Bounds on the viscosity coefficient of continental lithosphere from removal of mantle lithosphere beneath the Altiplano and Eastern Cordillera

Peter Molnar[1] and Carmala N. Garzione[2]

Received 8 March 2006; revised 29 November 2006; accepted 12 December 2006; published XX Month 2007.

[1] The rapid rise of the central Andean plateau between ∼10 and 6.8 Ma implies that mantle lithosphere, including eclogitized lower crust, was removed from beneath the region in that time interval; we infer from that removal that the average viscosity coefficient of mantle lithosphere was quite low when removal occurred. Using scaling laws for the growth of perturbations to the thickness of a dense layer over an inviscid substratum (Rayleigh-Taylor instability), we place bounds on the average viscosity coefficient for central Andean lithosphere. When compared with laboratory measurements of flow laws for olivine and eclogite, the allowed range of viscosity coefficients yields bounds on the temperature of ∼500–800°C at the Moho beneath this region and suggests that mean stresses across mantle lithosphere during continental deformation are less than ∼50 MPa. This range of temperature is comparable with, if a slightly lower, than we might expect for lithosphere approximately doubled in thickness and not yet equilibrated with the doubled crustal radioactivity. The mean deviatoric stress is comparable to that associated with stresses that drive plates and hence shows that lithospheric material is not too strong to prevent removal of its mantle part. **Citation:** Molnar, P., and C. N. Garzione (2007), Bounds on the viscosity coefficient of continental lithosphere from removal of mantle lithosphere beneath the Altiplano and Eastern Cordillera, *Tectonics*, *26*, XXXXXX, doi:10.1029/2006TC001964.

1. Introduction

[2] Removal of part or all of the mantle lithosphere offers a mechanism that can account for rapid changes in elevation of the overlying terrain and for rapid warming of the crust and remaining mantle lithosphere. To reconcile field observations of different kinds, many have appealed to some form of this process, either delamination of crust from mantle lithosphere [*Bird*, 1978, 1979] or convective instability [e.g., *England and Houseman*, 1989; *Houseman et al.*, 1981]. Although evidence consistent with such hypothesized processes exists, convincing demonstrations that such removal of mantle lithosphere has occurred have been harder to find. Many doubt that removal can occur, because the high viscosity of cold mantle lithosphere is thought to prevent it [e.g., *Morency et al.*, 2002; *Schott and Schmeling*, 1998]. Accordingly, a demonstration that this process has occurred requires evidence that cannot be explained by another process. We contend that the Altiplano of the central Andes offers the most convincing evidence for removal of mantle lithosphere, and we show that the implied average viscosity of the mantle lithosphere beneath the Altiplano, before it was removed, is consistent with laboratory experiments on the flow of olivine or eclogite at temperatures that are reasonable for the Altiplano at the time when dense material was removed.

[3] We carry out an analysis similar to that of *Molnar and Jones* [2004], who discussed the average viscosity beneath the Sierra Nevada before mantle lithosphere was removed from that area between ∼10 and 3.5 Ma. A variety of observations suggest that eclogite-rich mantle lithosphere was present beneath the Sierra at 10–12 Ma, but absent at 3.5 Ma. The most compelling evidence comes from a change in xenolith composition from eclogite and garnet pyroxenite prior to ∼8 Ma to a spinel peridotite composition since ∼1 Ma [*Ducea and Saleeby*, 1996, 1998; *Lee et al.*, 2000, 2001]. These xenoliths, derived from 40 to 100 km depths, suggest that eclogitic lower crust and mantle lithosphere were removed. The eruption of potassic basalts at 3.5 Ma is also consistent with removal of mantle lithosphere and provides a minimum age of ∼3.5 Ma on the timing of removal [*Farmer et al.*, 2002].

2. Removal of Eclogite-Rich Mantle Lithosphere From Beneath the Altiplano

[4] The central Andean plateau (Figure 1), with a width of ∼400 km and an average elevation of ∼4 km, is the second largest high plateau on Earth after the Tibetan plateau. In the central Andes where the plateau is widest, the Western and Eastern Cordillera, with peak elevations exceeding 6 km, bound the internally drained Altiplano basin whose average elevation is ∼3800 m. Active magmatism characterizes the Western Cordillera. The Eastern Cordillera and Altiplano basin preserve a history of folding and faulting. Constraints on the timing and distribution of

[1]Department of Geological Sciences and Cooperative Institute for Research in Environmental Science, University of Colorado, Boulder, Colorado, USA.

[2]Department of Earth and Environmental Sciences, University of Rochester, Rochester, New York, USA.

0278-7407/07/2006TC001964$12.00

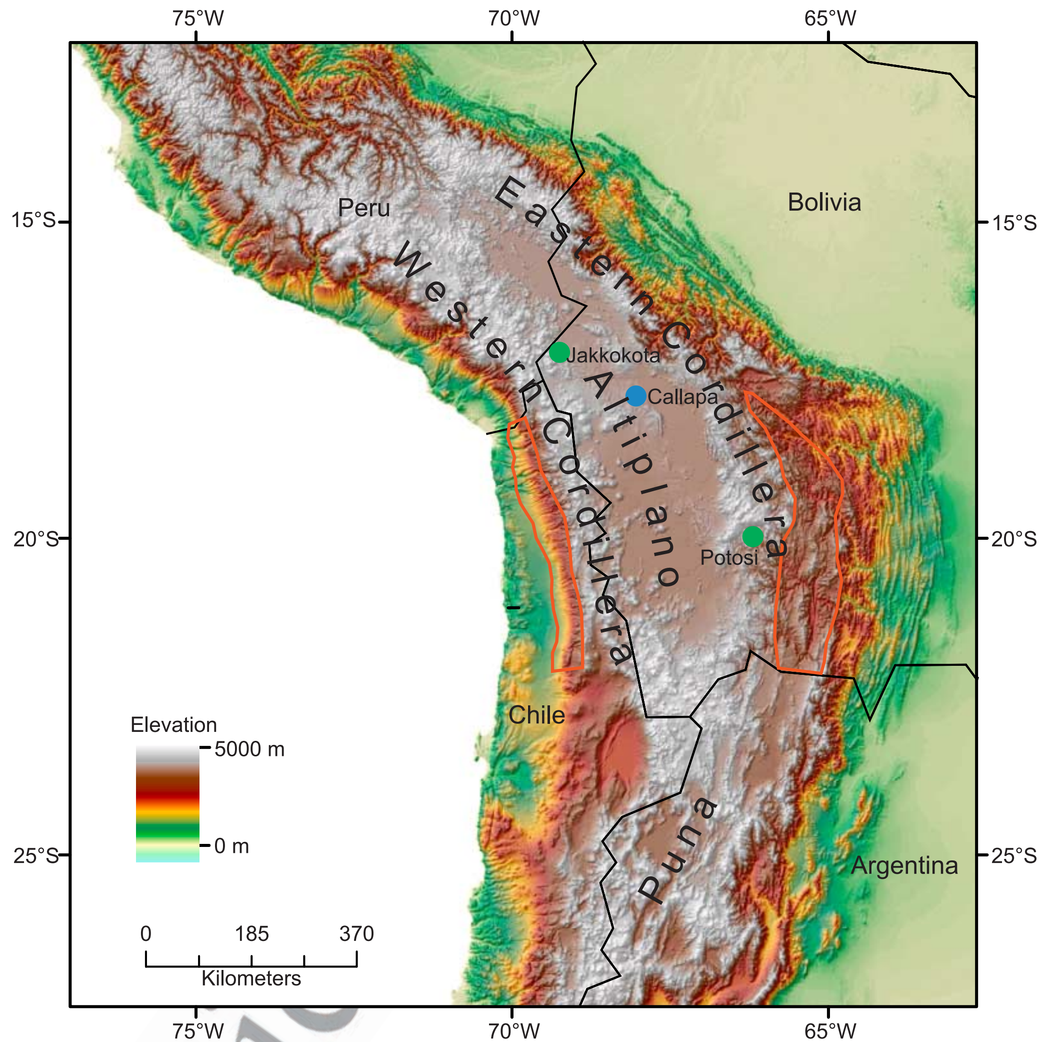

Figure 1. Elevations of the central Andean plateau between 12°S and 27°S, constructed with SRTM30 data set. Regions outlined in red show the extent of middle-late Miocene low-relief paleosurfaces that underwent rotation and incision beginning in late Miocene time. Green dots show locations of Miocene paleobotanical estimates of paleoelevation. Blue dots show locations of late Miocene paleoelevation estimates from $\delta^{18}O$ and Δ_{47} of authigenic carbonates shown in Figure 2.

deformation indicate that the Andean plateau, from the monoclinal structure that forms the western flank through the Altiplano and Eastern Cordillera, underwent horizontal shortening between ~40 and ~7 Ma [*Barnes et al.*, 2006; *Carrapa et al.*, 2005, 2006; *Deeken et al.*, 2006; *Ege et al.*, 2007; *Elger et al.*, 2005; *Farías et al.*, 2005; *Gillis et al.*, 2006; *Horton*, 2005; *Horton and DeCelles*, 1997; *Kley*, 1996; *Kraemer et al.*, 1999; *McQuarrie*, 2002; *Müller et al.*, 2002; *Sheffels*, 1990; *Victor et al.*, 2004].

[5] In the context of the shortening history of the central Andes, several authors have used geomorphic observations to infer that the surface of this area rose significantly since ~10 Ma [e.g., *Gubbels et al.*, 1993; *Isacks*, 1988; *Kennan et al.*, 1997]. Widespread, low-relief paleosurfaces on both the eastern slope of Eastern Cordillera [*Gubbels et al.*, 1993; *Kennan et al.*, 1997] and the western slope of the Western Cordillera [*Farías et al.*, 2005; *García and Hérail*, 2005; *Hoke et al.*, 2004; *Kober et al.*, 2006; *Schlunegger et al.*, 2006; *von Rotz et al.*, 2005; T. F. Schildgen et al., Tectonics of the western margin of the Altiplano in Southern Peru from river incision history, unpublished manuscript, 2006; G. D. Hoke et al., Geomorphic evidence for post-10 Ma uplift of the western flank of the central Andes (18°30′–22°S), submitted to *Tectonics*, 2007, hereinafter referred to as Hoke et al., submitted manuscript, 2007] reflect the remnants of low-relief drainage systems between ~7 Ma to 12 Ma in the Eastern Cordillera and until ~10 Ma on the western slope, after which widespread incision of both the eastern and western paleosurfaces had begun, in some areas more tightly dated to be before ~6.5 Ma. Reconstructions of the relief in these drainage systems have been used to infer ~1 to 2 km of surface uplift of the flanks of the

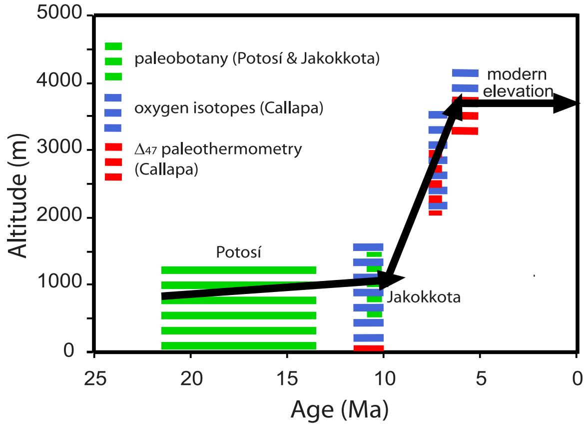

Figure 2. History of elevation change from multiple proxies in the northern Altiplano and Eastern Cordillera. The location of each record is shown on Figure 1. Paleobotany estimates from *Gregory-Wodzicki et al.* [1998] and *Gregory-Wodzicki* [2000]; Oxygen isotope estimates from *Garzione et al.* [2006]; and Δ_{47} estimates from *Ghosh et al.* [2006].

Andean plateau [e.g., *Kennan et al.*, 1997; *Hoke*, 2006; *Kober et al.*, 2006; Hoke et al., submitted manuscript, 2007]. Finally, concurrent with the rise of the Andean plateau, the locus of active crustal shortening migrated to the sub-Andes on the eastern margin of the belt after ~10 Ma [e.g., *Echavarria et al.*, 2003; *Moretti et al.*, 1996; *Uba et al.*, 2006].

[6] Quantitative estimates of paleoelevation come from marine deposits in the Altiplano basin, fossil leaf physiognomy, oxygen isotopes from carbonate sediment, and Δ_{47} paleothermometry (Figure 2). Shallow marine deposits of the 70–60 Ma El Molino Formation require that the Altiplano lay at sea level at the end of Cretaceous time [*Sempere et al.*, 1997]. Paleotemperature estimates derived from fossil leaf physiognomy in the northern Altiplano and Eastern Cordillera (Figure 1) suggest paleoelevations of no more than a third of the plateau's modern average height of ~4 km at ~15 to 20 Ma [*Gregory-Wodzicki*, 2000] and no more than half by ~10 Ma [*Gregory-Wodzicki et al.*, 1998]. Recent oxygen-isotope paleoaltimetry [*Garzione et al.*, 2006] and Δ_{47} paleothermometry [*Ghosh et al.*, 2006] suggest 3 ± 1 km between ~10 Ma and 6.8. Ma.

[7] Rapid surface uplift of several kilometers of a region as wide as the Andean plateau in ~3 ± 1 Myr of the late Miocene Epoch (Figure 2) reflects a rate of surface uplift that is too high to be generated by crustal shortening alone and requires the removal of relatively dense eclogite and mantle lithosphere [*Garzione et al.*, 2006]. Flow of middle-lower crust [e.g., *Gerbault et al.*, 2005; *Hindle et al.*, 2005; *Husson and Sempere*, 2003] may have redistributed crustal material and contributed to the nearly flat surface of the Altiplano. Lower crustal flow, however, cannot explain the simultaneous rise not only of the Altiplano but also of both the Eastern and Western Cordilleras some 300 km apart, for there is no obvious source of the volume of crustal material that must be injected into crust beneath these regions. In fact, crustal thickening in the Eastern and Western Cordilleras associated with crustal shortening and magmatism should result in a source of excess lower crustal material from the cordilleras, not injection of material beneath them.

[8] Constraints on the thermal and structural characteristics of the crust and mantle as well as the volcanic history are consistent with removal of mantle lithosphere and eclogitic lower crust from below the Andean plateau. The crustal thickness in the central Andes varies between 70 km below the highest topography in the Eastern and Western Cordilleras and 59 to 64 km in the central Altiplano, suggesting that the region is in approximate Airy isostatic balance [*Beck and Zandt*, 2002; *James*, 1971; *Yuan et al.*, 2002]. Seismic tomography of the mantle between 16° and 20°S, however, shows the lowest P wave speeds below the Altiplano–Eastern Cordillera transition and suggests that virtually all of the mantle lithosphere has been removed in this region [*Dorbath and Granet*, 1996; *Myers et al.*, 1998]. In addition, the crustal column beneath the Altiplano lacks typical high-speed lower crust, suggesting a felsic composition that is typical of upper crust [*Beck and Zandt*, 2002]. All of these observations support previous suggestions [*Kay and Mahlburg-Kay*, 1991; *Kay and Kay*, 1993] that both mantle lithosphere and eclogitic lower crust were removed from below the Altiplano and the western part of the Eastern Cordillera [*Beck and Zandt*, 2002]. The eruption of mafic lavas throughout the northern and central Altiplano beginning at ~7.5 to 5.5 Ma [*Carlier et al.*, 2005; *Lamb and Hoke*, 1997] and at ~7 to 3 Ma in the southern Altiplano and Puna [*Kay et al.*, 1994, 1999] have been inferred to reflect Late Miocene to Pliocene removal of eclogitic lower crust and mantle lithosphere beneath the Altiplano and Puna plateaus [*Kay et al.*, 1994; *Lamb and Hoke*, 1997]. High $^{3}He/^{4}He$ ratios across much of the Altiplano and Eastern Cordillera, interpreted to result from degassing of mantle-derived magmas, also support the inferred removal of mantle lithosphere, in this case for virtually the entire Altiplano [*Hoke et al.*, 1994].

3. Rayleigh-Taylor Instability

[9] The relatively low temperature of mantle lithosphere makes this material in most regions denser than asthenosphere when at the same pressure, even where eclogite comprises a negligible fraction of the lithosphere. The notable exception is beneath Archean cratons, where chemical differentiation has made mantle lithosphere intrinsically less dense than asthenosphere at the same temperature [e.g., *Jordan*, 1975; *Poudjom Djomani et al.*, 2001]. Thus mantle lithosphere in most regions should be unstable, and perturbations to its thickness should grow with time, unless diffusion of heat can erase the perturbations before they can grow. The presence of eclogite enhances this instability [e.g., *Kay and Mahlburg-Kay*, 1991; *Nelson*, 1991; *Jull and Kelemen*, 2001], particularly where a thick layer of eclogite overlies cold mantle lithosphere.

Table 1. List of Symbols

Symbol	Definition
A	empirical constant relating strain rate to stress difference in laboratory measurements of high-temperature flow
B	viscosity coefficient for non-Newtonian viscosity
C	empirically determined dimensionless factor that scales the growth rate of an instability for non-Newtonian viscosity; C depends on the wavelength of the perturbation and weakly on n
d	thickness of cooling layer
E	second invariant of the strain rate tensor
g	gravity
H_a	activation enthalpy
L	thickness of unstable layer (lithosphere)
ΔL_0	perturbation to the thickness of unstable layer (lithosphere)
n	power that relates strain rate to deviatoric stress in the constitutive relationship
R	universal gas constant
t'_b	elapsed time between initiation of a perturbation and when perturbations sinks to infinite depth
T	temperature (in kelvins)
$\Delta\rho$	density difference between unstable top layer (lithosphere) and underlying layer (asthenosphere)
e_{ij}	strain rate tensor
κ	coefficient of thermal diffusivity
τ_{ij}	deviatoric stress tensor

[10] The nonlinear relationship between deviatoric stress and strain rate that applies to most rock-forming minerals can make lithosphere more stable than it would be if deformation of the lithosphere obeyed Newtonian viscosity [e.g., *Conrad and Molnar*, 1999; *Houseman and Molnar*, 2001]. In the case of the Altiplano, however, where the crust has been thickened to roughly twice its normal value, this large perturbation to the thermal structure should be sufficient to prevent stabilization in a period shorter than tens of millions of years. The time constant for decay of thermal perturbations by diffusion of heat in a layer of thickness d is given by d^2/κ where κ is the coefficient of thermal diffusivity ($\sim 10^{-6}$ m^2 s^{-1}). For a layer 70 km thick (the maximum crustal thickness beneath the Andes, and much thinner than thickened lithosphere), the thermal time constant is 160 Myr, much longer than the time interval over which material was removed. Thus we need not consider diffusion of heat, and we can use scaling laws based on Rayleigh-Taylor instability, which results when an intrinsically dense layer overlies a less dense fluid.

[11] By exploiting the time for perturbations to grow by Rayleigh-Taylor instability as a mechanism for removal of lithosphere, we assume that the material that descends, first, does so in narrow plumes or sheets, and second, draws material from neighboring lithosphere. Thus we treat the understanding and the timescale for the downwelling of material as surrogates for those of removal of material from adjacent regions. As mass must be conserved, if some lithosphere descends, other, surely hotter, and therefore less dense material must rise to replace it. Because many factors can affect the distance scales involved in such removal, and much understanding of Rayleigh-Taylor instability remains to be gained, we do not try to specify where downwelling occurred or what its planform might have been.

3.1. Scaling Laws

[12] Using different approaches, *Canright and Morris* [1993] and *Houseman and Molnar* [1997] obtained scaling laws for the time that must elapse for a perturbation in the thickness of an unstable layer to sink to infinite depth. *Canright and Morris* [1993] treated the layer as a thin viscous sheet, and hence with no shear stress on vertical or horizontal planes, including at the top and bottom of the sheet. *Houseman and Molnar* [1997] considered the complete deformation field of a layer with a rigid top boundary. The extent to which mantle lithosphere (with or without eclogite) is detached from the overlying crust or tightly attached to it remains controversial, but by considering these extremes, we span the range of likely top boundary conditions on the mantle lithosphere. Although the analysis by *Canright and Morris* [1993] does not lend itself to consideration of additional complexity in the distribution of density or viscosity, numerical experiments by others [e.g., *Jull and Kelemen*, 2001; *Molnar et al.*, 1998] allow assessments of some such complexity. Also, *Conrad and Molnar* [1999] showed that convective instability, not only with vertically varying density and viscosity, but also with diffusion of heat, obeys a scaling law similar to that found by *Houseman and Molnar* [1997] for Rayleigh-Taylor instability. Thus ignoring diffusion of heat and considering Rayleigh-Taylor instability should introduce a negligible error to our estimates of viscosity coefficient.

[13] Most treatments of convection within the mantle beneath the lithosphere consider Newtonian viscosity, but in the lithosphere, where temperatures are relatively low, we must allow for a non-Newtonian constitutive relationship between stress and strain rate. As discussed below, laboratory measurements of rock-forming minerals suggest a relationship between deviatoric stress, τ_{ij}, and strain rate, e_{ij}, that can be written as

$$\tau_{ij} = BE^{(1-n)/n} e_{ij} \tag{1}$$

B is a temperature-dependent (and pressure-dependent) viscosity coefficient (with units of Pa s$^{1/n}$), and $E^2 = (1/2)\sum_{i,j} e_{ij}\, e_{ij}$ is the second invariant of the strain rate tensor, e_{ij}, and for olivine and most earth-forming minerals, the exponent $n = 3$–3.5. (Symbols are defined in Table 1.)

[14] The scaling laws developed by *Canright and Morris* [1993] and *Houseman and Molnar* [1997] consider constitutive laws like (1), but expressed with dimensionless quantities. To use these theoretical analyses and numerical experiments, we consider dimensionless distances obtained by dividing distances by lithospheric thickness, L, and dimensionless times, t', scaled using

$$t' = t(\Delta\rho g L/B)^n \tag{2}$$

in which $\Delta\rho$ is the difference between the density of the unstable layer (lithosphere) and that of the substratum (asthenosphere), and g is gravity.

[15] To approximate the behavior of an unstable layer, *Canright and Morris* [1993] used a dense thin viscous sheet, in which shear stresses on horizontal and vertical planes are negligible. This assumption treats any substratum as inviscid and the top boundary as free of shear stress. Moreover, to obtain an analytic solution, they assumed a constant $\Delta\rho$ within the layer. They derived an expression that can be integrated to give the time, t'_b, that a perturbation to the thickness of the sheet, $\Delta L'_0 = \Delta L_0/L$ (where ΔL_0 is the amplitude of the dimensional perturbation), grows to infinite thickness, $\Delta L' \to \infty$ (or thins to zero thickness). For $n = 3$, an approximate value of n for olivine, they obtained an analytic solution for t'_b:

$$t'_b(\Delta L'_0) = \frac{1+\Delta L'_0}{4\left[\left(1+\Delta L'_0\right)^2 - 1\right]^2} + \frac{1+\Delta L'_0}{8\left[\left(1+\Delta L'_0\right)^2 - 1\right]} + \frac{1}{16}\ln\frac{\Delta L'_0}{2+\Delta L'_0} \quad (3)$$

Not surprisingly, numerical integration of their governing equation for $n = 3.5$ yields a solution that does not differ much from that for $n = 3$ [*Molnar and Jones*, 2004].

[16] Using a less elegant approach than did *Canright and Morris* [1993], *Houseman and Molnar* [1997] considered a dense sheet with no horizontal or vertical movement of its top surface (rigid top boundary) and found that the elapsed time for a dimensionless harmonic perturbation $\Delta L'_0$ to grow to infinite depth could be approximated by

$$t'_b(\Delta L') = \left(\frac{n}{C}\right)^n \frac{\Delta L_0'^{\,1-n}}{n-1} \quad (4)$$

Here C is an empirically determined constant that is $\sim$0.76 for $n = 3$ in a layer of constant density (and using the definition of E given above). *Houseman and Molnar* [1997] interpolated between estimates of C from numerical experiments for $n = 3$ and $n = 5$ to suggest that for $n = 3.5$, $C \approx 0.9$.

[17] Although the different forms of (3) and (4) might suggest markedly different dependences of t'_b on $\Delta L'_0$, *Molnar and Jones* [2004] showed them to be quite similar at least for $n = 3$ and $n = 3.5$. Thus, if we can assign values not only to $\Delta\rho$, g, and L in (2), but also to t_b and $\Delta L'_0$, we can combine (3) or (4) with (2) to deduce a value of B from

$$B = \Delta\rho g L \left(\frac{t_b}{t'_b}\right)^{\frac{1}{n}} \quad (5)$$

3.2. Application of Scaling Laws to the Andean Plateau

[18] With a crustal thickness of 70 km for the Eastern Cordillera, roughly twice that of typical crust, we may infer that mantle lithosphere had also thickened approximately two times; hence $\Delta L'_0 \approx 1$. If $n = 3$, (3) yields $t'_b = 0.07$, and (4) gives $t'_b = 30.8$. With $n = 3.5$, appropriate for olivine or eclogite, an integration of *Canright and Morris*'s expressions yields $t'_b = 0.1$, and with (4) $t'_b = 46.4$. The large differences in values of t'_b result from the resistance to flow imposed by the top boundary in the case considered by *Houseman and Molnar* [1997] and the freedom of a layer to slip horizontally if its top is stress-free.

[19] As discussed above, the Andean plateau rose within an interval of $\sim$3 ± 1 Myr. We treat that interval as defining the duration between the time when a large perturbation to the thickness of the lithosphere grew to effectively infinite depth, and hence as defining $t_b = 3 \pm 1$ Myr. Although this process strictly applies to the time needed for a perturbation in thickness to grow to great depth, because the material that sinks must be drawn from the adjacent lithosphere, we treat that interval of time as also defining the period in which lithosphere is removed from beneath the Andean plateau. Also as noted above, the recent rise of the Altiplano of 3 ± 1 km, if compensated by removal of mantle lithosphere gives an estimate of the product $\Delta\rho L = 8.4(\pm 2.8) \times 10^6$ kg m^{-2}. This combination of parameters suggests that for a rigid top boundary $B \approx 9.1 \times 10^{12}$ Pa s$^{1/3}$ (for $n = 3$) or $B \approx 1.57 \times 10^{12}$ Pa s$^{1/3.5}$ ($n = 3.5$) and for a free top $B \approx 1.20 \times 10^{12}$ Pa s$^{1/3}$ ($n = 3$) and $B \approx 2.7 \times 10^{11}$ Pa s$^{1/3.5}$ ($n = 3.5$).

3.3. Discussion of Uncertainties

[20] Several sources contribute to uncertainty in B. First, consider the effect of errors in assumed parameters (see Table 2). Because the values of t'_b and t_b contribute to estimates of B in (5) only when the cube or 3.5th root is taken, their uncertainties are not very important. For instance, the 1-Myr uncertainty in the time for deblobbing contributes less than a 10% error to the estimates of B. If we assumed that the initial perturbation were 50% ($\Delta L'_0 \approx 0.5$) instead of 100%, the estimates of B for either free slip or rigid top boundaries would be 65% smaller, than those given above. Finally, the 1-km (33%) uncertainty in the elevation change between 10 and 6.8 Ma implies a 33% uncertainty in the product $\Delta\rho L$ in (5), and a corresponding uncertainty in B. Thus the combined uncertainties in the assumed parameters leads to an uncertainty in B that is less than 100% (less than a factor or 2) and closer to 50%.

[21] Assumptions of boundary conditions and in the applicability of scaling laws derived for idealized structures cause the largest uncertainties in estimates of B. The values of B for free slip at the top and a rigid top differ by a factor of 7.6 or 5.8 depending on the value of n. Surely the top is neither free nor rigid, and the likely value of B ought to lie between these values: 9.1×10^{12} Pa s$^{1/3.5}$ and 1.20×10^{12} Pa s$^{1/3.5}$ for $n = 3$, or between 1.57×10^{12} Pa s$^{1/3.5}$ and 2.7×10^{12} Pa s$^{1/3.5}$ for $n = 3.5$.

[22] The scaling laws in (2) and (3) consider an inviscid fluid beneath the unstable layer, but a finite viscosity of the asthenosphere will retard growth of an unstable mantle lithosphere. *Molnar et al.* [1998] ran numerical experiments for cases in which the viscosity coefficient decreased exponentially across the unstable layer to a constant value equal to that of the lower layer and found that the presence of a viscous substratum decreased values of C in (3) by about 20% below those with an inviscid substratum. Thus ignoring the viscosity of the substratum makes our estimates of B too large by $\sim$20%. Also we use the elapsed time for

Table 2. Summary of Contributions to Uncertainties

Assumption in Error	Effect on Estimated Value of B
30% error in t_b	uncertain by <10%
Smaller perturbation, $\Delta L'_0 \approx 0.5$ (not 1.0)	smaller by 65%
1-km (33%) error in the change in elevation	uncertain by 33%
No shear stress on top	overestimated by 5.8 to 7.6 times
No slip on top	underestimated by 5.8 to 7.6 times
Negligible asthenospheric viscosity	overestimated by ∼20%
Descent to 300–400 km (not infinite depth)	underestimated by ∼5%
Constant density in lithosphere	overestimated by 20–30%

the unstable layer to sink to infinite depth, but were we to use a depth of 300–400 km, we would augment the estimated values of B by only 5% [*Molnar and Jones*, 2004].

[23] The scaling laws also are based on theory or experiment for which the density anomaly is constant throughout the layer, but if the density anomaly decreases from a maximum at the top of the layer to zero at its bottom, growth is slower. *Houseman and Molnar*'s [1997] numerical experiments for such a density distribution yield values of C ∼20–30% smaller than those for constant density. Such a decrease in C leads to a comparably smaller inferred viscosity coefficient B.

[24] Finally, the assumption of a constant viscosity coefficient in the layer is unrealistic. A more realistic viscosity coefficient that decreased with depth in the layer would make the bottom part sink rapidly, but the upper part more slowly than the average value. Correcting for such a decrease with depth is difficult. Houseman's finite element code, which solves the equations of motion in a Lagrangian frame, becomes numerically unstable when elements deform from roughly equidimensional to elongated triangles, as they do when the large strains develop as blobs sink to great depth and the remaining lithosphere thins. We cannot quantify easily the error introduced by using the scaling laws to estimate the average viscosity of the lithosphere, but the numerical experiments on convective instability by *Conrad and Molnar* [1999] show that the same scaling laws used here apply to those cases for which viscosity decreased exponentially with temperature and hence approximately exponentially with depth through the unstable layer.

[25] The arguments given above suggest that allowance for a linearly decreasing density anomaly across the layer and of a finite viscosity of the substratum would make the average value of B for the Altiplano ∼50% smaller than the values given above, and uncertain by ∼50%. Thus, if n = 3.5 we use $B \approx 7.9\ (\pm 3.9) \times 10^{11}$ Pa s$^{1/3.5}$ for a free top and $B \approx 1.4\ (\pm 0.7) \times 10^{11}$ Pa s$^{1/3.5}$ for a rigid top, or if n = 3, $B \approx 4.6\ (\pm 2.3) \times 10^{12}$ Pa s$^{1/3}$ for a free top boundary and $B \approx 6.0\ (\pm 3.0) \times 10^{11}$ Pa s$^{1/3}$ for a rigid top.

4. Discussion

[26] Because few readers have experience with the viscosity coefficient B (or its units of Pa s$^{1/3}$ or Pa s$^{1/3.5}$), we address the significance of the range of inferred values given above in two ways: what they imply (1) for magnitudes of stress in the lithosphere and (2) for temperatures at depth. First, B relates magnitudes of stress to strain rates as shown in equation (1), and second, using laboratory measurements of B as a function of temperature, we may use the estimates given above to infer temperatures of the upper mantle at the Moho.

[27] Much of the crustal thickening in the Altiplano and Eastern Cordillera occurred between 40 and ∼10 Ma [e.g., *Elger et al.*, 2005; *McQuarrie*, 2002]. Suppose that crust doubled in thickness in 30 Myr, so that the average strain rate was $e = 1 \times 10^{-15}$ s^{-1}. If thickening developed by pure shear so that horizontal compression occurred at the same (but negative) rate, then $E = 1 \times 10^{-15}$ s^{-1} also. Corresponding average deviatoric stresses across the mantle lithosphere, given by $\tau = Be^{1/n}$, are $7.3 \le \tau \le 41$ MPa for $1.4 \le B \le 7.9 \times 10^{11}$ Pa s$^{1/3.5}$ and $6.0 \le \tau \le 46$ MPa for $0.60 \le B \le 4.6 \times 10^{12}$ Pa s$^{1/3}$. Because these are averages for the entire mantle lithosphere, maximum values should be larger. These average bounds span those inferred for Tibet [e.g., *England and Molnar*, 1997; *Flesch et al.*, 2001], where lithosphere has been thinned. Most important, they are not so large as to prohibit deformation of mantle lithosphere, for such average stresses are comparable to those that drive plates [e.g., *Chapple and Tullis*, 1977; *Forsyth and Uyeda*, 1975; *McKenzie*, 1972].

[28] To compare these inferred values of B with those derived from laboratory measurements, we first must take into account the temperature dependence of B. Experimentally determined power law relationships between stress and strain rate, such as those for olivine [e.g., *Goetze*, 1978; *Karato and Wu*, 1993] or eclogite [*Jin et al.*, 2001], can be written as [e.g., *Molnar et al.*, 1998]

$$B(T) = 3^{-\frac{n+1}{2n}} \left(\frac{A}{2}\right)^{-\frac{1}{n}} \exp\left(\frac{H_a}{nRT}\right) \tag{6}$$

where A and n are parameters determined experimentally in the laboratory, H_a is the activation enthalpy, T is temperature in kelvins, and R is the universal gas constant. *Hirth and Kohlstedt* [1996, 2003] showed that for olivine either saturated with hydrogen (“wet”) and with small amounts of hydrogen (“dry”), n differs little. The addition of hydrogen reduces the effective viscosity largely by reducing A, by nearly 2 orders of magnitude. Here, we ignore the high-stress, low-temperature flow law suggested by *Evans and Goetze* [1979] which shows lower effective values of B than

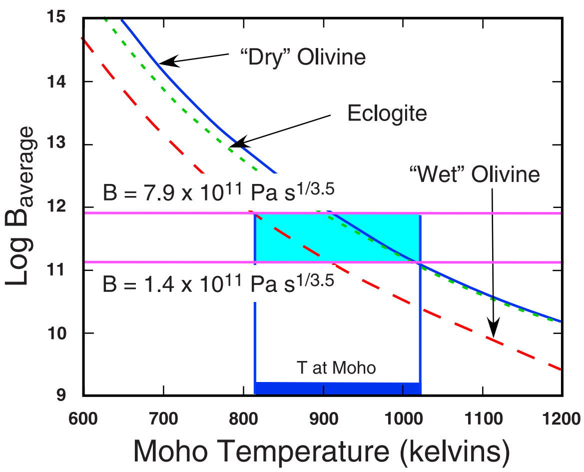

Figure 3. Average values $B(T)$ in mantle lithosphere as function of temperature at the Moho, bounds of average values $B(T)$ obtained from applying scaling laws for Rayleigh-Taylor instability to conditions in the Altiplano (horizontal lines), and corresponding ranges of Moho temperatures (vertical lines). We integrated values of B appropriate for "wet" and "dry" olivine and for eclogite from 1600 K (asthenosphere) to different values of temperature at the Moho. For "wet" olivine (red line) we assumed $n = 3$, $H_a = 420$ kJ mol^{-1}, and $A = 1.9 \times 10^3$ MPa^{-3} s^{-1}, for "dry" olivine (blue line) $n = 3.5$, $H_a = 540$ kJ mol^{-1}, and $A = 2.4 \times 10^5$ MPa$^{-3.5}$ s^{-1} [*Hirth and Kohlstedt*, 1996, 2003], and for eclogite (green line) $n = 3.4$, $H_a = 480$ kJ mol^{-1}, and $A = 2.0 \times 10^3$ MPa$^{-3.4}$ s^{-1} [*Jin et al.*, 2001].

those given by (6) when temperatures drop below ~900 K [e.g., *Evans and Goetze*, 1979; *Goetze*, 1978]. We recognize that (6) cannot be extrapolated reliably to low temperatures, a result corroborated by deformation using a diamond anvil [*Wenk et al.*, 2004]. In the temperature range of interest here, however, including the high-stress flow law will alter inferred Moho temperatures by amounts much smaller than those due to uncertainties in B.

[29] We may calculate average values of B by integrating $B(T)$ in (6) from the temperature at the Moho to 1600 K, a reasonable temperature for the base of the lithosphere. To carry out this integration, we assume that the temperature gradient across the mantle lithosphere is constant, and we used parameters in flow laws for both "wet" and "dry" olivine [*Hirth and Kohlstedt*, 1996, 2003] and for eclogite [*Jin et al.*, 2001] (Figure 3). A comparison of the calculations based on flow laws with the ranges of values of B given above based on the scaling laws for Rayleigh-Taylor instability, and for $n = 3.5$, suggest temperatures at the Moho of ~814–905 K (540–630°C) for "wet" olivine and ~910–1015 K (640–740°C) for "dry" olivine or eclogite (Figure 3). Allowance for an uncertainty of a factor of 2 in the values of B, estimated either from laboratory measurements or from the scaling laws and geological observations widens the range of temperatures by ~50 K to 490–680°C and 590–790°C. The upper of these bounds differs little from the Moho temperature that *Morency and Doin* [2004] inferred to be necessary for removal of mantle lithosphere obeying a plastic constitutive relationship.

[30] These ranges of inferred temperatures call for a relatively cool upper mantle (including eclogite) beneath the Altiplano before mantle lithosphere was removed, but if in fact, a thick layer of eclogitized lower crust were removed, such values are not unreasonable for thickened lower crust. For instance, if normal lithosphere with a temperature at the Moho of 490–790°C doubled in thickness between ~40 and ~10 Ma, then at 10 Ma, the temperature structure would have been slowly evolving toward a warmer crust and uppermost mantle, because of the doubled thickness of radiogenic heat production in it. With the long thermal time constant of thickened lithosphere, however, the temperature near the Moho should have changed little during this 30-Myr span, a result corroborated by numerical calculations of *Babeyko et al.* [2002].

[31] This simple scenario of homogenous thickening overlooks any interaction of the lower lithosphere with the downgoing slab, which clearly could affect the thermal structure of the overlying crust. It also overlooks the possibility that some or all of the mantle lithosphere was removed earlier, for instance, at ~25 Ma. Widespread magmatism spanning the width of the present-day plateau seems to have begun near this time [e.g., *Allmendinger et al.*, 1997; *James and Sacks*, 1999]. On the basis of the westward sweep of magmatic activity, *James and Sacks* [1999] associated this magmatism with a steepening of Nazca slab subduction from nearly flat to its present-day dip of ~30°. Such a change in the dip of the subducting slab beneath the Western Cordillera might have led to some removal of Andean lithosphere, but the occurrence of volcanism implies that the slab was deeper than ~120 km. Thus it does not imply that the Andean lithosphere was unusually thin.

[32] Most measurements of heat flux from the Altiplano show higher than normal values [*Henry and Pollack*, 1988; *Springer and Förster*, 1998; *Uyeda and Watanabe*, 1982], with *Henry and Pollack* [1988] giving an average of 84 mW/m^2. Values as large as 84 mW/m^2 would suggest steady state Moho temperatures above 1000°C if relatively typical distributions of radiogenic heat production were assumed [e.g., *Springer*, 1999], but for a thickened layer of radiogenic heat production and the same surface heat flux, equilibrium Moho temperatures 700°C would also be possible. Because of the transient temperature distribution in the crust, however, the heat flux measurements place only weak constraints on the Moho temperature at ~10 Ma. We consider the range of 490–790°C to be cooler than most would expect for a subduction zone, but not unreasonably cool. *Babeyko et al.* [2002], in fact, suggest that convection within the crust has altered the thermal structure considerably.

5. Conclusions

[33] Recent analyses of stable isotopes [*Garzione et al.*, 2006; *Ghosh et al.*, 2006] in authigenic carbonates in the

Altiplano corroborate previous inferences of a post-10 Ma rise of the Andean plateau [e.g., *Allmendinger et al.*, 1997; *Gregory-Wodzicki*, 2000; *Gubbels et al.*, 1993; *Isacks*, 1988; *Kay et al.*, 1994; *Lamb and Hoke*, 1997] and suggest that an elevation change of 3±1 km occurred in only 3 ± 1 Myr between ~10 and 6.8 Ma. The combination of such a large change in mean elevation and the rapidity with which it occurred virtually requires removal not only of all of the mantle lithosphere but also a layer of dense eclogite beneath the overlying crust [*Garzione et al.*, 2006].

[34] We use scaling laws that relate the time for a perturbation to the thickness of dense layer overlying a less dense layer to grow to infinite depth and the measured time of 3 ± 1 Myr to estimate the average viscosity coefficient for the mantle lithosphere that underlay the Altiplano before 10 Ma and then was removed: $1.4\ (\pm 0.7) \times 10^{11}$ Pa $s^{1/3.5}$ B < $8.8\ (\pm 4.4) \times 10^{11}$ Pa $s^{1/3.5}$. The change in elevation of 3 ± 1 km constrains the mass per unit area of the layer that was removed, so that the largest sources of uncertainty in the estimated average viscosity coefficient are the boundary condition at the top of the layer and the approximation of lithosphere as a layer of constant viscosity used to derive the scaling laws.

[35] A comparison of the range of estimates of viscosity coefficient with those based on calculated averages from temperature-dependent constitutive relations based on laboratory measurements yield a range of likely temperatures between ~500°C and ~800°C at the Moho of the Altiplano at ~10 Ma, just before mantle lithosphere was removed. Such values are consistent with those appropriate for doubling the thickness of the lithosphere between ~40 and ~10 Ma. In addition, this range of average values of viscosity coefficient implies average deviatoric stresses within the mantle lithosphere less than ~50 MPa. Thus, insofar as these scaling laws for Rayleigh-Taylor instability can be applied to removal of mantle lithosphere, the corresponding material parameters match those deduced from laboratory experiments and show that high strengths of earth-forming minerals offer no obstacle to removal of tectonically thickened mantle lithosphere.

[36] **Acknowledgments.** One of us (P.M.) thanks the Miller Institute of Berkeley for support while the first draft of this paper was written, and we thank Greg Hoke for help with Figure 1 and Philip England and Nadine McQuarrie for constructive criticism. This research was supported in part by the National Science Foundation through grants EAR-0230232 and EAR-0337509.

References

Allmendinger, R. W., T. E. Jordan, S. M. Kay, and B. L. Isacks (1997), The evolution of the Altiplano-Puna plateau of the central Andes, *Annu. Rev. Earth Planet Sci.*, *25*, 139–174.

Babeyko, A. Yu, S. V. Sobolev, R. B. Trumbull, O. Oncken, and L. L. Lavier (2002), Numerical models of crustal scale convection and partial melting beneath the Altiplano-Puna Plateau, *Earth Planet. Sci. Lett.*, *199*, 373–388.

Barnes, J. B., T. A. Ehlers, N. McQuarrie, P. B. O'Sullivan, and J. D. Pelletier (2006), Eocene to recent variations in erosion across the central Andean fold-thrust belt, northern Bolivia: Implications for plateau evolution, *Earth Planet. Sci. Lett.*, *248*, 118–133.

Beck, S. L., and G. Zandt (2002), The nature of orogenic crust in the central Andes, *J. Geophys. Res.*, *107*(B10), 2230, doi:10.1029/2000JB000124.

Bird, P. (1978), Initiation of intracontinental subduction in the Himalaya, *J. Geophys. Res.*, *83*, 4975–4987.

Bird, P. (1979), Continental delamination and the Colorado Plateau, *J. Geophys. Res.*, *84*, 7561–7571.

Canright, D., and S. Morris (1993), Buoyant instability of a viscous film over a passive fluid, *J. Fluid Mech.*, *255*, 340–372.

Carlier, G., J. P. Lorand, J. P. Liégeois, M. Fornari, P. Soler, V. Carlotto, and J. Cárdenas (2005), Potassic-ultrapotassic mafic rocks delineate two lithospheric mantle blocks beneath the southern Peruvian Altiplano, *Geology*, *33*, 601–604.

Carrapa, B., D. Adelmann, G. E. Hilley, E. Mortimer, E. R. Sobel, and M. R. Strecker (2005), Oligocene range uplift and development of plateau morphology in the southern central Andes, *Tectonics*, *24*, TC4011, doi:10.1029/2004TC001762.

Carrapa, B., M. R. Strecker, and E. R. Sobel (2006), Cenozoic orogenic growth in the Central Andes: Evidence from sedimentary rock provenance and apatite fission track thermochronology in the Fiambalá Basin, southernmost Puna Plateau margin (NW Argentina), *Earth Planet. Sci. Lett.*, *247*, 82–100.

Chapple, W. M., and T. E. Tullis (1977), Evaluation of the forces that drive the plates, *J. Geophys. Res.*, *82*, 1967–1984.

Conrad, C. P., and P. Molnar (1999), Convective instability of a boundary layer with temperature- and strain-rate-dependent viscosity in terms of 'available buoyancy,' *Geophys. J. Int.*, *139*, 51–68.

Deeken, A., E. R. Sobel, I. Coutand, M. Haschke, U. Riller, and M. R. Strecker (2006), Development of the southern Eastern Cordillera, NW Argentina, constrained by apatite fission track thermochronology: From early Cretaceous extension to middle Miocene shortening, *Tectonics*, *25*, TC6003, doi:10.1029/2005TC001894.

Dorbath, C., and M. Granet (1996), Local earthquake tomography of the Altiplano and the Eastern Cordillera of northern Bolivia, *Tectonophysics*, *259*, 117–136.

Ducea, M. N., and J. B. Saleeby (1996), Buoyancy sources for a large, unrooted mountain range, the Sierra Nevada, California: Evidence from xenolith thermobarometry, *J. Geophys. Res.*, *101*, 8229–8244.

Ducea, M., and J. Saleeby (1998), A case for delamination of the deep batholithic crust beneath the Sierra Nevada, California, *Int. Geol. Rev.*, *40*, 78–93.

Echavarria, L., R. Hernández, R. Allmendinger, and J. Reynolds (2003), Subandean thrust and fold belt of northwestern Argentina: Geometry and timing of the Andean evolution, *AAPG Bull*, *87*, 965–985.

Ege, H., E. R. Sobel, E. Scheuber, and V. Jacobshagen (2007), Exhumation history of the southern Altiplano plateau (southern Bolivia) constrained by apatite fission track thermochronology, *Tectonics*, *26*, TC1004, doi:10.1029/2005TC001869.

Elger, K., O. Oncken, and J. Glodny (2005), Plateau-style accumulation of deformation: Southern Altiplano, *Tectonics*, *24*, TC4020, doi:10.1029/2004TC001675.

England, P. C., and G. A. Houseman (1989), Extension during continental convergence, with application to the Tibetan Plateau, *J. Geophys. Res.*, *94*, 17,561–17,579.

England, P., and P. Molnar (1997), Active deformation of Asia: From kinematics to dynamics, *Science*, *278*, 647–650.

Evans, B., and C. Goetze (1979), The temperature variation of hardness of olivine and its implication for polycrystalline yield stress, *J. Geophys. Res.*, *84*, 5505–5524.

Farías, M., R. Charrier, D. Comte, J. Martinod, and G. Hérail (2005), Late Cenozoic deformation and uplift of the western flank of the Altiplano: Evidence from the depositional, tectonic, and geomorphologic evolution and shallow seismic activity (northern Chile at 19°30′S), *Tectonics*, *24*, TC4001, doi:10.1029/2004TC001667.

Farmer, G. L., A. F. Glazner, and C. R. Manley (2002), Did lithospheric delamination trigger Late Cenozoic potassic volcanism in the southern Sierra Nevada, California?, *Geol. Soc. Am. Bull.*, *114*, 754–768.

Flesch, L. M., A. J. Haines, and W. E. Holt (2001), Dynamics of the India-Eurasia collision zone, *J. Geophys. Res.*, *106*, 16,435–16,460.

Forsyth, D., and S. Uyeda (1975), On the relative importance of the driving forces of plate motions, *Geophys. J. R. Astron. Soc.*, *43*, 163–200.

García, M., and G. Hérail (2005), Fault-related folding, drainage network evolution and valley incision during the Neogene in the Andean Precordillera of northern Chile, *Geomorphology*, *65*, 279–300.

Garzione, C. N., P. Molnar, J. C. Libarkin, and B. MacFadden (2006), Rapid Late Miocene rise of the Bolivian Altiplano: Evidence for removal of mantle lithosphere, *Earth Planet. Sci. Lett*, *241*, 543–556.

Gerbault, M., J. Martinod, and G. Hérail (2005), Possible orogeny-parallel lower crustal flow and thickening in the central Andes, *Tectonophysics*, *399*, 59–72.

Ghosh, P., C. N. Garzione, and J. M. Eiler (2006), Rapid uplift of the Altiplano revealed in ^{13}C-^{18}O bonds in paleosol carbonates, *Science*, *311*, 511–515.

Gillis, R. J., B. K. Horton, and M. Grove (2006), Thermochronology, geochronology, and upper crustal structure of the Cordillera Real: Implications for Cenozoic exhumation of the central Andean plateau, *Tectonics*, *25*, TC6007, doi:10.1029/2005TC001887.

Goetze, C. (1978), The mechanisms of creep in olivine, *Philos. Trans. R. Soc. London, Ser. A*, *288*, 99–119.

Gregory-Wodzicki, K. M. (2000), Relationships between leaf morphology and climate, Bolivia: Implications for estimating paleoclimate from fossil floras, *Paleobiology*, *26*(4), 668–688.

Gregory-Wodzicki, K. M., W. C. McIntosh, and K. Velasquez (1998), Climatic and tectonic implications of the late Miocene Jakokkota flora, Bolivian Altiplano, *J. S. Am. Earth Sci.*, *11*, 533–560.

Gubbels, T., B. Isacks, and E. Farrar (1993), High level surfaces, plateau uplift, and foreland basin development, Bolivian central Andes, *Geology*, *21*, 695–698.

Henry, S. G., and H. N. Pollack (1988), Terrestrial heat flow above the Andean subduction zone in Bolivia and Peru, *J. Geophys. Res.*, *93*, 15,153–15,162.

Hindle, D., J. Kley, O. Oncken, and S. Sobolev (2005), Crustal balance and crustal flux from shortening estimates in the central Andes, *Earth Planet. Sci. Lett.*, *230*, 113–124.

Hirth, G., and D. L. Kohlstedt (1996), Water in the oceanic upper mantle: implications for rheology, melt extraction and the evolution of the lithosphere, *Earth Planet. Sci. Lett.*, *144*, 93–108.

Hirth, G., and D. Kohlstedt (2003), Rheology of the upper mantle and the mantle wedge: a view from the experimentalists, in *The Subduction Factory*, *Geophys. Monogr. Ser.*, vol. 138, edited by J. Eiler, pp. 83–105, AGU, Washington, D. C.

Hoke, G. D. (2006), The influence of climate and tectonics in the geomorphology of the western slope of the central Andes, Chile and Peru, Ph.D. dissertation, Cornell Univ., Ithaca, N. Y.

Hoke, G. D., B. L. Isacks, T. E. Jordan, and J. S. Yu (2004), Groundwater sapping origin for the giant quebradas of northern Chile, *Geology*, *32*, 605–608.

Hoke, L., D. R. Hilton, S. H. Lamb, K. Hammerschmidt, and H. Friedrichsen (1994), ^{3}He evidence for a wide zone of active mantle melting beneath the central Andes, *Earth Planet. Sci. Lett.*, *128*, 341–355.

Horton, B. K. (2005), Revised deformation history of the central Andes: Inferences from Cenozoic foredeep and intermontane basins of the Eastern Cordillera, Bolivia, *Tectonics*, *24*, TC3011, doi:10.1029/2003TC001619.

Horton, B. K., and P. G. DeCelles (1997), The modern foreland basin system adjacent to the central Andes, *Geology*, *25*, 895–898.

Houseman, G. A., and P. Molnar (1997), Gravitational (Rayleigh-Taylor) instability of a layer with nonlinear viscosity and convective thinning of continental lithosphere, *Geophys. J. Int.*, *128*, 125–150.

Houseman, G., P. Molnar (2001), Mechanisms of lithospheric renewal associated with continental orogeny, in *Continental Reworking and Reactivation*, edited by J. A. Miller et al., *Geol. Soc. Spec. Publ.*, *184*, 13–38.

Houseman, G. A., D. P. McKenzie, and P. Molnar (1981), Convective instability of a thickened boundary layer and its relevance for the thermal evolution of continental convergent belts, *J. Geophys. Res.*, *86*, 6115–6132.

Husson, L., and T. Sempere (2003), Thickening the Altiplano crust by gravity-driven crustal channel flow, *Geophys. Res. Lett.*, *30*(5), 1243, doi:10.1029/2002GL016877.

Isacks, B. L. (1988), Uplift of the central Andean plateau and bending of the Bolivian orocline, *J. Geophys. Res.*, *93*, 3211–3231.

James, D. E. (1971), Andean crustal and upper mantle structure, *J. Geophys. Res.*, *76*, 3246–3271.

James, D. E., I. S. Sacks (1999), Cenozoic formation of the Andes: A geophysical perspective, in *Geology and Ore Deposits of the Central Andes*, edited by B. J. Skinner, *Spec. Publ. Soc. Econ. Geol.*, *7*, 1–25.

Jin, Z. M., J. Zhang, H. W. Green, and S. Jin (2001), Eclogite rheology: Implications for subducted lithosphere, *Geology*, *29*, 667–670.

Jordan, T. H. (1975), The continental tectosphere, *Rev. Geophys.*, *13*, 1–12.

Jull, M., and P. B. Kelemen (2001), On the conditions for lower crustal convective instability, *J. Geophys. Res.*, *106*, 6423–6446.

Karato, S.-I., and P. Wu (1993), Rheology of the upper mantle: A synthesis, *Science*, *260*, 771–778.

Kay, R. W., and S. M. Kay (1993), Delamination and delamination magmatism, *Tectonophysics*, *219*, 177–189.

Kay, R. W., and S. Mahlburg-Kay (1991), Creation and destruction of lower continental crust, *Geol. Rundsch.*, *80*(2), 259–278.

Kay, S. M., B. Coira, and J. Viramonte (1994), Young mafic back arc volcanic rocks as indicators of continental lithospheric delamination beneath the Argentine Puna plateau, central Andes, *J. Geophys. Res.*, *99*, 24,323–24,339.

Kay, S. M., C. Mpodozis, B. Coira (1999), Magmatism, tectonism, and mineral deposits of the Central Andes (22°–33°S latitude), in *Geology and Ore Deposits of the Central Andes*, edited by B. J. Skinner, *Spec. Publ. Soc. Econ. Geol.*, *7*, 27–59.

Kennan, L., S. Lamb, L. Hoke (1997), High altitude paleosurfaces in the Bolivian Andes: Evidence for Late Cenozoic surface uplift, in *Paleosurfaces: Recognition, Reconstruction and Paleoenvironmental Interpretation*, edited by M. Widdowson, *Geol. Soc. Spec. Publ.*, 120, 307–324.

Kley, J. (1996), Transition from basement-involved to thin-skinned thrusting in the Cordillera Oriental of southern Bolivia, *Tectonics*, *15*, 763–775.

Kober, F., F. Schlunegger, G. Zeilinger, H. Schneider (2006), Surface uplift and climate change: The geomorphic evolution of the Western Escarpment of the Andes of northern Chile between the Miocene and present, in *Tectonics, Climate, and Landscape Evolution*, edited by S. D. Willett et al., *Spec. Pap. Geol. Soc. Am.*, *398*, 75–86.

Kraemer, B., D. Adelmann, M. Alten, W. Schnurr, K. Erpenstein, E. Kiefer, P. van den Bogaard, and K. Görler (1999), Incorporation of the Paleogene foreland into the Neogene Puna Plateau: The Salar de Antofalla area, NW Argentina, *J. S. Am. Earth Sci.*, *12*, 157–182.

Lamb, S., and L. Hoke (1997), Origin of the high plateau in the Central Andes, Bolivia, South America, *Tectonics*, *16*, 623–649.

Lee, C.-T., Q.-Z. Yin, R. L. Rudnick, J. T. Chesley, and S. B. Jacobsen (2000), Re-Os isotopic evidence for pre-Miocene delamination of lithospheric mantle beneath the Sierra Nevada, California, *Science*, *289*, 1912–1916.

Lee, C., R. L. Rudnick, and G. H. Brimhall Jr. (2001), Deep lithospheric dynamics beneath the Sierra Nevada during the Mesozoic and Cenozoic as inferred from xenolith petrology, *Geochem. Geophys. Geosyst.*, *2*(12), doi:10.1029/2001GC000152.

McKenzie, D. P. (1972), Plate tectonics, in *The Nature of the Solid Earth*, edited by E. C. Robertson, pp. 323–360, McGraw-Hill, New York.

McQuarrie, N. (2002), The kinematic history of the Central Andean fold-thrust belt, Bolivia; implications for building a high plateau, *Geol. Soc. Am. Bull.*, *114*, 950–963.

McQuarrie, N., and P. DeCelles (2001), Geometry and structural evolution of the central Andean backthrust belt, Bolivia, *Tectonics*, *20*, 669–692.

Molnar, P., and C. H. Jones (2004), A test of laboratory based rheological parameters of olivine from an analysis of late Cenozoic convective removal of mantle lithosphere beneath the Sierra Nevada, California, USA, *Geophys. J. Int.*, *156*, 555–564.

Molnar, P., G. A. Houseman, and C. P. Conrad (1998), Rayleigh-Taylor instability and convective thinning of mechanically thickened lithosphere: Effects of non-linear viscosity decreasing exponentially with depth and of horizontal shortening of the layer, *Geophys. J. Int.*, *133*, 568–584.

Morency, C., and M.-P. Doin (2004), Numerical simulations of the mantle lithosphere delamination, *J. Geophys. Res.*, *109*, B03410, doi:10.1029/2003JB002414.

Morency, C., M.-P. Doin, and C. Dumoulin (2002), Convective destabilization of a thickened continental lithosphere, *Earth Planet. Sci. Lett.*, *202*, 303–320.

Moretti, I., P. Baby, E. Mendez, and D. Zubieta (1996), Hydrocarbon generation in relation to thrusting in the Sub Andean zone from 18 to 22°S, Bolivia, *Pet. Geosci.*, *2*, 17–28.

Müller, J. P., J. Kley, and V. Jacobshagen (2002), Structure and Cenozoic kinematics of the Eastern Cordillera, southern Bolivia (21°S), *Tectonics*, *21*(5), 1037, doi:10.1029/2001TC001340.

Myers, S. C., S. Beck, G. Zandt, and T. Wallace (1998), Lithospheric-scale structure across the Bolivian Andes from tomographic images of velocity and attenuation for *P* and *S* waves, *J. Geophys. Res.*, *103*, 21,233–21,252.

Nelson, K. D. (1991), A unified view of craton evolution motivated by recent deep seismic reflection and refraction results, *Geophys. J. Int.*, *105*, 25–35.

Poudjom Djomani, Y. H., S. Y. O'Reilly, W. L. Griffin, and P. Morgan (2001), The density structure of subcontinental lithosphere through time, *Earth Planet. Sci. Lett.*, *184*, 605–621.

Schlunegger, F., G. Zeilinger, A. Kounov, F. Kober, and B. Hüsser (2006), Scale of relief growth in the forearc of the Andes of Northern Chile (Arica latitude, 18°S), *Terra Nova*, *18*, 217–223.

Schott, B., and H. Schmeling (1998), Delamination and detachment of a lithospheric root, *Tectonophysics*, *296*, 225–247.

Sempere, T., R. F. Butler, D. R. Richards, L. G. Marshall, W. Sharp, and C. C. Swisher III (1997), Stratigraphy and chronology of Upper Cretaceous–lower Paleogene strata in Bolivia and northwest Argentina, *Geol. Soc. Am. Bull.*, *109*, 709–727.

Sheffels, B. (1990), Lower bound on the amount of crustal shortening in the central Bolivian Andes, *Geology*, *l8*, 812–815.

Springer, M. (1999), Interpretation of heat-flow density in the central Andes, *Tectonophysics*, *306*, 377–395.

Springer, M., and A. Förster (1998), Heat-flow density across the central Andean subduction zone, *Tectonophysics*, *291*, 123–139.

Uba, C. E., C. Heubeck, and C. Hulka (2006), Evolution of the late Cenozoic Chaco foreland basin, southern Bolivia, *Basin Res.*, *18*, 145–170.

Uyeda, S., and T. Watanabe (1982), Terrestrial heat flow in western South America, *Tectonophysics*, *83*, 63–70.

Victor, P., O. Oncken, and J. Glodny (2004), Uplift of the western Altiplano plateau: Evidence from the Precordillera between 20° and 21°S (northern Chile), *Tectonics*, *23*, TC4004, doi:10.1029/2003TC001519.

von Rotz, R., F. Schlunegger, F. Heller, and I. Villa (2005), Assessing the age of relief growth in the Andes of northern Chile: Magneto-polarity chronologies from Neogene continental sections, *Terra Nova*, *17*, 462–471.

Wenk, H.-R., I. Lonardelli, J. Pehl, J. Devine, V. Prakapenka, G. Shen, and H.-K. Mao (2004), In situ observation of texture development in olivine, ringwoodite, magnesiowüstite and silicate perovskite at high pressure, *Earth Planet. Sci. Lett.*, *226*, 507–519.

Yuan, X., S. V. Sobolev, and R. Kind (2002), Moho topography in the central Andes and its geodynamic implications, *Earth Planet. Sci. Lett.*, *199*, 349–402.

C. N. Garzione, Department of Earth and Environmental Sciences, University of Rochester, Rochester, NY 14627, USA.

P. Molnar, Department of Geological Sciences and Cooperative Institute for Research in Environmental Science, University of Colorado, Boulder, CO 80309, USA. (peter.molnar@colorado.edu)