexico near the extensional Gulf of California. I provide a summary below of crustal thicknesses and Moho topography.

2.1 Seismic Reflection and Refraction

There are two well-established techniques for crustal and upper-mantle seismic structure that gave good estimates of the depth to the Moho. They are deep seismic reflection (mainly 'Vibroseis') and wide-angle or refraction seismic surveys. The reflection times from subsurface layers give the depth, providing the seismic velocity of the section is known. In some areas, especially the Cordillera, the Moho is seen as a strong reflector. In others, especially stable areas, the base of the crust may be defined by a change in the reflection character with depth. The crust commonly has complex reflections from composition inhomogeneities, whereas the upper mantle is more uniform and seismically transparent. Wide-angle or seismic refraction, generally with large explosion sources, gives seismic velocities with depth as well as defining layering within the earth, but it has low spatial resolution. In the compilation of Chulick and Mooney (2002) the Cordillera crust is generally thin and quite uniform, although as noted above

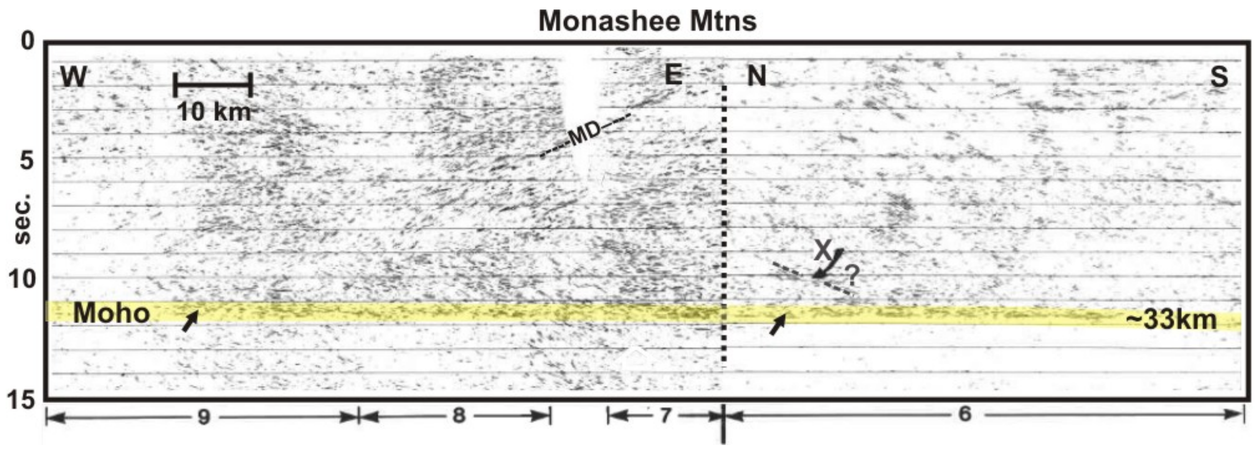

Figure 4. 150 km long reflection section across southeastern British Columbia showing the thin crust and very flat Moho in an area of both contractional and extensional faults with large offset. The seismic lines 9, 8, and 7 are east-west; line 6 connects and is north-south. MD is the Monashee decollement. 'X' is a possible listric fault (after Cook, 1995).

there are a few areas with substantial variations associated with recent tectonics, that we exclude from our primary backarc analysis.

Examples of deep seismic reflection data across the Basin and Range area of the U.S. Cordillera, are given by Allmendinger et al., (1987a) and Catchings and Mooney (1991). There are no clear Moho offsets associated with extensional faulting. The crust is thin with only small variations in the crustal thickness, slightly thicker on the western and eastern ends of their Cordillera profile. Another area where multichannel reflection and wide-angle refraction data give well-defined nearly constant crustal thicknesses and flat Moho is southern British Columbia just north of the U.S. border, by the Canadian Lithoprobe program, (e.g., Clowes et al. 1995; Cook et al., 2010, and references therein). In this area there has been little tectonic deformation since the Eocene except for some more recent transcurrent motion in the western portion. From surface structural data, Parrish et al. (1988) estimated at least 30% crustal extension in the Eocene. As noted by Cook (1995), the nearly 25 km of structural relief identified in outcrop and on seismic reflection data is not evident in the Moho which is remarkably flat over the horizontal resolution of a few kilometers. As in the Basin and Range to the south, the crust is several kilometers thicker under the higher elevation coast belt and toward the eastern Cordillera beneath the higher elevation Monashee complex, as expected by Airy isostasy (see compilation of reflection sections by Cook, 1995). Otherwise the crust is about 32 km (11.5 seconds reflection time) across the Intermontane Belt, with no significant Moho offsets (**Figure 4**). In northern British Columbia and the Yukon where no extension has been identified, Clowes et al. (2005) found crustal thicknesses mainly about 33 km with a few areas of 30 km and 36 km. A uniform thin crust of about 32 km also has been found for most of the Alaska backarc south of the Brooks Range (e.g., Beaudoin et al., 1992; 1994; Fuis et al., 2008; Ruppert, 2008).

2.2 Noise Tomography and Receiver Functions

A similar uniform thin crust has been obtained for the Cordillera from two other methods, tomography surface wave inversions that use a range of frequencies to resolve velocities at different depths, and receiver functions that use phase conversions at the Moho from distant earthquake sources. These analyses have given us a much improved mapping of crustal thickness with consistent methods over the whole Cordillera. The crustal thickness contrast in western U.S.A. from Cordillera to craton is especially well resolved using noise tomography data from the closely spaced stations of USArray (e.g., Shen et al., 2013; Levandowski et al., 2014).

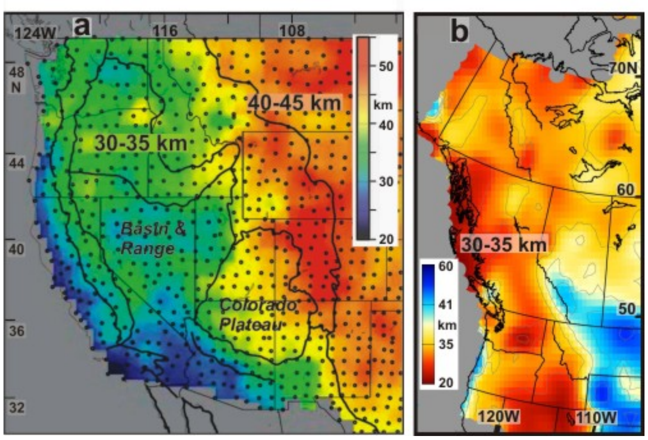

Figure 5. Noise tomography crustal thickness maps of western U.S. and western Canada showing uniform thin Cordillera crust; data from (a) Levandowski et al., 2014, and (b) Kao et al., 2014. Note different color scales.

In western Canada and Alaska, although

there are fewer seismic stations so lower spatial resolution, the thicknesses are also well resolved. There is a clear division between the Cordillera (30-35 km) and the stable craton to the east (40-45 km) (e.g., Kao et al., 2014) (**Figure 5).** The area of north-central Canada also has a thin crust partly because of near sea level elevation and partly because of intermediate thermotectonic ages.

There are numerous examples of receiver function analyses that give crustal thicknesses that are very consistent with those from the other seismic structure methods, mainly 31-35 km. For western United States, see Ramesh et al. (2002), and for western Canada in the area of the Lithoprobe data noted above, see Cassidy (1995). Data are given for Alaska by Ai et al. (2005) and O'Driscoll and Miller (2015). California is complex because of very recent transcurrent fault juxtaposition of forearc and backarc and local recent tectonic thinning, but most of that area still has the characteristic backarc thin crust (Zhu and Kanamori, 2000). These data show that, whereas there are small variations in crustal thickness in the Cordillera, they are much less than the contrast with stable areas in central and eastern North America.

2.3 Western Europe

There is a related association of hot thin lithosphere and thin, about 30 km, crust beneath western Europe, compared to the cold Fennoscandian Shield to the northeast, with crustal thicknesses from 40-45 km. The crust is generally a few kilometers thinner in western Europe compared to the North American Cordillera so the corresponding thickness difference between western Europe and the adjacent shield is larger. The especially thin crust of western Europe results in lower elevations, and there is little difference in elevation beneath western Europe and the adjacent shield (see Figure 2 of Tesauro et al., 2008). Relative to the shield, the thermal elevation effect of western Europe balances the thin crust. Western Europe, most of which is a current or thermally recent backarc, is equivalent to the Cordillera, and the Fennoscandian Shield with a stable thick lithosphere is equivalent to the Canadian Shield. Isostasy crustal density corrections have not been made, but elevation difference between the two regions for the same crustal thickness is similar to that for North America, about 1,600 m. This difference is in agreement with the contrast in thermal regime implied by the lithospheric thickness difference between western Europe of 50-100 km compared to about 200 km for the Fennoscandian Shield (e.g., Plomerová and Babuška, 2010). From the inferred high temperatures, it is likely that much of western Europe is subject to lower crustal flow and detachment.

3. Temperatures in the Lower Crust

Many models of subduction zones have high temperatures near volcanic arcs, but we now recognize that high temperatures usually extend across the entire continental backarc (e.g., Hyndman et al., 2005; Currie and Hyndman, 2006). Although some lateral temperature variations are resolved in the North American Cordillera, most are small compared to the large contrast with the adjacent cold Canadian Shield and other stable areas to the east. There are five main constraints to lower crust temperatures that are complementary and give consistent results.

(1) surface heat flow and heat generation

(2) temperature dependence of seismic velocity in the upper mantle

(3) xenolith temperature-pressure (depth)

(4) Thermal control of elevation

(5) lithosphere thickness assuming the base of the lithosphere is thermally controlled

An additional constraint that supports high temperatures in the lower crust is the effective elastic lithosphere thickness Te that I discuss below. The different methods provide complementary data such that the Cordillera thermal regime is now well constrained. Heat flow-heat generation and mantle xenoliths can give good local thermal estimates but with very limited and irregular coverage. In contrast, surface wave velocities from seismic tomography provide coverage over the whole Cordillera and show the uniformity of thermal regimes, but have low spatial resolution. Within the recognized uncertainties, crustal and upper mantle temperatures from all five methods are generally in agreement and have shown that, in most of the Cordillera from Mexico to Alaska, the lower crust is very hot, 800-850^{o}C at the Moho.

3.1 Surface Heat Flow Temperature Estimates

A well-established constraint on deep-crustal temperatures and lithosphere thicknesses is provided by surface heat flux measurements (e.g., Chapman and Pollack, 1977; Chapman, 1986; Morgan and Gosnold, 1989). However, extrapolating surface heat flow to deep temperatures has a large uncertainty. In addition to the measurement uncertainties and near-surface thermal disturbances, there is the effect of variations in near-surface radioactive heat generation that affect the heat flow directly but have only a small influence on deep temperatures. The use of regional heat flow maps for estimating deep temperatures therefore can be misleading. This uncertainty can be much reduced if we have measurements of upper crust radioactive element abundance (U, Th, K), and can allow for variations in radioactive heat generation. The high temperatures in the Cordillera backarc, were shown for Washington and Oregon by Blackwell et al. (1990) and by Lewis et al., (1992) for southern British Columbia allowing for variations in upper crust heat generation, and by Lewis et al. (2003) for the northern Canadian Cordillera. All three areas give similar estimates of lower crustal temperatures. An example profile where a

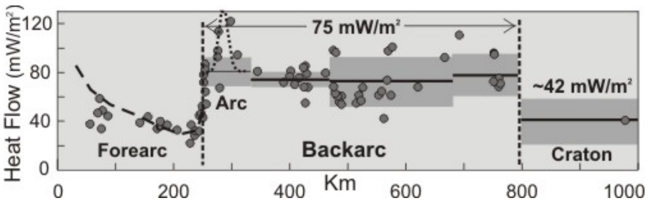

Figure 6. Heat flow data across the southern Canadian Cordillera, corrected for variations in near surface heat generation.

first-order correction to the heat flow has been made for the effect of variations in near-surface radioactive heat generation is shown in **Figure 6** (Lewis et al., 1992; Hyndman and Lewis, 1995; Hyndman, 2010). There is a clear contrast between the nearly laterally uniform corrected high heat flow across the Cordillera and the low heat flow for the stable areas to the east. Heat flow in the Cordillera is almost double that of the shield. **Figure 7** shows the average and variability of temperature-depth estimates from the heat flow-heat generation data and other constraints for the northern Cordillera (Hyndman et al., 2009).

Although heat flow-heat generation data give regionally similar lower crustal temperatures for most of the Cordillera, there are some variations in the estimates. Upper crust radioactive heat generation variations usually have a limited effect on deep temperatures, but very large heat generation differences can be important. Examples of measured unusually high near-surface heat generation and resulting higher than average inferred crustal temperatures are the southeastern British Columbia Omineca Belt and northern British Columbia-southern Yukon where estimated Moho temperatures are about 900^{o}C (Lewis et al., 1992; Flück et al., 2003; Lewis et al., 2003).

3.2 Temperatures From Upper Mantle Seismic Velocities

For large scale mapping of regional deep temperatures over the whole Cordillera, the best estimator is temperature-dependent seismic velocity in the upper mantle. Within the continental crust, seismic velocities are mainly controlled by rock composition. However, in the upper mantle, velocity is mainly controlled by temperature; higher temperatures give lower velocities. The second-order effect of upper-mantle composition can be corrected, especially using mantle xenoliths.A sometimes complicating factor for this temperature constraint is the poorly constrained effect of upper mantle partial melt on the velocity,

especially in parts of the U.S. Cordillera (e.g., discussions by Hammond and Humphreys, 2000; Dixon et al., 2004), such as beneath the Yellowstone region, so I show mainly examples of this temperature constraint for the Canadian Cordillera where partial melt appears generally not to be sufficiently wide-spread to be important. However, the effect of partial melt remains a significant source of uncertainty in temperature estimates from seismic velocities. Areas with partial melt may have biases in estimated temperatures by this method (e.g., Schilling et al., 2006). Velocity estimates can come from both local wide angle seismic structure studies and from regional seismic tomography. Temperatures can be estimated from both compressional and shear wave velocities but most analyses are for the better determined shear wave estimates. Tomography can use both distant earthquakes and "noise" sources, noise tomography.

Figure 7 gives temperature-depth and its variability for the Canadian Cordillera and adjacent craton (Hyndman

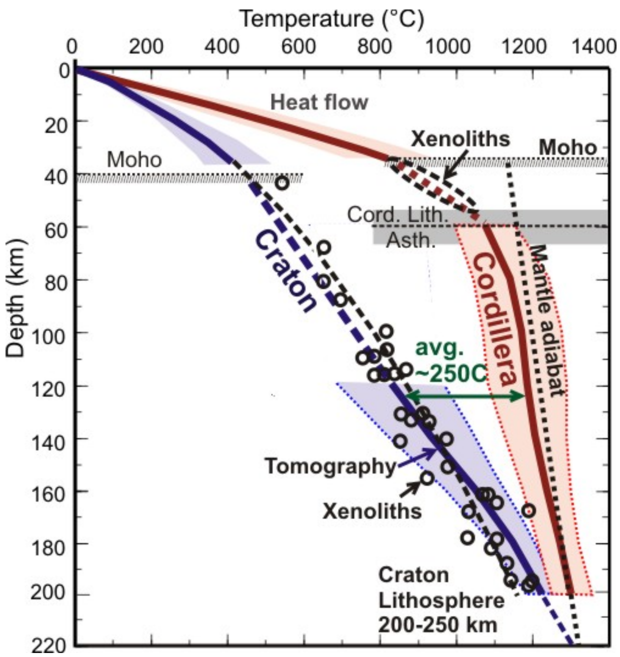

Figure 7. Average northern Cordillera temperature-depth estimates based on surface heat flow-heat generation data, on mantle seismic velocities (Hyndman et al., 2009) and on xenoliths (Canil, 2008; Greenfield et al., 2013). The area is shown in Figure 8.

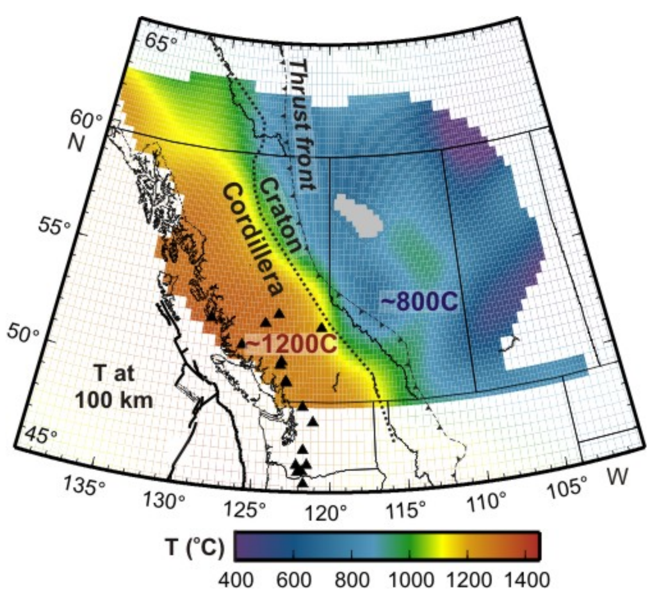

Figure 8. Map of estimated temperatures of the Canadian Cordillera and adjacent craton at a depth of 100 km (data from Hyndman et al., 2009). The average and variability in T-Z estimates is shown in Figure 7.

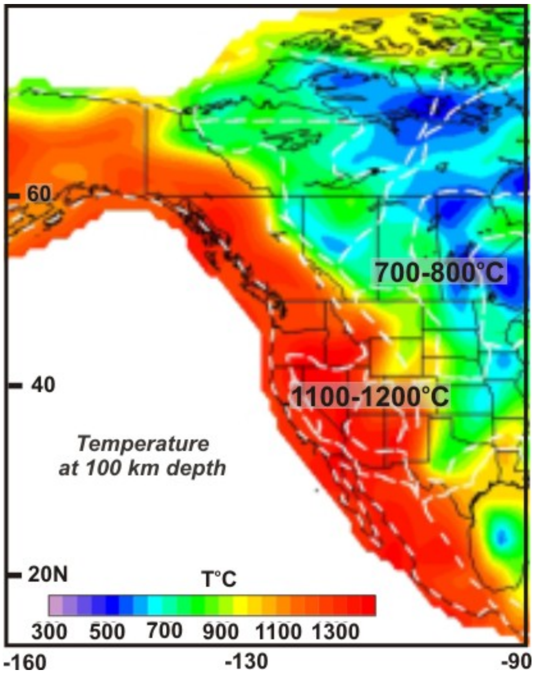

Figure 9. Seismic tomography velocities at a depth of 100 km for the Cordillera from Mexico to Alaska, illustrating the first order uniform high temperatures (after Tesauro et al., 2014).

et al., 2009). **Figure 8** gives a map of estimated temperatures at a depth of 100 km for that area (Hyndman et al., 2009) based on velocity data from van der Lee and Frederiksen (2005). The absolute temperatures have a significant uncertainty especially associated with attenuation (e.g., Goes and van der Lee, 2002) but the depth to the base of the conductive lithosphere is robust. The Cordillera-craton mantle temperature contrast is very sharp, for example at the Rocky Mountain Trench in southern British Columbia. Some of the variability shown in the calculated temperatures represents real temperature variation and some comes from measurement uncertainty. In any case, the variability within the Cordillera backarc and the adjacent craton is much smaller than the difference between the two regions. The maximum Cordillera-craton temperature difference of about 500°C occurs at about 60 km depth and the average temperature difference to 200 km is about 250°C.

In several recent studies, Bedle and van der Lee (2009) and Schaeffer and Lebedev (2014; and models that they summarize) found low velocities in the Cordillera upper mantle from western Mexico to southern Alaska indicating consistently high temperatures. The inferred temperatures are remarkably high and laterally uniform across the Cordillera and much lower and quite uniform in the adjacent craton (e.g., Tesauro et al., 2014; Kaban et al., 2014) (**Figure 9**). The very constant temperature at 100 km, over most of the Cordillera indicates that, although the base of the lithosphere may be at a variable depth, it is everywhere shallower than 100 km. As discussed below, this depth indicates a Moho temperature of greater than about 700°C. This is a minimum temperature; our other constraints indicate that the Moho temperature is at least 100°C higher.

3.3 Xenolith Temperature-Pressure (Depth).

Samples of rocks from the deep crust and upper mantle occasionally are carried to the surface entrained in volcanic magmas in the Cordillera and through kimberlite pipe eruptions (“diamond pipes”) in the craton. Commonly the exhumation rates are rapid enough that chilling retains the mineral equilibria representative of the temperature and pressure (depth) at their source, allowing calculation of temperature-depth profiles at the time of emplacement. Reliable temperatures can be obtained for both the Cordillera and for the craton. However, depth calculations are more reliable for the xenoliths from cratonic regions which have minerals with better pressure-sensitive equilibria than for the high temperature Cordillera. However, through some estimators and through indirect methods the xenolith origin depths for the Cordillera have been constrained to useful accuracy.

Upper-mantle xenoliths have been recovered from numerous localities in the Cascadia backarc that give Moho temperatures consistent with estimates from shear wave velocities, Vs. Ross (1983) estimated 1000°C at a depth of about 40 km in 8 localities in British Columbia. This temperature estimate gives about 850°C for an average Cordillera 33 km Moho assuming a conductive gradient. Saruwatari et al. (2001) estimated 900°C at 35-50 km depth in southern British Columbia to Alaska. A detailed study by Harder and Russell (2006) of the Llangorse/Edziza volcanic field in northwest British Columbia constrained the Moho temperature to be 800-850°C. Greenfield et al. (2013) also estimated a Moho temperature at 33 km of 825±25°C for southern British Columbia (**Figure 7**). For all the Cordillera studies, the estimated temperatures at the Moho are 800-850°C, very consistent with the other constraints.

For comparison, the craton and stable platform, from numerous studies from mantle xenoliths give temperatures (e.g., MacKenzie and Canil, 1999; Canil, 2008 and references therein) that are very consistent with temperatures from Vs, 400-500°C, (**Figure 7**) and show the contrast with the backarc Cordillera. There are well-resolved lateral variations in the craton but they are small compared to the contrast with the Cordillera. Other cratons globally give similar temperatures (e.g., Griffin et al., 2004). The base of the thermal lithosphere is usually at 200–250 km. This depth is similar to that obtained from seismic and magnetotelluric data (e.g., Eaton et al, 2009).

3.4 Thermal Regime From Surface Elevation

Simple mapping of surface elevation and crustal thickness provide a strong regional constraint to deep temperatures through the effect of temperature on density, thermal isostasy. It has been concluded that surface elevation is controlled mainly by the thermal regime after allowance is made for variations in crustal thickness and crustal density, the latter estimated from average crustal seismic velocity (Hyndman and Currie, 2011). Although a systematic difference in upper mantle composition between the Cordillera and adjacent stable areas is indicated by xenolith data, the temperature difference appears to have a dominant control on elevation. A surprising observation is that the high-elevation Cordillera has a thinner crust, about 33 km average, compared to the adjacent low-elevation craton, about 40 km, which is a clear violation of simple Airy Isostasy. There is no Cordilleran mountain root. The averages from a compilation of surface elevation and crustal thickness data for North America by Hasterok and Chapman (2007) are 33±5 km for the Cordillera and 40±4 km for the stable Canadian Shield. Becker et al. (2013) found that much of the variations in surface elevation within the U.S. Cordillera can be explained by crustal thickness and crustal density variations, but suggested a dynamic component for some of the elevation variability. However, Hasterok and Chapman (2007) and Hyndman and Currie (2011) showed that using the average crustal velocity to estimate average crustal density, all of the elevation variations can be explained by crustal thickness and density within the recognized uncertainties. Goes and van der Lee (2002) estimated 1,500 m thermal elevation for the Basin and Range relative to stable North America, but it can now be seen that this thermal elevation applies to the whole Cordillera.

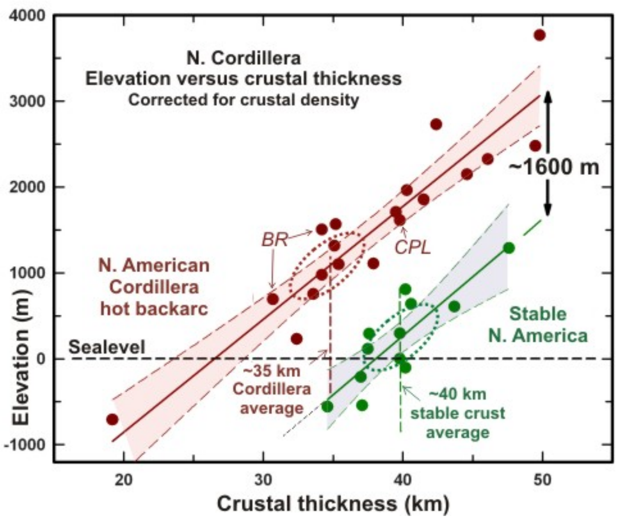

Figure 10. Elevation versus crustal thickness showing 1600 m difference in elevation for the Cordillera compared to the adjacent stable areas for the same crustal thickness (after Hyndman and Currie, 2011; data from Hasterok and Chapman, 2007). CPL, Colorado Plateau; BR, Basin and Range.

Much work has been done on the role of mantle traction and mantle dynamics on elevation (e.g., reviews by Braun, 2010 and Flament et al., 2013). Although the latter may be important, especially for cratons, the simple relations between crustal thickness and elevation, corrected for crustal density, indicates that base-of-lithosphere traction must be second order to the effect of thermal isostasy for the Cordillera. Also, for the Cordillera and other hot backarcs, the estimated asthenosphere viscosities appear to be too low to allow significant lithosphere basal traction and asthenosphere convective buoyancy forces for that area (see also Levandowski et al., 2014).

Although most of the Cordillera has crustal thicknesses of 30-35 km and the variability is low, there are significant variations in a few areas which can be used to illustrate the difference in thermal buoyancy between the Cordillera and shield more graphically. Within each of the regions, after correction for the usually small variations in crustal density, there is the expected correlation between crustal thickness and elevation for Airy isostasy. Plots of elevation vs crustal thickness show a remarkable separation between the Cordillera and craton of 1,600 m for the same crustal thickness, with no overlap (Hyndman and Currie, 2011). The average deviation from the two linear elevation-vs-crustal thickness relations is about ±200 m in elevation and less than ±2 km in crustal thickness (**Figure 10**). This scatter represents about ±30°C in the average temperature to 200 km depth for the sites within each of the two regions. This difference in elevation for the same crustal thickness is concluded to be mainly due to thermal density reduction in the Cordillera (e.g., Hyndman et al., 2005), although there are some differences in mantle density for the two areas due to composition (e.g., Tesauro et al., 2014). An average temperature difference of 250^{o}C to 200 km depth where Cordillera-craton temperature converge is required for a thermal origin of the elevation difference, in good agreement with the temperature-depth differences from the other thermal constraints discussed above. The high elevation for most of the Cordillera with a thin ~33 km crust (see Figure 5) is strong evidence for low density due to high temperatures. If the thin crust Cordillera had the density and thermal regime of the craton, the elevation would generally be below sealevel.

As noted above, there is a related comparison between the hot thin lithosphere beneath western Europe, with crustal thicknesses of about 30 km, compared to the cold Fennoscandian Shield to the northeast, with crustal thicknesses of 40-45 km. The crust is generally thinner in western Europe compared to the North American Cordillera so the elevation is lower, and there is little difference in elevation beneath western Europe and the adjacent shield (see Figure 2 of Tesauro et al., 2008). Although corrections for variations in crustal density have not been applied, the crustal thickness and elevation averages for these two European regions agree very well with the plots of crustal thickness versus elevation for North America. Western Europe, most of which is a current or thermally recent backarc, is equivalent to the Cordillera, and the Fennoscandian Shield is equivalent to the Canadian Shield (**Figure 9**). The elevation difference between the two regions for the same crustal thickness is also approximately 1,600 m. Goes et al. (2000) showed that there are high temperatures and thin lithosphere beneath western Europe, similar to under the North American Cordillera.

3.5 Thermal Regime from Lithosphere Thickness (LAB)

A strong constraint to lower crust temperatures may be provided by the thickness of the lithosphere from seismic structure methods (e.g., Shen et al., 2013), assuming that the base of the lithosphere (lithosphere-asthenosphere boundary, LAB) is thermally controlled at the adiabatic temperature for that depth (e.g., Hansen et al., 2015). This method has been little used but has significant potential. More work is needed to make this lower crust and upper mantle temperature constraint secure, especially, (1) that the base of the lithosphere is at a reliable known adiabatic temperature for that depth and, (2) that the temperatures above the LAB represent a conductive gradient. There may be a downward transition from conductive to advective adiabatic temperature gradients, at the base of the lithosphere, but the common receiver function LAB reflection suggests that the transition is quite abrupt. In the several temperature estimates below, I have taken the adiabat as the temperatures from upper mantle seismic velocity data at depths greater than 60 km in the northern Cordillera as shown in **Figure 7** (Hyndman et al., 2009) which is in general agreement with the temperature estimates of Hansen et al. (2015) and with the xenolith-based temperatures by Greenfield et al. (2013). For a higher temperature adiabat temperature-depth estimate such as by McKenzie and Bickle (1988), the Moho temperatures in the Cordillera are even higher.

As well as a reflection or phase conversion in receiver functions, the LAB sometimes is seen as a deep reflection and as a velocity boundary in wide-angle seismic structure data. The reflection requires an abrupt change in seismic impedance, not just a change in velocity gradient with depth. There may be a negative velocity gradient immediately below this boundary. The observed seismic velocity gradients appears to require a contrast in mantle hydration, fertility, or melt content, in combination with a vertical gradient in velocity anisotropy (e.g., Fischer et al., 2010). The seismic wavelengths used in receiver functions are long, so the depth resolution is at best a few kilometers. The differences between Ps and Sp depths obtained by Levander and Miller (2012) indicate a significant uncertainty. However, in a number of locations the base of the lithosphere corresponds well to the downward change from conductive to adiabatic in temperatures from seismic velocities (**Figure 7).** Below this boundary there must be strong small scale convection that maintains the the low gradient adiabatic temperatures.

Another approach for determining the lithosphere thickness is from mantle xenoliths. Xenoliths estimate the LAB by the maximum depth of origin and by the depth of significant sheared textures. In British Columbia lithosphere thicknesses of 52-66 km have been estimated from xenoliths by Harder and Russell (2006, and references therein), again consistent with an average 33 km deep Moho of 800-850°C from the other estimators. In summary, most estimates have the Cordillera lithosphere thicknesses of from 60 to 70 km. This depth range constrains the Moho temperature to about 800-850°C assuming a conductive gradient in the overlying lithosphere and assuming that the temperature at the base of the lithosphere is the asthenosphere adiabatic temperature at that depth as shown in **Figure 7**.

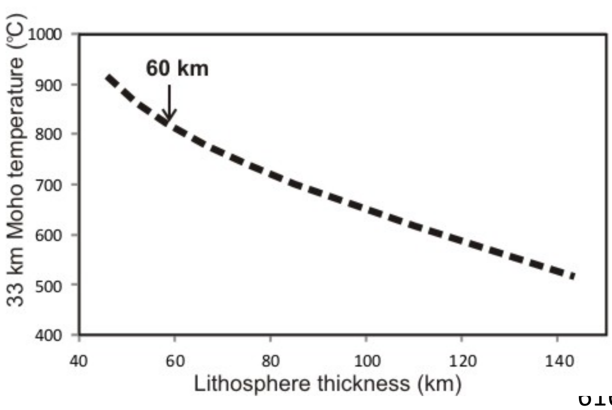

Figure 11. Estimated Moho temperature versus lithosphere thickness, assuming the base of the lithosphere is at the upper mantle adiabatic temperatures.

Figure 11 shows an estimate of the Moho temperature for varying lithosphere thicknesses. The temperature-depth relations for varying lithosphere thicknesses are from Flück (2009) which are very similar to those of

Chapman and Pollack (1977). From receiver function definition of the base of the lithosphere, Hansen et al. (2015) estimated a temperatures of 1200-1400^{o}C at 60-85 km beneath the western United States. For a normal temperature-depth profile (e.g., **Figure 7**), these temperatures and depths correspond to a temperature of 800-900^{o}C at an average Cordillera 33 km depth Moho. This is in good agreement with our other estimators of Moho temperatures. Similar lithosphere thicknesses of 60-80 km were found by Lekić and Fischer (2014) and Hopper et al. (2014) for most of the western United States backarc, with variations of about 10 km. Levander and Miller (2012) found 62±3 km and Kumar et al. (2012) and Yuan (2011) found 70-80 km. There are somewhat greater thicknesses beneath the Colorado Plateau and adjacent areas. This suggests a lower temperature gradient, but because the crust is thicker than the Cordillera average, the estimated temperature at the deeper Moho in that area is similar. In the Alaska backarc O'Driscoll and Miller (2014) found an average lithosphere thickness of about 75 km. In British Columbia and the Yukon, Clowes et al. (2010) and Cook et al. (2004) summarized Cordillera lithosphere thickness of 50-70 km from a number of reflection and wide angle seismic structure studies, giving estimated temperatures that are again about 800-850^{o}C with an uncertainty of about ±50^{o}C. In the few Cordillera areas with thicker than average lithosphere, for example for an 80 km thick lithosphere, the estimated Moho temperature is 700-750^{o}C from this method (see **Figure 11**).

For lithosphere less than about 80 km thick, there is low sensitivity of elevation to lithosphere thickness, less than about 15% smaller predicted thermal elevation for 80 km vs 60 km, because in backarcs the deeper part of the thermal regime has everywhere approximately the same convective adiabat to the reference depth of about 200 km. The thermal elevation anomaly relative to the craton reference decreases more rapidly for lithosphere thicknesses greater than about 100 km, reaching zero at the reference depth of about 200 km for craton lithosphere. Hyndman and Currie (2011, Appendix) showed that the difference in predicted elevation due to thermal isostasy for a 60 versus 80 km lithosphere is about 15% or 240 m. The average deviation from the best fit elevation versus crustal thickness line for the Cordillera is about 200 m, so the thermal elevation effect (after correction for crustal density) should be evident for a 75 km or greater lithosphere compared to 60 km.

From all of the temperature constraints I estimate the variability in temperatures at the Moho to be about ±50^{o}C. Some part of this variability in temperature estimates is real variability and some part of measurement uncertainty. This variability is only 10% of the difference between the average Cordillera and craton.

4. Origin of Cordillera Backarc High Temperatures

Most backarcs globally have high temperatures so likely have a common origin (e.g., Currie and Hyndman, 2006). The high temperatures have been explained by rapid upward convective heat transfer beneath a thin lithosphere (see Hyndman et al., 2005, for discussion). This process was suggested by Hasebe et al. (1970) who were concerned with the high heat flow in the Japan Sea backarc. Based on high heat flow, high electrical conductivity and other results that presented strong evidence for high temperatures and partial melting at shallow mantle depths beneath the Canadian Cordillera, Gough (1986) proposed “mantle upflow tectonics”. Many models of backarc convection have assumed one large-scale circulation cell driven by the downward traction and negative thermal buoyancy of the cold subducting oceanic plate. This model is conceptually reasonable but it has proved difficult to produce the observed uniform high heat flow across the backarc with such models (e.g., Currie et al., 2004; Kukačka and

Matyska, 2008). Heat should be lost from the top of the cell such that temperatures and surface heat flow decrease toward the arc, unless the convection speed is much faster than plate motion rates. This decrease is not observed. Regional small-scale convection that maintains adiabatic temperatures below about 60 km seems to be required, with local flow rates faster than relative plate-motion rates (e.g., Currie et al., 2004; Nyblade and Pollack, 1993; Arcay et al., 2006). **Figure 2** shows a schematic small-scale convection model. In a few areas such as the Basin and Range province, present or recent crustal extension may have an additional thermal effect (e.g., Lachenbruch and Sass, 1978), although it is not clear in our thermal constraints. Similarly, in oceanic backarcs where extension is occurring, it is difficult to separate the thermal effect of extension from that of convective heat transport in the underlying shallow asthenosphere. However, Watanabe et al. (1977) suggested that even in these basins, small-scale convection is needed to explain the thin lithospheres and the high heat flow that has been maintained for long times after the basins opened.

An explanation for shallow vigorous convection beneath the Cordillera and other backarc lithospheres is that the mantle viscosity is substantially lowered by incorporation of water and other volatiles expelled from hydrated minerals in the underlying subducting oceanic plate with increasing downdip temperature and pressure. The backarc convection system is poorly understood, but vigorous convection may mix the water throughout the whole backarc asthenosphere wedge. Another possibility for spreading oceanic plate dehydration fluids landward is that episodes of flat-slab subduction carry water far inland, initiating small-scale convection. Mantle rocks containing even quite small amounts of water in the mineral structure (>50 ppm), have a much lower effective viscosity than dry mantle rocks (e.g., Karato and Wu, 1993). Dixon et al. (2004) summarized the evidence for very low mantle viscosity beneath the

Cordillera current and recent backarc of the western U.S.A. and concluded that such low viscosities require significant water in the upper mantle, as well as high temperatures that are close to the solidus. In areas where the landward boundary of the backarc is a craton or an old platform, such as western North America, the shallow asthenosphere convection may be limited landward by thick, refractory lithosphere. However, the original craton margin rifting and associated asthenosphere upwelling may have extended and heated a considerable width of the margin of the craton or platform. This would have allowed subsequent backarc shallow thermal convection to continue beneath the thinned region to the edge of the unextended craton lithosphere. Royden and Keen (1980) illustrate such margin lithosphere thinning and heating from craton rifting by the opening of the Labrador Sea ocean basin between Labrador and Greenland. This provides one model for the margin lithosphere thickness and thermal regime at the start of subduction. The backarc also may be widened by the addition of accreted terranes, which has occurred in western North America.

Globally there are a few cool backarcs, mainly where there is flat-slab subduction such that there is no space for small-scale convection between the base of the backarc lithosphere and the underlying nearly-horizontal subducting slab (see discussion by Currie and Hyndman, 2006). These areas usually are also characterized by little or no arc volcanism. For a well-studied example, the Peru flat slab area is discussed by Gutscher et al. (2000).

5. Duration of High Temperatures in Former Backarcs

In the discussion and data shown above, I included the northern Canadian Cordillera in the backarc although subduction was cut off and stopped on that margin 40-50 Ma ago with the development of the transform Queen Charlotte Fault system (e.g., Engebretson et al., 1985;

Hyndman and Hamilton, 1993; McCrory and Wilson, 2013). The same is true in California where subduction was cut off more recently by the San Andreas Fault system (e.g., Atwater and Stock,1998). However, these backarcs must have cooled quite slowly following the termination of subduction such that the estimated lower-crust and upper-mantle temperatures are little different from those landward of the presently-active Cascadia subduction zone. There must be a finite life to the high temperatures in backarc mobile belts after the source of heat is removed, as most ancient mobile belts active in the Paleozoic or earlier no longer exhibit the characteristic backarc high lithosphere temperatures. In the backarc convection model, the vigorous free convection should decline following the termination of subduction. The processes involved are undoubtedly complex, including slab window and slab break-off effects, and may take tens of millions of years before temperatures start to decline significantly. However, lithosphere cooling and thickening are probably conductive following the increase in upper-asthenosphere viscosity due to water loss through partitioning into arc and backarc melt fractions, and through upward diffusion. The cooling time constant may be estimated from compilations of present heat flow, thermal elevation, and inferred lithosphere temperatures relative to the age of the most recent thermotectonic event defined by igneous activity, metamorphism, volcanism, etc. (e.g., Currie and Hyndman, 2006). The thermotectonic age is assumed to correspond approximately to the time since termination of subduction, commonly due to continental or terrane collision, and therefore of subduction water input. Collision may be a long-duration process and as noted earlier, there may be a delay before the start of decline of lower crust temperatures. For example, it has been at least 25 m.y. since the initial India-Asia collision and convergence is continuing. The most rapid decrease in heat flow appears to be in the several hundred m.y. following the last thermotectonic event, and the data suggest a 300-500 m.y. time constant (see also Sleep, 2005). A similar cooling and lithosphere thickening time is suggested by several examples, especially the former backarc mobile belt of Appalachia in eastern North America.

The Appalachian former backarc mobile belt in eastern North America, in which the last significant deformation occurred at about 300 Ma, is now cool and stable. Although heat flow and other thermal data suggest that it is still somewhat warmer than the cratons (e.g., Pollack et al., 1993), at least some of the difference may be due to greater upper-crust heat generation (Mareschal and Jaupart, 2004). The residual thermal elevation above that for the craton however, remains quite significant at 600-800 m (e.g., Hasterok and Chapman, 2007). The main Innuitian orogeny of the Canadian arctic islands is of a similar age (e.g., Trettin, 1991). The thermal regime for that area is poorly constrained but the thin crust and low elevation of that region suggests an intermediate thermal regime between the Cordillera and craton. In contrast to these examples, the northern Canadian Cordillera, where margin subduction was cut off by the Queen Charlotte-Fairweather transform fault zone in the Eocene, 40-50 m.y. ago, still has high heat flow and inferred high temperatures at depth (Lewis et al., 2003), similar to backarcs of currently active subduction zones. From these examples, backarc cooling must be limited before 50 m.y. but considerable after 300 m.y.

Currie and Hyndman (2006) used a simple conductive model with an initial thin backarc lithosphere and an underlying adiabatic asthenosphere to illustrate these features. This simple model is the limiting case of abrupt termination of convection to a depth of 250 km (approximate thickness of craton lithosphere) at the time of termination of subduction. The heat flow data are not corrected for variations in upper-crust heat generation, and the effect of erosion of the high-radioactivity upper crust should give a somewhat lower heat flow at long times than the simple cooling model. The heat flow as a function of age predicted by the model is in general agreement with that observed, which suggests that the termination of shallow free convection generally occurs a few tens of millions of years after subduction stops.

6. Effective Elastic Thickness, Te

The effective elastic thickness, Te, is an important constraint to the rheology of the deep crust and upper mantle that can be mapped over the whole Cordillera. It is closely related to the temperature regime. Te is an estimate of the thickness of the crust and upper mantle that can maintain elastic strain due variations in topographic and density loads for geological time scales. For areas with a single elastic layer, i.e., very hot or very cold lithosphere, Te may approximate the depth of the brittle-ductile transition. For intermediate thermal regimes, there may be a a layered structure with weak lower crust layer and a strong upper mantle. Te is then the equivalent single layer. There are good gravity and topography data over the whole Cordillera that can be used for Te mapping. There are a number of methods to estimate Te, but most recent analyses use the coherence as a function of horizontal wavelength between topography and gravity (e.g., Burov and Diament, 1995).

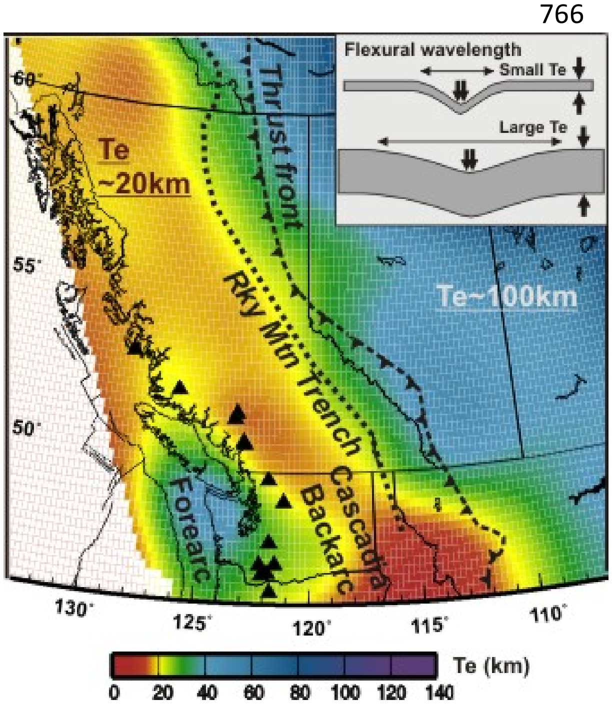

Figure 12. Effective elastic thickness Te for the northern Cordillera (after Fluck et al., 2003).

There are many uncertainties in the calculation and interpretation of Te, such as the different methods of Te calculation, the duration of the loads, and the regional stress, so caution is required for quantitative interpretation. However, the effective elastic thickness provides our best direct regional constraint on the depth to the weak lower crust. Lowry and Smith (1995), Flück et al. (2003), Hyndman et al. (2009), and Audet and Mareschal (2006) have given Te results for the Cordillera and adjacent craton. **Figure 12** shows Te for the northern Cordillera from Flück et al. (2003). Te is everywhere thin, less than 20 km, for the Cordillera backarc and over 60 km for the adjacent craton. The Cascadia cool forearc has a thick Te. In the Cordillera backarc, only the upper crust has significant strength. For a Te average in the Cordillera of 18 km, the horizontal flexural wavelength is about 100 km, so the horizontal resolution of Te estimates is similar, approximately 100 km. The horizontal resolution is larger for the adjacent craton which has a much thicker Te. Te is especially thin in western USA (Lowry and Smith, 1995), probably because of the regional extensional regime in that area since the temperature estimates are not higher than for the rest of the Cordillera. The upper crust does not act as a simple elastic plate because of the active normal faulting. For the Cordillera, the main loads may be topography generated by the spatially variable erosion, especially during the Pleistocene glaciation for the northern Cordillera. If so, the time scale for the loads is 1-2 million years, which is similar to the time scales for lower crustal flow inferred for the Basin and Range noted above. The Te results are therefore directly relevant for studying lower crustal flow in the Cordillera.

The thin Te for most of the Cordillera is consistent with the other estimators of a weak layer in the lower 10-20 km of backarc crusts. To match the thin observed Te values for the Cordillera, Hyndman et al. (2009) estimated 800-900°C at the Cordillera Moho and 400-500°C at the craton Moho, as discussed below.

7. Strength versus depth from thermal regime and laboratory data

Lithosphere strength versus depth envelopes may be estimated using the average

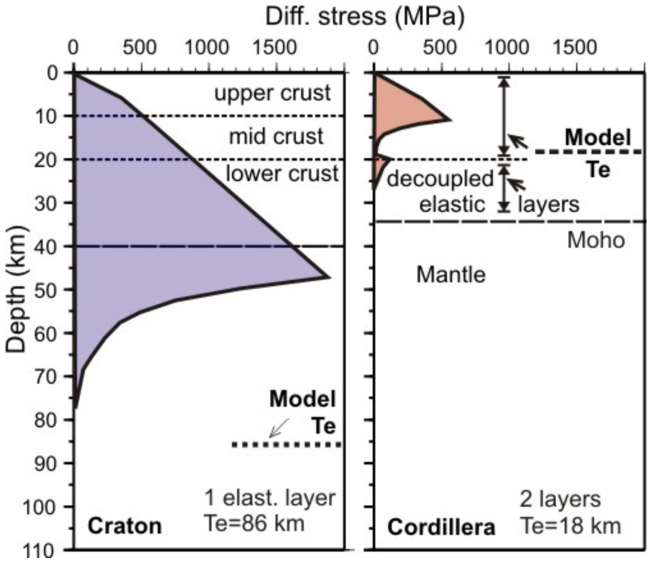

Figure 13. Strength envelopes estimated using Cordillera and craton temperatures, and Te models (Modified from Hyndman et al., 2009).

Cordillera temperature-depths and estimates of average compositions with depth (see Ranalli, 1995, for a discussion). Ranalli (2003) concluded that that the ductile strength becomes comparable to plate boundary and elevation forces for strength less than about 100 MPa (see also Bürgmann and Dresen, 2008 for field estimates of paleo stress in deep crust). This strength therefore may be taken as an approximate base of the elastic lithosphere. **Figure 13** show estimates of strength versus depth given by Hyndman et al. (2009). The crust is usually inferred to be more mafic with less SiO_2 downward, so the deep crust is stronger than the shallow crust at the same temperature. However, Wheeler (1992) suggested that polyphase mafic rocks deforming by dissolution-precipitation creep (DPC) may be as weak as quartz-rich rocks. If this is correct, the lower crust may be weaker than these estimates shown. For most reasonable models, the Cordillera backarc envelope is very weak in the lower 10-15 km of the crust, and there is little strength in the upper mantle. This is in marked contrast to that for the adjacent stable craton which has considerable strength through the crust and to about 100 km deep in the mantle. If there is sufficient quartz to control the rheology, hot thermal gradient like the Cordillera, and for a strain rate of 10^{-15} s^{-1}, i.e., for significant tectonic strain rates at small stresses, Ranalli (2003) estimated that the ductile strength becomes less than 100 MPa at 10-19 km depth depending on the rheology parameters used and whether the rocks are wet or dry.

The total strength of the lithosphere of the Cordillera is comparable to plate tectonic and high elevation gravitational potentials so it may be readily deformed by elevation and plate boundary force stress perturbations. If only the upper strong crust of the Cordillera is involved, deformation may occur at even smaller stresses. In contrast, the lithosphere is much too strong to be deformed except under exceptional rare circumstances.

A first order model effective elastic thickness Te can be obtained from the temperature data using laboratory-derived rheology. Hyndman et al. (2009) showed that the Te is closely related to temperature. Using temperatures from upper mantle seismic velocities as shown in **Figure 7**, they calculated strength versus depth for a model for the Cordillera (common Cordillera 10^{-15} s^{-1} strain rate assumed) and for the craton (10^{-19} s^{-1} strain rate assumed) (**Figure 13**). There may be thin weaker crustal layers (not shown) depending on the details of the assumed crustal compositions with increasing strength downward; in these examples any such layers do not produce decoupling so do not significantly affect the effective elastic thicknesses. The model Te are in general agreement with those measured; Cordillera about 18±5 km, craton Te ~100 km. Our study supports the conclusion that lithosphere elastic thickness and strength are controlled primarily by temperature, and that laboratory-based rheology provides a good estimate of the deformation behaviour of the crust and upper mantle.

8. Lower Crust Horizontal Seismic Reflectors, Exhumed Sheared Outcrops, and Flow Layer Thickness

8.1 Lower crust reflective bands

In the Cordillera and a number of other current or recent backarcs, there is common laminated near-horizontal reflectivity in the lower crust in deep crustal seismic reflection sections (Matthews, 1986; Allmendinger et al., 1987b; Clowes and Kanasewich, 1970; Fuchs,

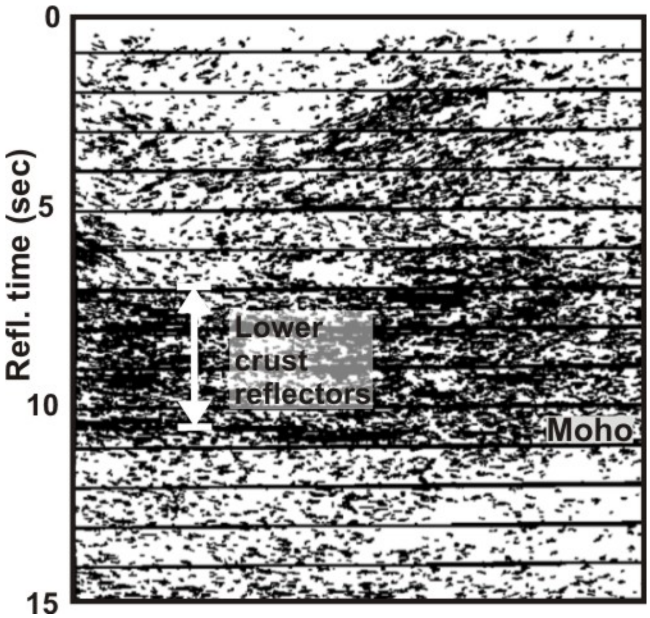

Figure 14. Reflectors in the ~10 km of the lower crust in southern British Columbia (after Cook et al., 1992).

1969; McCarthy and Thompson, 1988; Klemperer, 1989). There have been many discussions of possible causes of the reflectors (e.g., Warner, 1990; Weaver and Meissner, 1987). They include compositional layering, low-velocity fluid-filled shear zones, and horizontal magmatic intrusions. As summarized by Meissner et al., (2006), some of the clearest densely laminated reflectivity in the lower crust are in Phanerozoic extensional areas that we now interpret to be in or have recently been in backarcs. Especially strong lower-crustal reflectivity has been found in the Basin and Range province, around the British Isles, in the German Variscides, and in the Paris Basin. An example of strong lower crust reflectors in the central northern Cordillera is shown in **Figure 14** (Cook 1992). What may be an important correlation is that the temperature is usually about 450^{o}C at the top of the reflective layers, and also at the top of the Cordillera lower crust electrically conductive layer. The conductive and reflective layers are often approximately coincident (Marquis et al., 1993; Hyndman and Lewis, 1991). There is a similar correlation of the top of the reflective layers with temperature of about 350^{o}C in a detailed study for the Japan backarc (Ito, 1999). The conclusion discussed above for lower crustal flow, supports the explanation that the reflectors result from stretching and flow of heterogeneities such as mafic intrusions to form long sub-horizontal reflecting layers with marked contrasts in seismic impedance.

In a few areas, the horizontal seismic reflectivity is 5-10 km above the base of the crust. It may be that in the hot backarc, the lowermost crust at ~800°C is in granulite facies conditions so has dry mineralogy and therefore stronger than slightly shallower where there is amphibolite conditions and more bound water results in less strength even though cooler.

The conclusion of a tectonically very weak lower crust in backarcs supports the idea that lower-crust horizontal reflectivity is due to horizontal shearing. In some areas, the horizontal seismic reflectivity extends up to the mid-crust. Although the current temperatures at mid-crustal levels may be too low for ductile deformation, temperatures may have been hotter at earlier times and when that level was deeper in a thickened crust in the past, or before upper crust layers were eroded. The reflectors are interpreted to result from stretching and flow of heterogeneities such as mafic intrusions, producing long thin sub-horizontal reflecting layers with marked contrasts in seismic impedance. Such layers could be frozen in and remain as temperatures decline and the crust strengthens. Oueity and Clowes (2010) proposed such a development to explain seismic reflection and refraction characteristics at the base of the crust beneath the Paleoproterozic Great Bear magmatic arc in the Canadian Northwest Territories. However, lower-crust horizontal reflectors are less common in geologically older areas, so a variety of geological processes may reduce the layered impedance contrasts over long periods of time.

8.2 Exhumed horizontally sheared lower crust

Important support for horizontal shearing as the origin of lower crustal reflectors, is the observation of banding or lamination and anisotropy in regional amphibolite to granulite facies rocks formed in the lower crust, as inferred from petrological studies. In several large areas, rocks with strong horizontal compositional laminations are now exposed at the surface (e.g., Sandiford et al., 1989; Weiss et al., 1999; Pohl et al., 1999; Dumond et al., 2010; Klepeis and

Clarke, 2004; Regan et al., 2014). Temperatures at the time of the lower-crust ductile flow were estimated to be about 800°C at 30 km depth by Dumond et al. (2010), similar conditions to those estimated for the current Cordillera. In addition to these examples, lower crust flow that produces sub-horizontal layers also may provide an explanation for the globally widespread banded gneisses (several examples, Dietrich, 1963 and Myers, 1978). Banded gneiss layers with strong impedance contrasts in the lower crust, are one model for the observed horizontal seismic reflectivity.

8.3 Lower crust flow thickness

The thickness of the lower crust channel flow has three main constraints. (1) Although flow is easiest at the higher temperatures near the base of the crust, there is a theoretical constraint that flow in thinner layers requires much higher temperatures and lower effective viscosities than thick layers (e.g., McKenzie et al., 2000). A thickness of about 10 km appears to be needed for reasonable effective viscosities. (2) There is a minimum temperature for ductile flow of about 450^{o}C for most rocks of compositions expected in the lower crust (e.g., Ranalli, 1995). McKenzie et al. (2000) concluded temperatures above 400-500^{o}C. The temperature range from 450 to 800-850^{o}C, from the mid crust to the base of the crust, also gives a thickness of about 10 km. (3) Lower crust reflectors that may represent channel flow, commonly have a thickness of about 10 km as in the example shown in **Figure 14** and in numerous examples given in the references cited above. However, as I noted earlier, the upper reflectors could be frozen in from when that depth was at a higher temperature. In contrast to channel flow, lower crust detachment or shear can occur in a quite thin layer.

9. Lower Crust Effective Viscosity

9.1 Viscosity required for lower crust flow

A number of lower crust viscosity estimates have been obtained using the constraint that topography on the Moho is interpreted to relax over times of a few million years. Much has been written, and I give only a summary of the conclusions and a number of references. The discussions usually consider only simple linear Newtonian viscosity, since it has been concluded to be an adequate approximation for most such modeling (e.g., Kaufman and Royden, 1994). Because of this approximation, we should use the expression "effective viscosity", but for simplicity I generally will use simply viscosity. From their model analyses of the Basin and Range, Kruse et al. (1991) concluded that, over the length scale of ~500 km and extension factors of 1.4–3.0 over 10 m.y. the required maximum viscosity is less than 10^{18}-10^{20} Pa s for flow in 10-25 km thick channels. Flow over shorter length scales, <150 km and a thick layer, may occur for higher viscosities. Block and Royden (1990) calculated that 1 km difference in surface elevation gives 500 Pa/m for a 10 km thick channel and 1cm/yr needs a viscosity less than 10^{19} Pa s, comparable to the values estimated by other methods. Similar values were also estimated by Kaufman and Royden (1994) for a 10 km thick flowing layer. McKenzie et al. (2000) estimated a maximum viscosity of 10^{20} Pa s for length scales of 100-150 km. For maintaining a flat Moho over ~30 km lengths they estimated a maximum of $6x10^{19}$ Pa. In their models for lower crust channel flow, Jamieson et al. (2011) used 10^{19} Pa s at 700-750^{o}C with some weakening due to a small amounts of partial melt. However, as I now conclude, backarc temperatures at the base of the crust are commonly hotter, 800-850^{o}C for normal 32 km crust and even hotter for thick crust as for Tibet.

The conclusions from these modeling studies is that lower crust channel flow can occur with modest driving force, i.e., Moho topographic gradients of a few degrees or topographic variations of 1 km or less, for viscosities of 10^{19} Pa s. Under some conditions, such as for an especially thick layer, flow may occur for higher viscosities of 10^{20} Pa s. The maximum viscosity is strongly dependent on the thickness of the flowing layer and on the length of the flowing channel.

Relevant to lower crust detachment shearing is the development of core complexes over length scales of 10's of km. Tirel et al. (2008) estimated that viscosities of less than 10^{20} Pa s are required in the lower crust, and 10^{22} Pa s in the underlying mantle. Therefore, lower crust detachment probably can occur for viscosities of 10^{20} Pa s or lower. Longer detachments such as across the Cordillera when foreland thrusting in the eastern Rocky Mountains is driven by terrane collision or a strong subduction thrust on the western margin, may require lower viscosities.

9.2 Viscosity from response to local loading

In a few places backarc lower crust viscosities have been estimated from the response with time of crustal loading, such as from ice sheet melting or changes in large lake loads. A major source of uncertainty is the time dependence of effective viscosity and the flexural wavelength of the loading. These data give an important confirmation of low effective viscosities in the Cordillera and other continental backarcs. In the models the thickness of the lower crust weak layer is poorly constrained and there has been uncertainty as to whether the concluded low viscosities are in the lower crust or in the upper mantle. I give several backarc examples. From the vertical crustal response to loading by Lake Bonneville in the Basin and Range area of western United States, Bills et al. (1994) estimated an effective viscosity of 10^{20} Pa s at 30 km depth. The time scales involve range from 100 to 10,000 years. An estimate of 4×10^{19} Pa s was found for the lower crust by Kaufmann and Amelung (2000) from the response to reservoir-induced deformation by Lake Mead, Nevada. In the central Andes backarc Bills et al. (1994) found effective viscosities from tilting of Lake Minchin shorelines, of lower than 5×10^{20} Pa s. From modeling the response to glacial unloading of the Holocene mass fluctuation of the Patagonian icefields in the southern South American backarc, Ivins and James (1999), estimated a viscosity of 5.0×10^{18} - 5.0×10^{19} Pa s, and Dietrich et al. (2010), 4×10^{18} Pa s. In these models the thickness of the elastic layer is not well constrained and the low viscosity could be in the lower crust or shallow mantle.

9.3 Rheology from laboratory flow laws and estimated temperatures

The viscosities calculated as required for lower crust flow and the viscosities estimated from response to loading may be compared to those from laboratory data and lower crust estimated temperatures and other conditions. There has been much study and much has been written about the rheology of the crust as estimated from laboratory data. There are many complexities, including: Newtonian versus power law rheology, the effect of water, the effect of varying strain rates, the nature of the deformation mechanisms, the effect of localization of deformation, the effect of polymineralic rocks, and the relation of estimated maximum strength to effective viscosity. Good summaries and references are provided by Kohlstedt et al. (1995), Bürgmann and Dresen (2008), and Burov (2011). In one relevant estimate for the lower crust, Kaufman and Royden (1994) estimated 10^{18} Pa s for 825°C near the base of the crust, and an order of magnitude decrease in effective viscosity for each 75°C increase in temperature. This is at least as low a viscosity as estimated to be required for lower crust flow.

10. Discussion and Conclusions

Lower crust flow and detachment has been recognized for some time for the thick crust backarc areas of Tibet and the high Andes, but also for the thin crust area of the Basin and Range of western United States. I have documented the evidence that lower crust flow must have occurred in many areas of the North America Cordillera from Mexico to Alaska, and that there are surprisingly uniform high temperatures such that flow and detachment can readily occur. From four seismic structure constraints, everywhere in the Cordillera the Moho topography is remarkably flat, both on short spatial scales of a few 10's of kilometers, especially from seismic reflection, and on scales of 100's of kilometers, especially from seismic tomography. Short scale detachment and flow accommodation deformation is shown by the lack of Moho offsets associated with extensional normal faults (core complexes) and other local deformation. The long wavelength accommodation is evident by the constant crustal thickness over large areas in the Cordillera, in spite of major extension of up to a factor of two and major shortening and thickening that also results in exhumation of lower crust rocks, such as in the Laramide deformation of the eastern Cordillera. Flow and detachment appear to happen whenever normal faulting extension and thinning or thrust thickening occur, such as due to time variations in plate boundary forces. The time required for the flow to flatten the Moho appears to be less than a few 10s of m.y. This short time constant suggests that significant lateral contrasts in elevation and crustal thickness must be transient and be maintained by ongoing processes at time scales of 10s of m.y.

Cordillera areas with well-studied large extension but smooth Mohos include the currently extending Basin and Range area of USA and past extension in southwestern British Columbia. Well-studied examples of extension in other current or past backarcs, are the northern British Isles and Germany. An example of well-studied shortening is the areas of Laramide deformation in the eastern North American Cordillera. These areas are inferred to have had past high elevations with thick crust; they now have thin crust. Although erosion may contribute to thinning the crust, it is unlikely to be sufficient. The Cordillera-wide very weak lower crust also helps the understanding of the very large-scale crustal translations and bending around horizontal axes such as in oroclines. Only the upper crust is likely involved. An excellent example is the translation to the northwest and bending of a large area of the northern Cordillera crust as inferred by Johnston (2001; 2008). The translation and bending likely involved only the upper crust.

The temperature at the Moho beneath most of the Cordillera and probably most other continental backarcs is 800-850^{o}C by five constraints. A few areas are inferred to be a little hotter where the upper crust radioactive heat generation is unusually large, and where there is current or thermally recent extension, like the Basin and Range. However, differences in these areas from the Cordillera average temperatures are not clearly resolved by the other temperature constraints.

The effective elastic thickness Te is thin everywhere in Cordillera, average about 18 km. There is very little strength in the lower crust or in the upper mantle. This thickness matches that of the upper crust elastic layer from the estimated temperatures and a reasonable range of laboratory rheologies, although the applicable rheology parameters have considerable uncertainty. The thin Te values support the conclusion of the lower crust being very weak, having low effective viscosity.

Model estimates of the maximum viscosity for lower crust flow to occur are in the range 10^{18}-10^{20} Pa s for flow in a 10-25 km thick channel, with most viscosity estimates less than 10^{19} Pa s. The maximum viscosity for lower crust flow is strongly dependent on the thickness of the channel. A few estimates of lower crust viscosity from transient loading are in the range 10^{19}-

10^{20} Pa s. Lower crust horizontal detachment may occur at higher viscosities and therefore at somewhat lower temperatures.

Most of this article has focussed on the current North American Cordillera but similar high temperatures in the lower crust occur beneath most continental backarcs and lower crust detachment and channel flow is expected. I noted the high temperatures and thin crust in western Europe which is a current or recent backarc. Eastern China and Korea are other backarc regions with similar characteristics. The central portion of the South America subduction backarc in the high Andes has a thick crust, but to the north and south, much of that subduction zone has thin crust similar to North America. Of course, Tibet is a backarc that is still hot associated with the ongoing convergence of India. The few exceptions of cool backarcs are where there is flat slab subduction. There is little intervening space for vigorous small scale convection between the subducting slab and the overlying lithosphere, so crustal temperatures are usually low. These areas also usually have few if any arc volcanoes. Lower crust detachment and flow is also important in the interpretation of ancient terranes that were in former hot backarcs. This is an area for future productive study.

In summary, over 10's of m.y. geological time scales, the North America backarc Moho can be viewed as a boundary between almost 'liquid' lower crust overlying a low-viscosity upper mantle. The Moho boundary relaxes to a near-horizontal gravitational equipotential over time scales of a few 10's m.y., and the Cordillera lower crust readily accommodates horizontal detachment motion over long distances.

Acknowledgments

Many discussion with colleagues, postdoctoral fellows and graduate students are gratefully acknowledged.

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