

# Subduction zone backarcs, mobile belts, and orogenic heat

Roy D. Hyndman, Pacific Geoscience Centre, Geological Survey of Canada, P.O. Box 6000, Sidney, British Columbia V8L4B2, Canada, and School of Earth and Ocean Sciences, University of Victoria, Victoria, British Columbia V8W3P6, Canada, rhyndman@nrcan.gc.ca; Claire A. Currie, School of Earth and Ocean Sciences, University of Victoria, Victoria, British Columbia V8W3P6, Canada; and Stephane P. Mazzotti, Pacific Geoscience Centre, Geological Survey of Canada, P.O. Box 6000, Sidney, British Columbia V8L4B2, Canada

## ABSTRACT

Two important problems of continental tectonics may be resolved by recognizing that most subduction zone backarcs have hot, thin, and weak lithospheres over considerable widths. These are (1) the origin of long-lived active “mobile belts” contrasted to the stability of cratons and platforms, and (2) the origin of the heat of continental collision orogeny. At many continental margin plate boundaries, there are broad belts with a long history of distributed deformation. These regions are mobile because the lithosphere is sufficiently weak to be deformed by the forces developed at plate boundaries. We conclude that mobile belts are weak because they are hot, and they are hot as a consequence of their position in present or recent backarcs. Most continental backarcs have thin and hot lithospheres, not just those with active extension or rifting. Moho temperatures are 800–900 °C and lithosphere thicknesses are 50–60 km, compared to 400–500 °C and 150–300 km for cratons and other regions with a thermotectonic age greater than ca. 300 Ma. The temperature differences result in backarc lithospheres that are weaker than cratons by more than a factor of 10. Backarcs may be hot because shallow asthenosphere convection results from the reduction in viscosity due to water rising from the underlying subducting plate. Hot, weak, former backarcs are expected to be the locus of most deformation during continent or terrane collision orogeny. The heat indicated by orogenic plutonism, high grade metamorphism, and ductile deformation may come from the preexisting

hot backarc lithosphere, not from the orogenic deformation process itself.

## INTRODUCTION

The model of plate tectonics with narrow plate boundaries provides an excellent first-order description of global tectonics. Plate tectonics also provides an elegant explanation for orogenic crustal shortening and thickening in terms of continental or terrane collision. However, a number of large-scale tectonic problems are not explained by the simple rigid plate and continental collision models. In this article, we present explanations for two such tectonic problems: the origin of long-lived active “mobile belts” that lie along a number of continental margin plate boundaries, contrasted to the stability of cratons and platforms, and the origin of the heat of continental collision orogeny. Current mobile belts are up to 1000 km wide and cover nearly a quarter of the continents (e.g., Stein and Freymueller, 2002; Thatcher, 2003). For example, the North American Cordillera has been tectonically active for >180 m.y., with evidence for older tectonic events extending this history to >350 m.y. (e.g., Burchfiel et al., 1992) (Fig. 1). Rates of deformation in current mobile belts are commonly 5–15 mm/yr; i.e., 10%–20% of the main plate boundary rates as indicated by distributed seismicity, precision global positioning system (GPS) measurements, and geological studies.

The long geological histories of deformation suggest that these mobile belts exhibit long-term lithosphere weakness compared to cratons. In contrast, the Precambrian cratons and stable platforms have exhibited little internal



Figure 1. The North America Cordillera mobile belt. The high elevation and complex current tectonics illustrate the hot, weak, backarc lithosphere deformed by variable margin forces.

deformation over long geological periods. They have some characteristic that allows them to maintain an especially thick, strong lithosphere, such as a more refractory mantle composition (Jordan, 1978; Forte and Perry, 2000). The reason for the long histories of tectonic activity in the mobile belts has not been clear. Also, most continental mobile belts have high elevation; they are mountain belts. Surprisingly, however, there is commonly no crustal thickening. In this article, we emphasize the important distinction between long-lived mobile mountain belts where the elevation is mainly thermally supported and short-lived continent or terrane collision orogenies where there is major crustal shortening and thickening.

Plate tectonics provides an elegant explanation for orogenic crustal shortening and thickening in terms of continental or terrane collision when an ocean closes. However, an important question remains unresolved that we attempt to address: What is the origin of the heat of orogeny? Significant heat from frictional deformation processes has been discounted through studies by a number of authors, and most other orogenic processes should absorb rather than generate heat.

In this study, we examine the hypotheses that (1) most mobile belts are located in backarcs or recent backarcs that are characterized by hot, thin, and weak lithospheres; and (2) most continent or terrane collision orogenic belts involve shortening of former hot backarc mobile belts. We do not discuss less common weakening mechanisms that subsequently may be reactivated in collision, such as localized extension and hotspots.

### BACKARC THERMAL STRUCTURE

Why are mobile belt regions so weak and why are cratons so strong for long geological periods? The primary reason appears to be that the mobile belts are hot, whereas cratons are cold (e.g., Vitorello and Pollack, 1980; Chapman and Furlong, 1992), and, below the upper crust brittle zone, there is a strong decrease in strength with increasing temperature. The systematic decrease in heat flow with age, or time since the last “thermotectonic event,” has been well recognized. The greatest difference is between the currently active mobile belts (70–90 mW/m<sup>2</sup>) compared to the cratons (~40 mW/m<sup>2</sup>) (e.g., Chapman and Furlong, 1992; Vitorello and Pollack, 1980). Temperatures at the base of the crust are ~400 °C lower for cold cratons compared to young, hot mobile belts (i.e., 400–500 °C vs. 800–900 °C, respectively). Although only the coolest case of cratons is discussed here, Paleozoic stable areas are only slightly warmer than Archean cratons and are estimated to be almost as strong. As an example, Figure 2 shows temperature estimates for the Canadian Cordillera mobile belt compared to the adjacent Canadian Shield (Hyndman and Lewis, 1999; Lewis et al., 2003). With these thermal estimates, the base of the mobile belt lithosphere (hot enough for vigorous convection, close to the solidus) is at a depth of only 50–60 km. The mobile belt high temperatures and thin lithospheres also are inferred from many other deep temperature indicators, including temperature-dependent uppermost mantle seismic properties (Moho refraction velocity, P<sub>n</sub>, tomography compression and shear wave velocities, V<sub>p</sub> and V<sub>s</sub>, and attenuation,

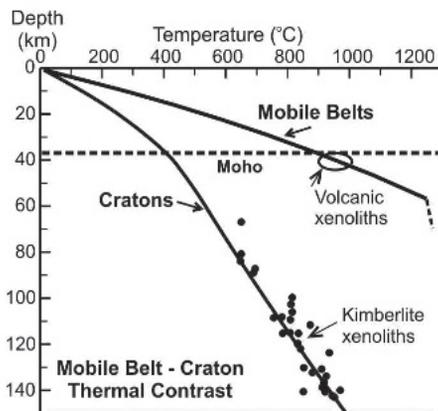


Figure 2. Temperature vs. depth for the Cascadia backarc and adjacent craton (from Hyndman and Lewis, 1999).

Q), effective elastic thickness, T<sub>e</sub>, thermally supported high elevations, seismic estimates of lithosphere thickness, and upper mantle xenoliths. The high temperatures of current mobile belts are also indicated by widespread sporadic Cenozoic basaltic volcanism, in contrast to the almost complete absence of recent volcanism in cratons and stable platforms. The same indicators of temperature show the lithospheres of cratons to be cold and >200 km thick.

We have carried out a global survey of continental backarcs, and, in all cases where there is sufficient data, we have found surprisingly high and uniform temperatures across wide zones, indicated by heat flow and other deep temperature constraints (also see Currie,

2004). It is well recognized that extensional backarcs are hot (e.g., Wiens and Smith, 2003), including the Basin and Range and extensional west Pacific backarc basins. In these regions, rifting and spreading are obvious sources of heat. We therefore focus on nine backarcs where we conclude that there is no current or recent thermally significant extension in the region, as indicated by GPS, seismicity, or geological studies (Fig. 3). Our areas are all more than 100 km from such extension; Morgan (1983) showed that the thermal effects of rifts extend only a short distance. The thermal effect of extension decays with a time constant of ~50 m.y., so we have also excluded areas with significant mid-late Cenozoic extension.

We summarize the two principal indicators of high temperatures in Figure 4 (for details and references, see the GSA Data Repository<sup>1</sup>): (1) surface heat flow greater than ~70 mW/m<sup>2</sup> (for normal upper crust heat generation of 1.0–1.5 μW/m<sup>3</sup>) (e.g., Lewis et al., 2003; Chapman and Furlong, 1992); and (2) low seismic velocities in the upper mantle (i.e., P<sub>n</sub> velocities <7.9 km/s compared to ~8.2 km/s for cratons) (e.g., Black and Braile, 1982; Lewis et al., 2003) or seismic tomography velocities lower than the global average by at least 2% for P-waves or 4% for S-waves (~4% and ~7% slower relative to cratons) (e.g., Goes et al., 2000; Goes and van

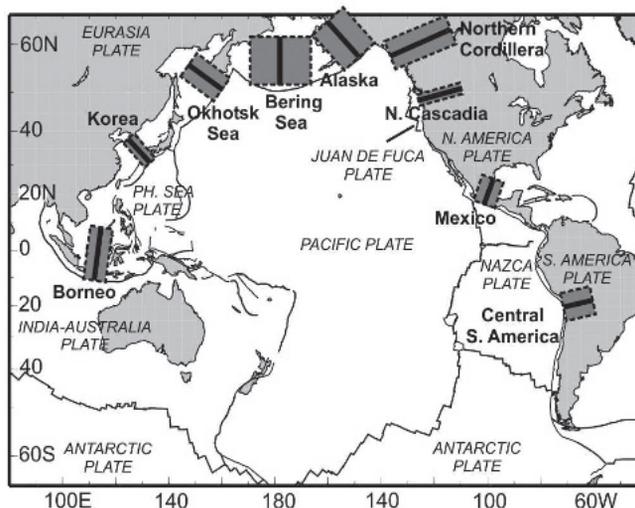


Figure 3. Examples of nonextensional backarcs with high heat flow and other high temperature indicators.

<sup>1</sup>GSA Data Repository item 2005030, Table DR1, notes, and references for Figure 4, is available on request from Documents Secretary, GSA, P.O. Box 9140, Boulder, CO 80301-9140, USA, editing@geosociety.org, or at www.geosociety.org/pubs/ft2005.htm.

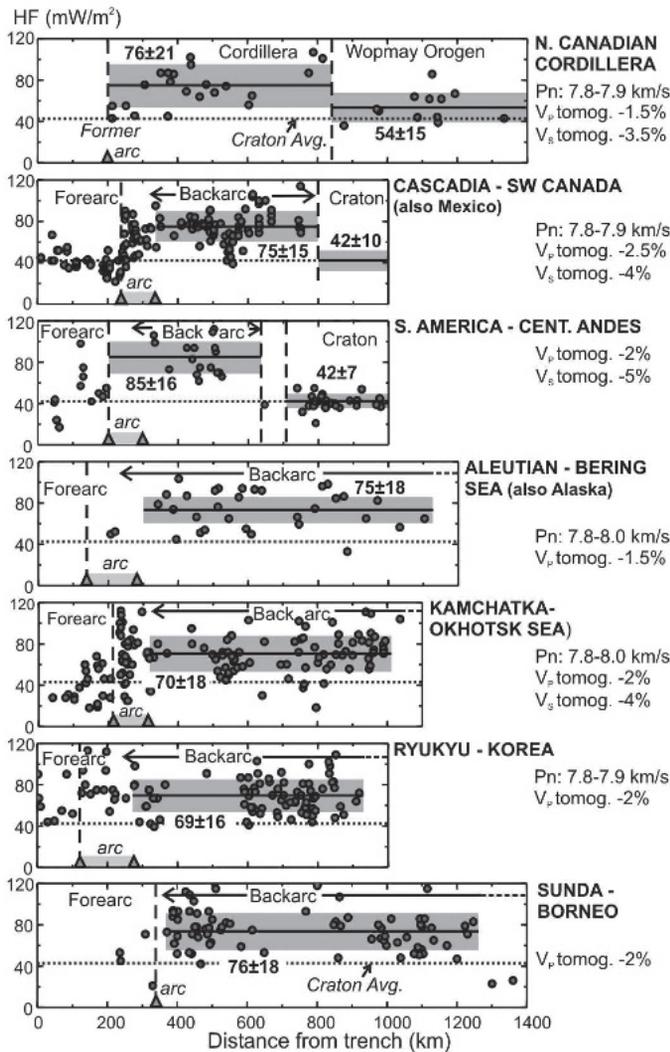


Figure 4. Heat flow (HF) data and other indicators of high temperatures across backarcs. References are given in the GSA Data Repository (see footnote 1). Pn—refraction uppermost mantle velocity;  $V_p$ —tomography compressional velocity;  $V_s$ —shear velocity.

der Lee, 2002). Temperature appears to be the dominant control of mantle velocity; the effects of composition, partial melt, anisotropy, etc., usually are much smaller (e.g., Goes et al., 2000; Wiens and Smith, 2003).

We include one oceanic backarc (Bering Sea) and one former continental backarc (the northern Canadian Cordillera that was cut off by transform faulting in the Eocene) that give very similar thermal results to the current continental backarcs. The heat flow averages range from 70 to 85 mW/m<sup>2</sup>; Pn velocities are low, 7.8–7.9 km/s, and seismic tomography velocities are low, consistent with high lithosphere temperatures (i.e., ~800 °C at the Moho). There are also only small variations in heat flow and in the other indicators of deep temperatures across the backarcs. All of the other backarcs we have examined show evidence that they are hot, although they provide less information. It is well recognized that an active volcanic arc requires high temperatures (~1200 °C) in the underlying mantle wedge at a depth of ~100 km; however, we find that the high lithosphere temperatures extend across the whole backarc, even where there is no extension.

## ELEVATION OF MOBILE BELTS

An important attribute of most continental mobile belts that indicates they are hot is high elevation. They are mountain belts. This elevation is usually in spite of relatively thin crust. For example, very little of the North America Cordillera has a crustal thickness greater than ~35 km (Mooney et al. 1998; Perry et al., 2002), but elevations are commonly ~1500 m. In contrast, cratons have elevations usually near sea level and average crustal thicknesses of ~40 km. Only in a few mountain belt regions of unusually high elevation and tectonic crustal thickening, such as Tibet, the central part of the Andes, and part of the Alps, is there thick crust. We argue that high elevations therefore provide a simple first order mapping association for hot, thin lithosphere.

Mobile belts are inferred to be in isostatic balance, in spite of the combination of high elevation and thin crust, because there is thermal expansion and density reduction to a depth of several hundred km relative to stable cratonic and platform areas. The effect of crust and upper mantle temperature on elevation has been discussed in detail by Lachenbruch and Morgan (1990). In a simple approximation, the density change due to a 200 °C average temperature difference to a depth of 150 km and a coefficient of thermal expansion of  $3.5 \times 10^{-5}$  per °C will approximately balance an elevation difference of 1 km or a crustal thickness difference of ~7 km (e.g., Lachenbruch and Morgan, 1990; Hyndman and Lewis, 1999).

## CONVECTION MODEL FOR BACKARCS AND THE ROLE OF WATER

An explanation for why most mobile belts are hot is that they are all in backarcs or recent backarcs, and all backarcs are hot, as discussed above. The high temperatures are likely due to rapid upward convective heat transfer beneath thin lithospheres (e.g., Hasebe et al., 1970). We show a schematic convection model in Figure 5, along with the heat flow across the southern Canadian Cordillera from the trench to the craton. Most models of backarc convection have assumed that the circulation is driven by the traction and thermal effects of the downgoing oceanic plate. However, it has proved difficult

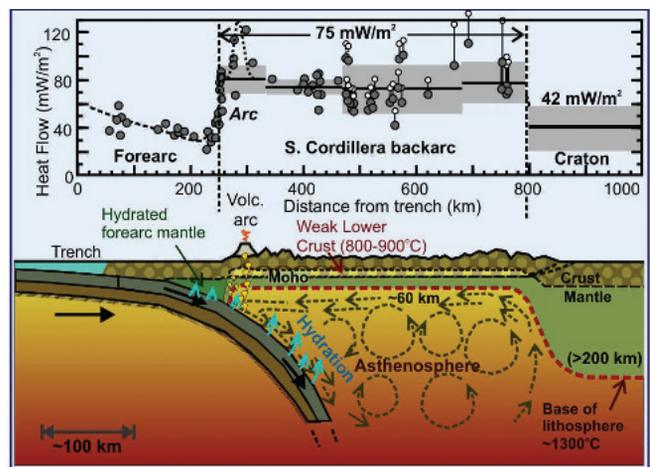


Figure 5. Heat flow across the N. Cascadia backarc and thermal-tectonic model. Solid points are corrected, open points uncorrected, for crustal heat generation, following Hyndman and Lewis (1999).

to produce uniform high heat flow across the backarc with such models, and regional small-scale convection seems to be required (Currie et al., 2004; see also Nyblade and Pollack, 1993). To maintain the high backarc heat flow, convection must be vigorous, with flow rates faster than relative plate rates (e.g., Currie et al., 2004).

In a few continental areas such as the Basin and Range, present or recent crustal extension has an additional thermal effect (e.g., Lachenbruch and Sass, 1977). Similarly, in the areas of 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 asymptote for long times after the basins opened.

An explanation for shallow vigorous convection beneath backarc lithospheres is that, in addition to high temperatures, the mantle viscosity is lowered by incorporation of water expelled from the underlying subducting oceanic plate (e.g., Dixon et al., 2004; Honda and Saito, 2003). The amount of water supplied is estimated to be very large (e.g., Peacock, 1993). The backarc convection system is poorly understood, but vigorous convection may mix the water throughout the whole wedge. Mantle rocks containing small amounts of water in the mineral structure (>50 ppm) have a substantially 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 (former backarc) of the western United States and concluded that such low viscosities require significant water in the upper mantle, as well as high temperatures that are close to the solidus. They also argue that the observed very low shear-wave velocities in the upper mantle from seismic tomography require water as well as high temperatures.

In areas where the landward boundary of the backarc is a craton or old platform, the asthenosphere convection may be limited by thick, refractory craton lithosphere. However, the original craton 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 thermal convection to continue beneath the thinned region to the edge of the unextended craton lithosphere. The backarc also may be widened by the addition of accreted terranes, such as in western North America.

### LITHOSPHERE STRENGTH OF BACKARC MOBILE BELTS

Lithosphere strength is expected to be the primary control on the nature of deformation processes. Lithosphere strength increases with depth in the shallow frictional regime, but at greater depths where there is ductile deformation, temperature and composition (especially crust vs. mantle) are the most important controls (e.g., Ranalli, 1995), so strength decreases with depth. A representative example of the strength of the lithosphere as a function of depth in a high temperature mobile belt and low temperature craton is shown in Figure 6 (top) for the Canadian Cordillera and adjacent Canadian Shield (see Flück et al., 2003; Lewis et al., 2003). In this simple model, the upper crust follows the Byerlee's Law

criteria for brittle failure (e.g., Ranalli, 1995) with hydrostatic pore pressure (e.g., Townend and Zoback, 2000), and deformation at greater depths follows power law creep (Karato and Wu, 1993) at a common plate boundary deformation rate of  $\sim 10^{-15} \text{ s}^{-1}$ . The upper crust has the rheology of granite, the lower crust of diabase, and the mantle of olivine (e.g., Ranalli, 1995). The much thinner and weaker lithosphere predicted for backarcs compared to cratons is clear.

The total strength of the lithosphere may be estimated by integrating the strength over depth (Fig. 6, bottom). Wet and dry refer to the laboratory rheologies that may apply to backarc mobile belts and cratons, respectively. The combined crustal and mantle lithosphere strength for backarc mobile belts is at least a factor of 10 less than for cold cratons. The mobile belt lithosphere strengths of  $< 5 \times 10^{12} \text{ N/m}$  are within the range suggested for plate tectonic forces and the gravitational potential of mountain belts of  $1\text{--}10 \times 10^{12} \text{ N/m}$  (Lynch and Morgan, 1987; Whittaker et al., 1992; Zoback et al., 2002). In contrast, craton lithospheres are too strong to be readily deformed. Alternatively, if one fixes the plate boundary forces at the estimated values, the calculated strain rates for backarc mobile belts are in the range observed for plate boundaries and are very small for stable cratons and platforms (Zoback et al., 2002). This large difference in lithosphere strength between hot backarcs and cold cratons

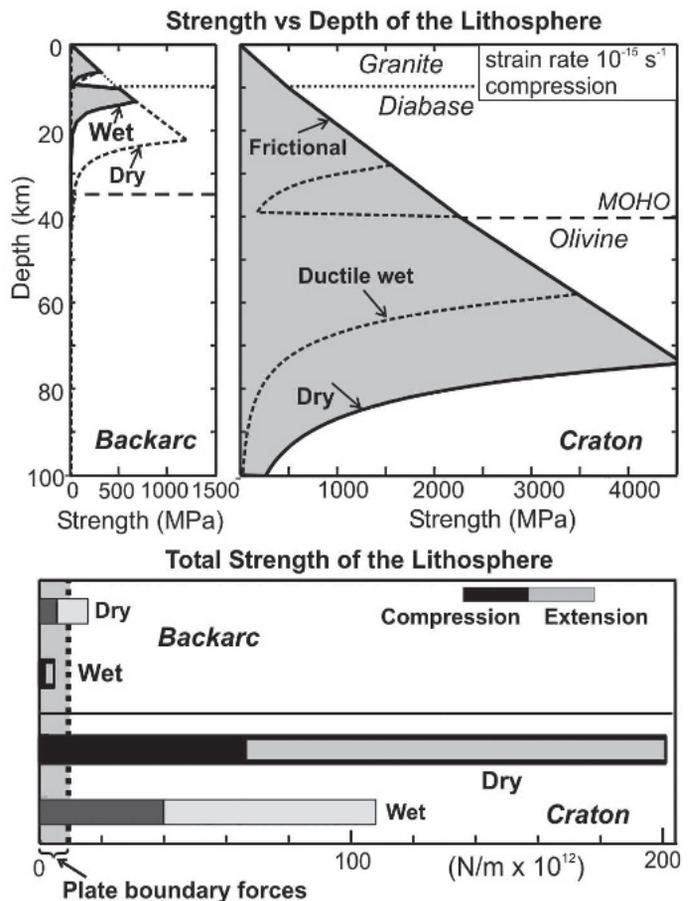


Figure 6. Top: strength vs. depth. Bottom: total lithosphere strength for hot backarc mobile belts and cold cratons.

should play a major role in subsequent orogenic deformation. Also, most backarc mobile belts are hot enough for the lower crust to be very weak and to allow lower crust detachment (Fig. 6, top).

### DURATION OF HIGH TEMPERATURES IN MOBILE BELTS

There must be a finite life to the high temperatures in backarc mobile belts after the source of heat is removed, since most past mobile belts active in the Paleozoic or earlier no longer exhibit the characteristic high lithosphere temperatures. In the backarc convection model, the vigorous free convection should decline rapidly following the termination of subduction and the loss of water as a flux for convection. Although the processes involved are undoubtedly complex, lithosphere cooling and thickening are probably conductive following the increase in upper mantle 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 heat flow (and inferred lithosphere temperatures) versus age of the last thermotectonic event (e.g., Chapman and Furlong, 1992; Pollack et al., 1993; Chapman and Pollack, 1975; Pavlenkova, 1996; Artemieva and Mooney, 2001) (Fig. 7). The thermotectonic age is assumed to correspond approximately to the time since termination of subduction and therefore of water input. Most of the rapid decrease in heat flow appears to be in the several hundred m.y. following the last thermotectonic event. A similar cooling and lithosphere thickening time is suggested by several examples. The Appalachian mobile belt, with its last significant deformation ca. 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), the difference may be primarily due to greater upper crust heat generation (Mareschal and Jaupart, 2004). In contrast, the northern Canadian Cordillera, where margin subduction was cut off by the Queen Charlotte transform fault in the Eocene, has high heat flow and inferred deep temperatures (Lewis et al., 2003), similar to currently active backarcs (Fig. 4).

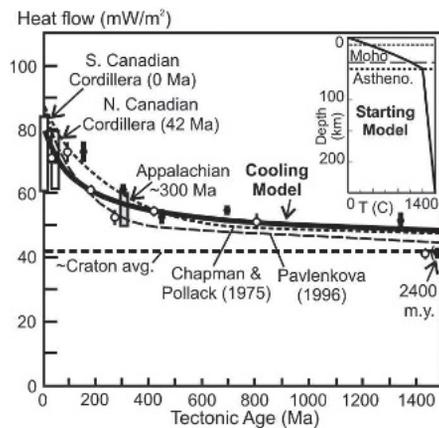


Figure 7. Heat flow vs. thermotectonic age. The zero time is interpreted to mark the termination of backarc convection (data references in text).

To illustrate the cooling, we have used a simple conductive model with an initial ~50-km-thick backarc lithosphere and an underlying adiabatic asthenosphere (Currie, 2004) (Fig. 7). This 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. Variations with depth of thermal properties and heat generation are included (e.g., Lewis et al., 2003). The addition of the effect of erosion of the high-radioactivity upper crust would give a somewhat lower heat flow at long times (Fig. 7). We also recognize that the thermotectonic age used in the heat flow data plot (peak regional metamorphism, major orogenic plutonism, etc.) is not precisely the same as the age of termination of subduction. However, the heat flow as a function of age predicted by the model is in good agreement with that observed (Fig. 7), which suggests that the termi-

nation of free convection occurs less than a few tens of millions of years after subduction stops.

### CONTINENTAL COLLISION AND OROGENIC HEAT

Recognizing that most continental or terrane collisions inherit preexisting hot, weak backarcs may provide a resolution to the long-standing problem of the origin of the heat of orogeny (e.g., Loosveld and Etheridge, 1990; Jamieson et al., 1998; Thompson et al., 2001; Vanderhaeghe and Teyssier, 2001; Collins, 2002). High temperatures are a defining feature of orogenic belts, as inferred from widespread granitic plutonism, high temperature regional metamorphism, and ductile deformation at mid-crustal depths. Often, the heat and high temperatures are simply ascribed to “orogenic heating” without reference to the mechanism. In fact, most orogenic processes should absorb heat.

We discuss only continent or terrane collision orogeny where the backarc is continental. Some backarcs are primarily oceanic and may be largely consumed by subduction during closure. Two main sources of heat have been previously proposed. The first ascribes high temperatures to backarc extension prior to collision. This explanation has the problem that the extension must occur less than 50 m.y. before the collision orogeny for the heat not to have decayed. Also, Thompson et al. (2001) argue that there must be thinning of the mantle lithosphere with little thinning of the crust to explain the inferred crustal thicknesses following collision. The second proposed heat source is high radioactive heat generation rocks emplaced in the lower crust during the shortening.

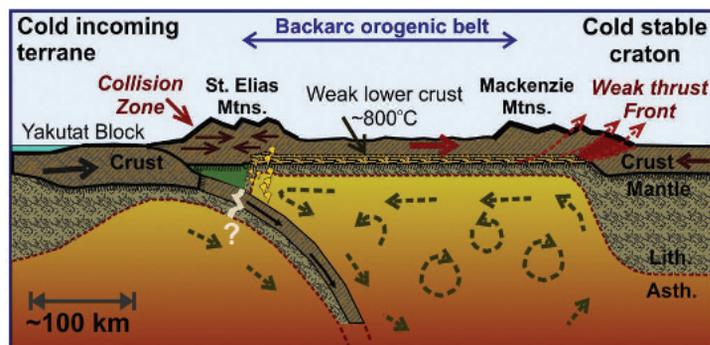


Figure 8. Model for initial collision of a weak backarc mobile belt—the northern Canadian Cordillera example.

This explanation has the problem that it results in maximum temperatures 50–100 m.y. after collision, whereas peak metamorphic assemblages are commonly argued to be synchronous with thickening (e.g., Thompson et al., 2001; Collins, 2002). This delayed heating also does not explain how shortening and thickening can be initiated in cold, strong lithosphere prior to the heating. We suggest that hot and thin former backarc lithosphere is commonly the primary locus of shortening (i.e., the vise or inherited weakness model; Ellis et al., 1998; Thompson et al., 2001; Collins, 2002). The orogenic heat comes from the preexisting hot backarc, not from the orogenic process itself.

The high temperature backarc thermal regime and inferred lithosphere strength model also provide insight into the initial development of backarc shortening in continental collision. Initial shortening and crustal thickening appear to be concentrated at the leading and trailing edges of the mobile belt. In the example of the collision of the Yakutat terrane in the Gulf of Alaska ca. 5 Ma, there has been strong shortening in the coastal collision zone and again in the Mackenzie Mountains of the eastern Cordillera. There has been little shortening in the intervening central Cordillera, and the crust remains thin (32–35 km thick) (e.g., Mazzotti and Hyndman, 2002) (Fig. 8). The backarc lower crust detachment must be sufficiently weak to transfer the collision stress to the weak foreland fold and thrust belt, even though the Cordilleran lithosphere is very thin. In contractional orogens, it has been argued that a through-going basal detachment in the lower crust separates the entire crustal section from the underlying lithosphere (Oldow et al., 1990). If this is correct for backarc mobile belts, thrust faults, transcurrent faults, and extension normal faulting are restricted to the overlying detached crustal section. Lower crust detachments seem necessary to explain the common characteristics of foreland fold and thrust belts where the upper mobile belt crust is thrust over the stable craton or platform (e.g., the northern Canadian Cordillera; Mazzotti and Hyndman, 2002) and the Andes of central South America (Hindle et al., 2002). With further shortening, the high temperatures

allow the crust of the whole backarc to be thickened to >50 km, such as for the Altiplano of the South America Cordillera, for Tibet, and perhaps for the Laramide orogeny of western North America. This thickening may be associated with lower crust ductile thickening and extrusion (e.g., Bird, 1991; Meissner and Mooney, 1998).

## CONCLUSIONS

The main conclusions of this work are (1) Continental mobile mountain belts exhibit long-term weakness compared to cratons and stable platforms. Mobile belts deform readily because they are hot and weak enough to be deformed by forces of the magnitude transmitted from plate boundaries. (2) Most mobile belts are hot and weak because they are, or recently were, located in backarcs. They may be hot because of shallow asthenosphere convection, facilitated by water from the underlying subducting slab. (3) Continental collision orogenic shortening is expected to be concentrated in former backarc mobile belts that are much weaker than the adjacent stable areas. The preexisting or inherited high temperatures provide an explanation for widespread orogenic plutonism, high temperature regional metamorphism, and ductile crustal deformation. Backarc and subsequent orogenic belt temperatures may be high enough for lower crust detachment that decouples complex surface tectonics from the uppermost mantle and for lower crustal flow.

## ACKNOWLEDGMENTS

We thank Trevor Lewis and Steven Johnston for helpful discussions. Reviewers Leigh Royden, Brendan Murphy, Sean Willett, Warren Hamilton, and Kevin Burke, although they did not always agree with us, provided exceptionally thorough and helpful reviews. We also thank *GSA Today* science editors Gerry Ross and Keith Howard for careful and thorough treatment of the manuscript. This is Geological Survey of Canada Publication number 2003201.

## REFERENCES CITED

Artemieva, I.M., and Mooney, W.D., 2001, Thermal thickness and evolution of Precambrian lithosphere: A global study: *Journal of Geophysical Research*, v. 106, p. 16,387–16,414, doi: 10.1029/2000JB900439.

Bird, P., 1991, Lateral extrusion of lower crust from under high topography, in the isostatic limit: *Journal of Geophysical Research*, v. 6, p. 10,275–10,286.

Black, P.R., and Braille, L.W., 1982, Pn velocity and cooling of the continental lithosphere: *Journal of Geophysical Research*, v. 87, p. 10,557–10,568.

Burchfiel, B.C., Cowan, D.S., and Davis, G.A., 1992, Tectonic overview of the Cordilleran orogen in the western United States, in Burchfiel, B.C., Lipman, P.W., and Zoback, M.L., eds., *The Cordilleran Orogen: Contemporaneous U.S.*: Boulder, Colorado, Geological Society of America, *The Geology of North America*, v. G-3, p. 407–479.

Chapman, D.S., and Furlong, K.P., 1992, Thermal state of the continental crust, in Fountain, D.M., Arculus, R.J., and Kay, R.W., eds., *Continental lower crust: Developments in Geotectonics*: Elsevier, Amsterdam, v. 23, p. 179–199.

Chapman, D.S., and Pollack, H.N., 1975, Global heat flow: a new look: *Earth and Planetary Science Letters*, v. 28, p. 23–32, doi: 10.1016/0012-821X(75)90069-2.

Collins, W.J., 2002, Hot orogens, tectonic switching, and creation of continental crust: *Geology*, v. 30, p. 535–538, doi: 10.1130/0091-7613(2002)030<0535:HOTSAC>2.0.CO;2.

Currie, C.A., 2004, *The thermal structure of subduction zones and backarcs* [Ph.D. thesis]: Victoria, British Columbia, Canada, University of Victoria, 267 p.

Currie, C.A., Wang, K., Hyndman, R.D., and He, J., 2004, The thermal effects of slab-driven mantle flow above a subducting plate: The Cascadia subduction zone and backarc: *Earth and Planetary Science Letters*, v. 223, p. 35–48, doi: 10.1016/j.epsl.2004.04.020.

Dixon, J.E., Dixon, T.H., Bell, D.R., and Malservisi, R., 2004, Lateral variation in upper mantle viscosity: Role of water: *Earth and Planetary Science Letters*, v. 222, p. 451–467, doi: 10.1016/j.epsl.2004.03.022.

Ellis, S., Beaumont, C., Jamieson, R., and Quinlan, G., 1998, Continental collision including a weak zone—the vise model and its application to the Newfoundland Appalachians: *Canadian Journal of Earth Sciences*, v. 35, p. 1323–1346, doi: 10.1139/cjes-35-11-1323.

Flück, P., Hyndman, R.D., and Lowe, C., 2003, Effective elastic thickness  $T_e$  of the lithosphere in Western Canada: *Journal of Geophysical Research*, v. 108, no. B9, 2430, doi:10.1029/2002JB002201.

Forte, A.M., and Perry, H.K., 2000, Geodynamic evidence for a chemically depleted continental tectosphere: *Science*, v. 290, p. 1940–1944, doi: 10.1126/science.290.5498.1940.

Goes, S., and van der Lee, S., 2002, Thermal structure of the North American uppermost mantle inferred from seismic tomography: *Journal of Geophysical Research*, v. 107, no. B3, 2050, doi: 10.1029/2000JB000049.

Goes, S., Govers, R., and Vacher, P., 2000, Shallow temperatures under Europe from P and S wave tomography: *Journal of Geophysical Research*, v. 105, p. 11,153–11,169, doi: 10.1029/1999JB900300.

Hasebe, K., Fujii, N., and Uyeda, S., 1970, Thermal processes under island arcs: *Tectonophysics*, v. 10, p. 335–355, doi: 10.1016/0040-1951(70)90114-9.

Hindle, D., Kley, J., Klosko, E., Stein, S., Dixon, T., and Norabuena, E., 2002, Consistency of geologic and geodetic displacements during Andean orogenesis: *Geophysical Research Letters*, v. 29, no. 8, 1188, doi: 10.1029/2001GL013757.

Honda, S., and Saito, M., 2003, Small-scale convection under the back-arc occurring in the low viscosity wedge: *Earth and Planetary Science Letters*, v. 216, p. 703–715, doi: 10.1016/S0012-821X(03)00537-5.

Hyndman, R.D., and Lewis, T.S., 1999, Geophysical consequences of the Cordillera-Craton thermal transition in S.W. Canada: *Tectonophysics*, v. 306, p. 397–422, doi: 10.1016/S0040-1951(99)00068-2.

Jamieson, R.A., Beaumont, C., Fulsack, P., and Lee, B., 1998, Barrovian regional metamorphism: Where's the heat?, in Treloar, P.J., and O'Brien, P.J., eds., *What drives metamorphism and metamorphic reactions?*: London, Geological Society Special Publication 138, p. 23–51.

Jordan, T.H., 1978, Composition and development of the continental tectosphere: *Nature*, v. 274, p. 544–548.

Karato, S., and Wu, P., 1993, Rheology of the upper mantle: a synthesis: *Science*, v. 260, p. 771–778.

Lachenbruch, A.H., and Morgan, P., 1990, Continental extension, magmatism, and elevation; formal relations

and rules of thumb: *Tectonophysics*, v. 174, p. 39–62, doi: 10.1016/0040-1951(90)90383-J.

Lachenbruch, A.H., and Sass, J.H., 1977, Heat flow in the United States and the thermal regime of the crust, in Heacock, J.G., Keller, G.V., Oliver, J.E., and Simmons, G., eds., *The Earth's crust; its nature and physical properties*: American Geophysical Union, Geophysical Monograph 20, p. 626–675.

Lewis, T.J., Hyndman, R.D., and Flück, P., 2003, Heat flow, heat generation, and crustal temperatures in the northern Canadian Cordillera: Thermal controls of tectonics: *Journal of Geophysical Research*, v. 108, no. B6, 2316, doi: 10.1029/2002JB002090.

Loosveld, R.J.H., and Etheridge, M.A., 1990, A model for low-pressure facies metamorphism during crustal thickening: *Journal of Metamorphic Geology*, v. 8, p. 257–267.

Lynch, H.D., and Morgan, P., 1987, The tensile strength of the lithosphere and the localization of extension, in Coward, M.P., et al., eds., *Continental Extensional Tectonics*: London, Geological Society Special Publication 28, p. 53–66.

Mareschal, J.C., and Jaupart, C., 2004, Variations of surface heat flow and lithospheric thermal structure beneath the North American craton: *Earth and Planetary Science Letters*, v. 223, p. 65–77, doi: 10.1016/j.epsl.2004.04.002.

Mazzotti, S., and Hyndman, R.D., 2002, Yakutat collision and strain transfer across the northern Canadian Cordillera: *Geology*, v. 30, p. 495–498.

Meissner, R.O., and Mooney, W.D., 1998, Weakness of the lower continental crust; a condition for delamination, uplift, and escape: *Tectonophysics*, v. 296, p. 47–60, doi: 10.1016/S0040-1951(98)00136-X.

Mooney, W.D., Laske, G., and Masters, T.G., 1998, CRUST 5.1: a global crustal model at 5 degrees x 5 degrees: *Journal of Geophysical Research*, v. 103, p. 727–747, doi: 10.1029/97JB02122.

Morgan, P., 1983, Constraints on rift thermal processes from heat flow and uplift: *Tectonophysics*, v. 94, p. 277–298, doi: 10.1016/0040-1951(83)90021-5.

Nyblade, A.A., and Pollack, H.N., 1993, A comparative study of parameterized and full thermal-convection models in the interpretation of heat flow from cratons and mobile belts: *Geophysical Journal International*, v. 113, p. 747–751.

Oldow, J.S., Bally, A.W., and Avé Lallemant, H.G., 1990, Transpression, orogenic float, and lithospheric balance: *Geology*, v. 18, p. 991–994, doi: 10.1130/0091-7613(1990)0182.3.CO;2.

Pavlenkova, N.I., 1996, Crust and upper mantle structure in northern Eurasia from seismic data: *Advances in Geophysics*, v. 37, p. 1–133.

Peacock, S.M., 1993, Large-scale hydration of the lithosphere above subducting slabs: *Chemical Geology*, v. 108, p. 49–59, doi: 10.1016/0009-2541(93)90317-C.

Perry, H.K.C., Eaton, D.W., and Forte, A., 2002, LITH5.0: a revised crustal model for Canada based on Lithoprobe results: *Geophysical Journal International*, v. 150, p. 285–294, doi: 10.1046/j.1365-246X.2002.01712.x.

Pollack, H.N., Hurter, S.J., and Johnson, J.R., 1993, Heat flow from the Earth's interior: Analysis of the global data set: *Reviews of Geophysics*, v. 31, p. 267–280, doi: 10.1029/93RG01249.

Ranalli, G., 1995, *Rheology of the Earth*: London, Chapman and Hall, 413 p.

Stein, S., and Freymueller, J., editors, 2002, *Plate Boundary Zones*: American Geophysical Union Monograph 30, 425 p.

Thatcher, W., 2003, GPS constraints on the kinematics of continental deformation: *International Geological Review*, v. 45, p. 191–212.

Thompson, A.B., Schulmann, K., Jezek, J., and Tolar, V., 2001, Thermally softened continental extensional zones (arcs and rifts) as precursors to thickened orogenic belts:

*Tectonophysics*, v. 332, p. 115–141, doi: 10.1016/S0040-1951(00)00252-3.

Townend, J., and Zoback, M.D., 2000, How faulting keeps the crust strong: *Geology*, v. 28, p. 399–402, doi: 10.1130/0091-7613(2000)28<399:HFKTCS>2.0.CO;2.

Vanderhaeghe, O., and Teyssier, C., 2001, Partial melting of crust and flow of orogens: *Tectonophysics*, v. 342, p. 451–472.

Vitarello, I., and Pollack, H.N., 1980, On the variation of continental heat flow with age and the thermal evolution of continents: *Journal of Geophysical Research*, v. 85, p. 983–995.

Watanabe, T., Langseth, M.G., and Anderson, R.N., 1977, Heat flow in back-arc basins of the western Pacific, in Talwani, M., and Pitman, W.C., III, eds., *Deep sea trenches and back-arc basins*: Maurice Ewing Series 1, Washington, American Geophysical Union, p. 137–161.

Whittaker, A., Bott, M.H.P., and Waghorn, G.D., 1992, Stresses and plate boundary forces associated with subduction: *Journal of Geophysical Research*, v. 97, p. 11,933–11,944.

Wiens, D.A., and Smith, G.P., 2003, Seismological constraints on structure and flow patterns within the mantle wedge, in Eiler, J., ed., *Inside the subduction factory*: American Geophysical Union Geophysical Monograph 138, p. 59–81, doi: 10.1029/138GM05.

Zoback, M.D., Townend, J., and Grollimund, B., 2002, Steady-state failure equilibrium and deformation of intra-plate lithosphere: *International Geology Review*, v. 44, p. 383–401.

*Manuscript received 11 June 2004; accepted 3 December 2004* ■



# Rock Solid.

## The Meiji ML Series of Polarizing Microscopes

Solid as a rock and crystal clear is what you will find when you look into these Polarizing Microscopes.

Each ML 9000 Series Microscope is equipped with swing-in, swing-out polarizer, analyzer and Bertrand lens for extra-bright, extra large interface figures. Every package includes DIN standard compensators (Mica 1/4 wave plate and first order red plate), strain free optics and a 360° graduated rotatable stage. You have a choice of 3 bodies - monocular, binocular or trinocular and a full range of accessories to create the ideal instrument for your specific needs and Meiji stands behind every instrument with its **“Limited Lifetime Warranty.”**

*For more information on these economically priced Microscopes, please call, FAX, write us or log on to our website today.*

## MEIJI TECHNO AMERICA

2186 Bering Drive, San Jose, CA 95131,  
Tel: 408.428.9654, FAX: 408.428.0472

**Toll Free Telephone: 800.832.0060 or visit our website at [www.meijitechno.com](http://www.meijitechno.com)**